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The Holocene Palaeoecology of Lorraine Lake, Jasper National Park, Alberta.

by

Rhonda Bear

A THESIS

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IN

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ABSTRACT

Palaeoenvironmental and palaeolimnological reconstructions from three limnic sediment cores are presented for Lorraine Lake (52° 44' N, 117° 40' W), Jasper National Park, based upon sediment stratigraphy, sedimentary plant pigments, pollen analysis, radiocarbon dating and tephrostratigraphy.

Organic sedimentation in the lake began ca. 8, 800 RYBP. Pollen and macroremains indicate that an established vegetation existed prior to the inception of organic sedimentation. In the early Holocene, water levels in the shallow portion of the basin were low, and reached depths similar to present day about 6, 800 RYBP. The pollen spectra indicate that there has been little compositional change in the vegetation throughout the Holocene. Climatic conditions were warm and dry in the early Holocene, followed by a short period of cooler and wetter conditions (ca. 7, 500 to 7, 000 RYBP). Ca. 4, 000 RYBP, regional cool and moist conditions appeared. After 4, 000 RYBP it appears as if modern vegetational trends were established at the site.

About 8, 800 RYBP a blue - green algal flora dominated in Lorraine Lake. High concentrations of oscillaxanthin show that populations of *Oscillatoria* spp. peaked in the early Holocene. Throughout the Holocene, the carotenoid and total chlorophyll a pigment concentrations steadily increase and show that the lake has become more productive in the late Holocene. Pyrite stratigraphies, and pigment records indicate that the surficial oxygen

concentrations at each site vary and these variations must be taken into account when interpreting fossil pigment spectra.

Bridge River, Mount St. Helen's set Y, and Mazama tephras are present in each of the cores. Evidence from this study suggests that it is likely that only one member of the Mount St. Helen set Y tephra is present in the Jasper Park area. The erroneous radiocarbon dates illustrate the hazards of dating lake sediments in calcareous areas, and that whenever possible terrestrial macrofossils should be utilized for dating.

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Ag	argilla granosa - silt composed of mineral particles between 0.06 and 0.002 mm in size
a.s.l.	above sea level
AMS	accelerator mass spectrometry
^{12}C	carbon - 12
^{14}C	carbon 14 (radiocarbon)
calc	calcareousness
CaO	calcium oxide
Cl	chloride
CO_2	carbon dioxide
Dl	detritus lignosus - fragments of wood and bark, > 2 mm in size
elas	elasticitas - degree of elasticity
FeO	ferrous oxide
K_20	potassium oxide
kV	kilovolt
Lc	limnus calcareous (marl)
Ld	limnus detrituosus (lake mud)
nA	nanoangstrom
part. test. (moll)	particulae testarum molluscorum - broken mollusc shells
RYBP	radiocarbon years before present
sicc	siccitas - degree of dryness
strf	stratificatio - degree of stratification

TAP	total a pigments
TC	total carotenoids
test. (moll.)	testae molluscarum - whole mollusc shells
yrs B.P.	years before present

1. Introduction

1.1 General Introduction

Ecology may be defined as the study of the relationships between organisms and their environment (Birks and Birks 1980). Similarly, palaeoecology may be defined as the study of the past relationships between organisms and the environment in which they lived (Birks and Birks 1980). Palaeoecology is a historical science and involves both geological and biological working methods. Since it is impossible to directly observe or study the relationships between the past biota and their environment, palaeoecological research usually focuses upon the reconstruction of past aquatic and terrestrial ecosystems and climate using biotic evidence, i.e., microfossils and macrofossils, and abiotic evidence, i.e., the physical and chemical characteristics of sediments.

Palaeoecology has several inherent limitations not found in ecology. These are: i) direct observation of the flora and fauna is impossible; ii) the physical and chemical attributes of an ecosystem cannot be directly studied; iii) generally, palaeoecologists have little control over temporal and spatial boundaries and species selection, while ecologists operate within specified boundaries of time and space and select the species to be studied and, iv) a palaeoecologist is concerned with evolution, migration, the processes of diagenesis, transportation and redeposition while an ecologist usually is not (Birks and Birks 1980).

Establishing a time - frame is an essential part of palaeoecological studies. The most common radiometric method for dating Holocene sediments and fossils is carbon - 14 analysis. This depends upon the steady formation of ^{14}C by the bombardment of ^{14}N molecules by neutrons in the upper atmosphere. The ratio of ^{14}C:^{12}C is assumed to be the same as when the sample or sediment was laid down (Moss 1980). The ^{14}C:^{12}C ratio is measured in the material to be dated then compared with the initial ratio. There are three major problems associated with radiocarbon dating: 1) lake sediments and fossils less than 300 years old cannot be dated because of the standard deviation of ± 100 years, and the large amounts of ^{14}C deficient carbon that have been released into the atmosphere with the burning of fossil fuels. 2) Contamination of limnic or terrestrial sediments and fossils by carbon that is not in equilibrium with atmospheric CO_2; this is known as the reservoir effect (Olsson 1979). Radiocarbon dating fossils of terrestrial origin and peat is generally preferred over aquatic sediments and fossils because of the increased chance of the reservoir effect in sediments and lake basins. However, MacDonald *et al.* (1987) have shown that old carbon can be preferentially absorbed by some aquatic moss species and that radiocarbon dates from mosses should be critically examined. 3) Relatively large amounts of material are needed to yield enough carbon for radiocarbon dating. For example, 200 to 400 g of gyttja is often necessary but is not always sufficient for radiocarbon dating, and as much as 10 g of wood is necessary (Olsson 1979). With the advent of mass spectrometry radiocarbon dating (AMS), it is possible to count the ^{14}C atoms in a sample weighing less than 15 mg (Litherland 1985). This method utilizes very small samples and provides accurate dates.

In addition to radiocarbon dating, in northwestern North America there are widespread, well dated volcanic ash layers (tephras) which are useful as stratigraphic markers and for dating. Examples of several widespread ashes are presented in Table 1.

Prior to 1980, Holocene palaeoecological investigations in Alberta were limited to a few studies (Hanson 1949a, 1949b, 1952; Heusser 1956; Litchi-Federorvich 1970, 1972). Since then palaeoecological research in Alberta has become more prominent with numerous published studies from central Alberta (Hickman and Klarer 1981, Hickman *et al.* 1984; Schweger *et al.* 1981; Vance *et al.* 1983), northern Alberta (MacDonald 1987a, 1987b; White and Mathewes 1986; White et al. 1985), and the foothills and Rocky Mountain regions (Beaudoin 1984, 1986; Bombin 1981; Kearney and Luckman 1983a, 1983b 1987; Luckman 1988; Luckman and Kearney 1986; Luckman and Osborn 1988; MacDonald 1982, 1989; Reasoner and Hickman 1989). There have also been general reviews of the postglacial palaeoecology of Alberta by Ritchie (1976), Schweger and Hickman (1989), and Vance (1986).

Several widespread trends can be observed from sites in Alberta. Arid and warm conditions prevailed in the early Holocene, with cold winters and warm summers (Schweger and Hickman 1989). After 8, 000 radiocarbon years before present (RYBP), many shallow lake basins filled, mostly in the period from 6, 500 to 4, 500 RYBP, and none after 3, 000 RYBP (Schweger and Hickman 1989). Palaeohydrological records and diatom stratigraphies indicate that

Table 1. Common Holocene Tephras Found in Northwestern North America (Modified from A.M. Sarna - Wojcicki *et al.* (1983) in "Late Quaternary Environments of the United States." H.E. Wright Jr. Ed. Vol. 2)

Tephra	Source Area	Age	Area of Distribution	References
Glacier Peak Set G	Glacier Peak, Washington	12, 750 ± 350	Washington, Montana British Columbia, Alberta.	Porter 1978.
Glacier Peak Set M	Glacier Peak	12, 000 (?)	Washington	Porter 1978.
Glacier Peak Set B	Glacier Peak	11, 200 ± 100	Washington, Idaho Wyoming	Porter 1978.
Mount St. Helen's Set J	Mount St. Helen's, Washington	11, 700 ± 400- 8, 300 ± 300	Washington	Mullineaux *et al.* 1978.
Mazama Set O	Crater Lake, Oregon	6, 800	Oregon, Washington, British Columbia, Alberta, Saskatchewan	Bacon 1983.
Mount St. Helen's	Mount St. Helen's,			
Set Yb		3, 900 ± 250	(?)	Crandell & Mullineaux 1973, Mullineaux *et al.* 1975, 1978.
Set Yn		3, 510 ± 230	Washington, Oregon, British Columbia, Alberta	
Set Ye		3, 350 ± 250	Washington, Oregon	

Table 1. Continued

Tephra	Source Area	Age[a]	Area of Distribution	References
Mount St. Helen's Set P	Mount St. Helen's	2, 930 ± 250 - 2, 580 ± 250	Washington, British Columbia.	Mullineaux *et al.* 1975, Westgate 1977
Bridge River	Meager Mountain, British Columbia	2, 350	British Columbia, Alberta	Westgate 1977 Mathewes and Westgate 1980.
Mount St. Helen's	Mount St. Helen's			
Set Wn		<1, 150	Washington, Idaho	Crandell and Mullineaux 1978
Set We		> 450	Washington, Idaho	Mullineaux *et al.* 1975

water levels of some lake basins decreased during the mid - Holocene (Ritchie 1976; Schweger and Hickman 1989). Most vegetational and climatic change seems to have taken place in the period from ca. 8, 000 to 3, 000 RYBP with the establishment of modern vegetation and climatic trends ca. 3, 000 RYBP in central Alberta (Schweger and Hickman 1989), and after 5, 000 RYBP in northern Alberta (MacDonald 1987a).

Holocene palynological and palaeoecological research in the Rocky Mountains, Jasper Park area mainly has focused on alpine sites (Beaudoin 1984, 1986; Kearney and Luckman 1981, 1983a, 1983b, 1987; Luckman and Kearney 1986), except for a few subalpine studies (Heusser 1956; Kearney and Luckman 1987; Bear and Hickman 1989). Pollen and macrofossil evidence from sites above timberline suggest that ca 8, 800 RYBP timberline was at least 100 m above present levels (Luckman and Kearney 1986). Higher than present timberlines persisted until about 5, 200 RYBP, with a short interval (7, 300 to 7, 500 RYBP) when treeline receded to modern elevations. Treeline has been at or close to its present limits from ca. 4, 500 RYBP (Luckman and Kearney 1986). Data from a subalpine site near Maligne Lake indicates that ca 2 600 to 3 400 RYBP, a warm interval occurred as determined by ^{18}O and ^{13}C variations in mollusc shells (Kearney and Luckman 1987). Beaudoin (1986) and Osborn and Luckman (1988) also report evidence from elsewhere in the park suggesting a warmer climate at about this time.

Alpine palynological studies indicate that the elevation of timberline has fluctuated throughout the Holocene, but there has been little compositional change in the vegetation. The alpine pollen graphy from the Sunwapta Pass (Beaudoin 1984) is an exception to this and shows low percentages of arboreal pollen and a high percentage of *Artemisia* pollen in the basal zone. There are few subalpine pollen stratigraphies which predate 8, 000 RYBP, and most of these records also indicate that there has been little compositional change in the vegetation. The apparent insensitivity of the pollen records in mountainous regions is probably due to the over or under representation of many taxa, and the wide variety of microclimates that exist in the region. The climate of the mountain region is extremely variable over short distances and is determined chiefly by geographic location (latitude, longitude, elevation) and topography (Holland and Coen 1982).

In the Lake O'Hara region of Yoho National Park, relatively warm climatic conditions prevailed during the early Holocene (Reasoner and Hickman 1989). The pollen stratigraphy from upper subalpine Lake O'Hara also shows a basal zone dominated by *Artemisia*. (Reasoner and Hickman 1989). Timberline declined gradually in the period from ca. 7, 000 to 3, 000 RYBP, reaching modern levels after 3, 000 RYBP (ibid).

In the mountains of southwestern Alberta, the climate was arid and cool in the early Holocene but by 9, 400 RYBP moist conditions had set in and open *Picea* and *Pinus* forest was established (MacDonald 1989). Warm and arid conditions prevailed during the mid - Holocene, ca. 8, 000 to 5, 000 RYBP, when the *Pinus* and *Picea* forest was largely replaced by grassland and *Pinus flexilis* . After 5, 000 RYBP, cooler and moist conditions set in and modern vegetation (*Pinus contorta* and *Picea glauca* forest) was established (MacDonald 1989).

Sedimentary plant pigments provide information on lake history, and changes in primary production and trophic status (Bengtsson 1979, Guilizzoni et al. 1982. Moss 1980, Sanger and Gorham 1972, Valletyne 1955, Wetzel 1983). The pigment content in sediments is usually correlated with levels of algal biomass and production at the time the sediment was deposited (Guilizzoni *et al.* 1982). Stratigraphic shifts in the concentration of fossil pigments can be related to changes in the physical/chemical nature of the lake and drainage basins Upon senescence and death of plant material, the photosynthetic pigments undergo degradation, the products of which are relatively stable, especially if in situ conditions are anoxic and light is limited (Wetzel 1983). Undegraded chlorophyll, chlorophyllides, pheophytin, pheophorbides, and carotenoids are the most widely used pigments for analysis (Wetzel 1983). Specific pigments synthesized by particular algal groups can be measured to infer invasion time and development of these algae. Analysis of the blue - green algal carotenoids oscillaxanthin and myxoxanthophyll have been widely documented to infer changes in trophic status (Brown and Coleman 1963, Moss 1980, Swain 1985, Züllig 1981).

This study was undertaken to further document Holocene vegetation development, climatic change and lake productivity in the Maligne Valley in Jasper National Park. Limnic sediment cores from a subalpine Lorraine Lake, Jasper National Park were subsampled for water content, organic matter content, pollen, diatoms, pigments, macrofossils, radiocarbon dating and tephras. This thesis will focus primarily on sediment stratigraphy, ^{14}C dating, fossil pigment records, tephrostratigraphy and pollen analysis to provide further insight on: i) chronology of deglaciation of the area; ii) subalpine Holocene vegetational development; iii) differences among pollen stratigraphies from shallow and deep sites in one lake; iv) apparent climatic trends which may be compared to trends already documented for the Jasper area, and possibly for the rest of Alberta; v) changes in lake productivity and, vi) the identification and distribution of Holocene tephras.

1.2 Site Description

Lorraine Lake (1, 750 m above sea level) is situated in the Maligne Valley (52° 44' N, 117° 40' W) in the Front Ranges of Jasper National Park, Alberta (Figure 1). The Maligne Valley divides the Front and Main Ranges, with quartzite and limestone rock on its western and eastern sides, respectively (Kuchar 1972). Prominent glacial features (hummocky moraine, outwash plain, kettle holes, kettle lakes, moulin kames) left by a valley glacier which flowed down the Maligne Valley from the Brazeau Icefield are present at the north end of Maligne Lake (Roed 1964). Lorraine Lake is a kettle basin likely formed by this valley glacier, and is located 2.4 km from the northwest shore of Maligne Lake.

Long term climatic data are only available from the Jasper meteorological station (1, 062 m above sea level). The mean summer temperature at Jasper is 13.9 °C and total precipitation is 561 mm/year (1951 - 1980 average, Alberta Environment 1986). The mean summer temperature at Lorraine Lake can be approximated

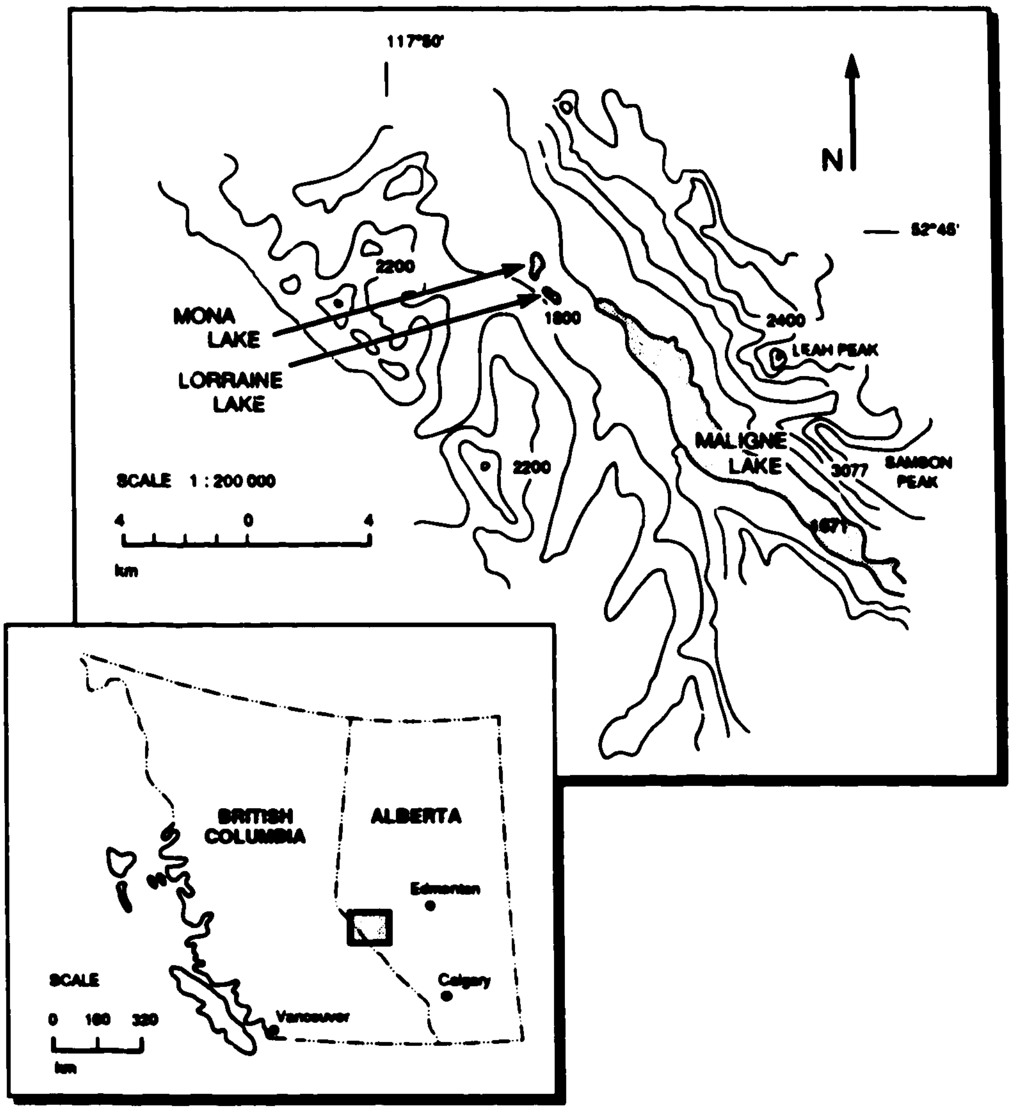


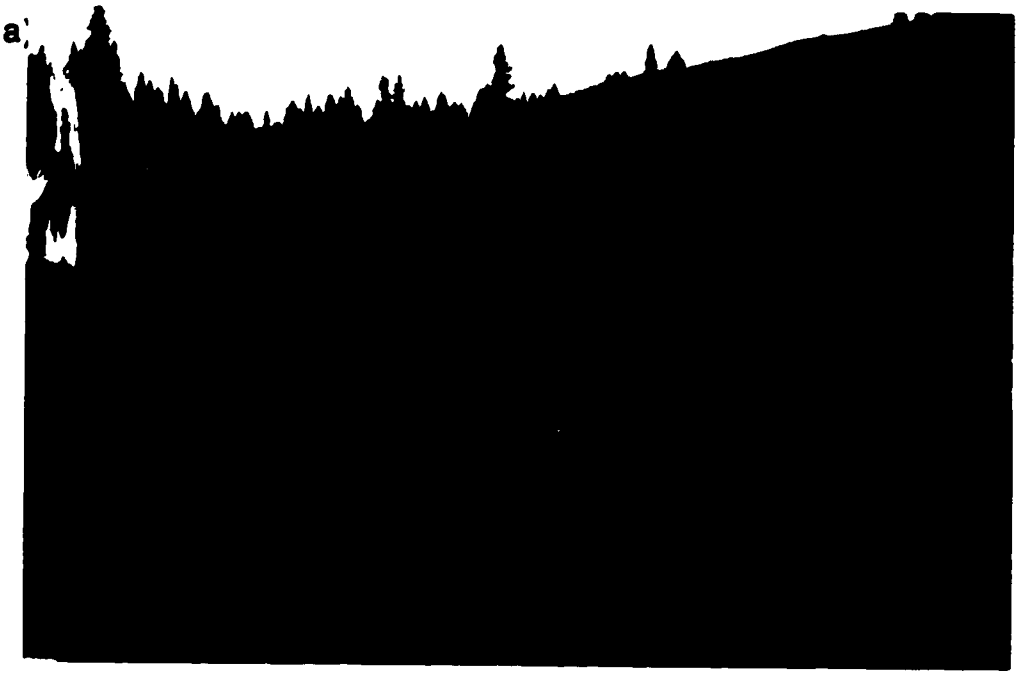
Figure 1. Map showing the location of Lorraine Lake.

by using the summer lapse rate of 0.5 °C/100 m (Holland and Coen 1982). The summer temperature at this site would be approximately 10.5 °C. Average annual precipitation recorded at Maligne Lake over a ten year period to 1980 is 565 mm/year (Holland and Coen 1982).

The subalpine ecoregion occurs at elevations above the montane ecoregion, and to approximately 2, 300 m a.s.l. in Jasper National Park (Holland and Coen 1982). Winter precipitation is greater than summer precipitation, with precipitation in the eastern regions generally less than that of the western regions (ibid). Winds are usually light in the subalpine but there is much variation. Lorraine Lake is situated in the lower subalpine and lies on the boundary between two ecosites; morainal with non - calcareous medium textured till, and ice - contact stratified drift with calcareous parent material (Holland and Coen 1982). The surrounding arboreal vegetation consists mainly of *Pinus contorta* Loudon, with some *Picea engelmanni* Parry *ex* Engelm. and *Abies lasiocarpa* (Hook) Nutt.. The high percentage of pine in the area is a result of fire which swept through the area in the early part of this century (Kuchar 1972). Shrub and dwarf vegetation surrounding the lake comprises *Ledum groenlandicum* Oeder, *Rosa acicularis* Lindl., *Betula glandulosa* Michx., *Salix* spp., *Juniperus communis* L., and *Empetrum nigrum* L.. *Carex rostrata* Stokes is the dominant emergent hydrophyte in the littoral zone. Plates 1a and 1b are of the east and west shores (respectively) of Lorraine Lake and show the extent of the growth of *Carex rostrata.*

Lorraine Lake is a small lake with a maximum depth of 8.5 m, and covers an area of 0.09 km^2. A small stream flows northward from the lake into the Maligne River, and several small streams flow into the lake but their origins are unknown. A gelatinous algal mat (up to 10 cm in depth) forms on the sediments on the east side of the basin in the early spring and persists until fall. The mat is primarily composed of the blue green alga *Apanothece gelatinosa*, diatoms and bacteria. In the deepest section of the lake basin the sediments are anoxic throughout most of the year.

1a)



1b)

Plate 1. View from Lorraine Lake facing: 1a) east, and 1b) west.

2. Materials and Methods

2.1 Coring Procedure and Initial Core Description

Five sediment cores were extracted from Lorraine Lake across an east - west transect with a percussion corer (Reasoner 1986). Locations of the coring sites, and water depths are presented in Plate 2. Plates 3 through 7 show the coring equipment and procedure. The percussion coring system is easily transportable and several sediment cores can be retrieved in one day. The sediments were collected in 7 cm diameter, 3 m lengths of plastic piping, capped and transported (without freezing) to the University of Alberta, where they were stored upright at 10 °C.

The core barrels were split lengthwise with a circular saw. Piano wire was pulled through the cut core barrel to separate the sediment before pulling the two halves of the core barrel apart. After initial examination and measurement, the sediments were described with respect to composition, texture, Munsell Colour Scale (Oyama and Takehara 1970), tephras and macrofossils. Three sediment cores LOR3, LOR4, and LOR5 were judged to be complete sections and selected for subsampling and analyses. The subsamples were removed from the center of the core barrel to ensure that contamination due to deformation of the sediments adjacent to the core barrel did not occur.

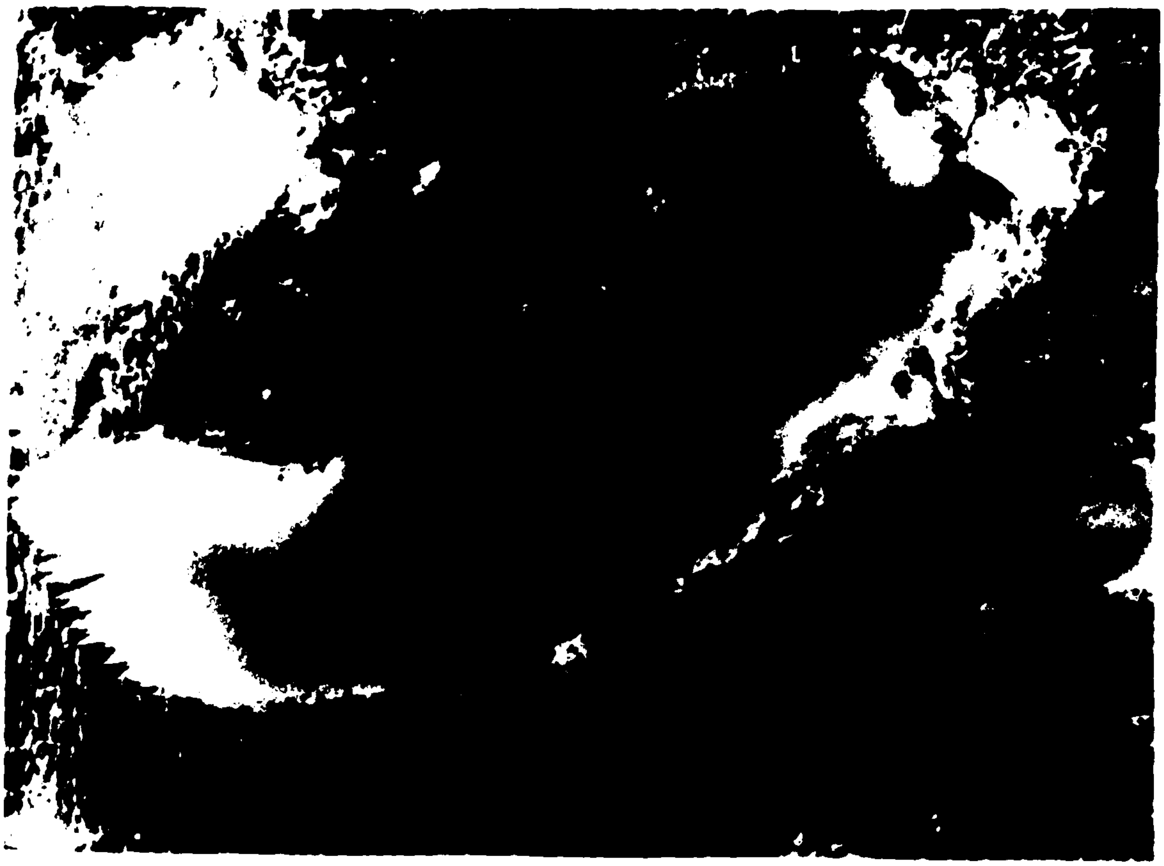


Plate 2. Aerial photograph of Lorraine Lake with coring sites marked by ✪ : from left to right, water depth = 6 m (LOR1), water depth = 8.5 m (LOR2 & LOR3), water depth = 4 m (LOR5), and water depth = 5 m (LOR3).

Plate 3. Close up of driver head.

Plate 4. Close up of the core barrel attached to the core head.

Plate 5. Lowering the core barrel with the core head attached into the water.

Plate 6. Standing over the coring site, raising and lowering the driver head.

Plate 7. Pulley system fastened to the ice with ice screws used to extract the core barrel from the sediments.

2.2 Analytical Methods

Percent water, organic matter, and carbonate content for cores LOR3 and LOR5 were calculated using 1 cm^3 subsamples of wet sediment taken at 5 cm intervals. Water content was determined by placing the pre - weighed fresh sediment subsamples in an oven at 105 °C for 24 hours. Organic matter content was determined after ignition of the samples at 550 °C for 1 hour, and carbonate content was determined by further ignition at 950°C for 3 hours (Wetzel 1970).

One cm^3 subsamples from cores LOR3 and LOR5 were processed for pollen analysis according to the procedure of Faegri and Iverson (1975). Some basal samples from LOR3 required additional treatment with heavy liquid, $ZnBr_2$, to remove silts and clays (Schweger 1976). To estimate pollen concentration two exotic pollen tablets were added to each sample (Stockmarr 1971). The pollen was mounted in silicon oil (refractive index 1.4). Pollen sums generally exceeded 500 grains per slide, and did not include aquatics or spores. The palynomorphs were identified and enumerated with a Leitz Laborlux microscope at 550X magnification. The pollen reference collection of the Palaeoenvironmental Laboratory at the University of Alberta, Anthropology Department was used to aid in identification along with the "Key to principal pollen types of Alberta" (Habgood 1978). Confidence limits were calculated according to Mosimann (1965). The numerical zonation programs SPLITLSQ, SPLITINF, CONLSINK-MAX AND CONSLINK-MIN were employed (Birks 1979, Birks and Gordon 1985) on the pollen records.

Gyttja samples from LOR3 were wrapped in foil and sent to the Alberta Environmental Center in Vegreville, Alberta for conventional radiocarbon assays. The samples were pretreated with NaOH and HCl to remove any mobile humic or carbonate contaminants. Charred macroremains from a basal clay and silt layer in LOR4 were also sent to the Alberta Environmental Center laboratory for radiocarbon dating. A *Pinus* ovuliferous scale from the layer of charred macroremains was sent to Beta Laboratories in Florida for an accelerator mass spectrometer radiocarbon date.

One cm^3 subsamples for pigment analysis were removed at 5 cm intervals for chlorophyll derivatives (CD), and bulk carotenoids (TC) in cores LOR3 and LOR5. CD and TC pigments were extracted in 10 ml of 90% methanol, followed by centrifugation. Absorbancy readings were taken at 750 nm (turbidity reading) , 665 nm (CD) and 480 nm (TC). Analyses for both total a pigments, and bulk carotenoids followed that of Marker (1972). Units are calculated and expressed as absorbance per gram organic matter where one unit is equal to an absorbance of 1.0 in a 10 cm cell when dissolved in 100 ml solvent (Sanger and Gorham 1972). Per cent native chlorophyll was calculated with Lorenzen's (1967) equation:

$$\text{percent native chlorophyll} = \frac{665B - 665A}{0.7\,(665A)} \times 100 \qquad (1)$$

where 665B is absorbance at 665 nm before acidification and 665A is absorbance at 665 nm after acidification with 0.0035 N HCl.

Ten gram fresh weight subsamples were removed at 10 cm intervals for oscillaxanthin and myxoxanthophyll pigment analyses.

These analyses followed the trichromatic method of Swain (1985). The pigment extract in 5 ml of ethanol was read at the absorbances of 412, 504, and 529 nm. Net oscillaxanthin and myxoxanthophyll can be calculated using the following equations:

$$O_{529} = (1.266)\ A_{529} - (0.219)\ A_{504} - (0.081)\ A_{412} \qquad (2)$$

$$M_{504} = (1.358)\ A_{504} - (1.308)\ A_{529} - (0.031)\ A_{412} \qquad (3)$$

Absorbance at each major peak is calculated with the equations:

$$O_{495} = (1.27) O_{529} \qquad (4)$$

$$M_{473} = (1.20) M_{504} \qquad (5)$$

The concentration of the pigments are calculated in ug pigment g^{-1} organic matter using the equation:

$$= \frac{10\ 000\ (V)\ (A)}{(E)\ (P)\ (g\ org.)} \times 100 \qquad (6)$$

where (A) is the appropriate major peak absorbance from equation (4) or (5), (g org) is the weight of organic matter in the extracted sediment, (P) is the proportion of the pigment extract, (V) is the volume of ethanol extract in ml, and (E) is the appropriate extinction coefficient (1450 for oscillaxanthin and 2100 for myxoxanthophyll).

2.2.1 Electron Microprobe Analysis

Tephras encountered in the cores were initially identified on the basis of their stratigraphic position, shard morphology and petrographic characteristics. The morphology of the shards is dependent upon the properties of the magma and the eruptive conditions (Westgate and Gorton 1981). Electron microprobe analysis, a grain discreet chemical analysis, was also employed to assist in identifying the tephras. Late Quaternary tephra deposits in southwestern Canada can be distinguished on the basis of their Fe, Ca, and K content (Westgate and Gorton 1981).

About 10 g of each sample was taken, disaggregated, and sieved through a size 70 mesh. The non - magnetic fraction was treated with heavy liquids and the heavy fraction dried. The heavy fraction was treated once more with heavy liquids, then dried and sieved. A small amount of each of the purified fractions was mounted in epoxy resin and polished thin sections were cut.

Tephra samples from core LOR5 were analyzed by Dr. Jim Beget at Washington State University, Pullman, Washington following Smith and Westgate (1969). At least 16 shards were analyzed for each sample and more were analyzed for those samples which showed heterogeneity. All determinations were made with a Cameca MBX microprobe at 15 kV, with an 11.87 nA probe current, and 8 micron beamspot. Count time was 10 seconds for each grain. Total Fe was calculated as Fe_2O_3. All other elements are presented as oxides except for Cl (elemental). The obsidian working standard CCNM211 (Smith and Westgate 1969) was included in the analytical run in order to calculate the element concentrations in the sample.

3. Results

3.1 Sediment Stratigraphy

3.1.1 Sediment Description

Sediment structure, texture, and type is classified according to the Troels - Smith sediment characterization system (Aaby 1979) and the Munsell soil colour code (Oyama and Takehara 1970). The sediment stratigraphy of core LOR3 is presented in Figure 2. Core LOR3 is 268 cm in length including a 6 cm dried sediment plug at the base. The sediment comprises primarily gyttja ($Ld^{O}4$). There are no distinct changes in texture or structure throughout the core (strf 2, elas 3, sicc 3), except for two marl layers (Lc3 Ag1 [part. test.(moll.) 2]) at 257.5 cm and 260 cm, 1.5 cm and 1 cm in width respectively. The sediment colour is described in Table 2. The sediment had a strong sulfurous odor throughout, and changed to an orange - brown colour after exposure to air. Four distinct tephras are present (Bear *et al.* in preparation): i) Mazama, 6, 800 years B.P.(Bacon 1983) is located from 221.5 to 226.5 cm; ii) two members from the Mount St. Helen's Set Y group, 3, 400 years B.P. (Luckman *et al.* 1984) at 123 to 127 cm and 118 to 118.5 cm, and iii) from 85.5 to 86.5 cm is Bridge River, 2, 350 years B.P. (Mathewes and Westgate 1980).

The stratigraphy of core LOR5 is presented in Figure 3. The core is considered to represent a complete record of the Holocene as clay was present at the base of the core barrel when it was extracted from

Table 2. Munsell sediment colours for core LOR3.

Sediment depth (cm)	Sediment colour
0 - 50	Black N 1.5/0 and olive black 5GY 2/1
50 - 76	Dark olive gray 5GY 3/1
76 - 86	2.5GY 3/1
86 - 115	7.5GY 2/1 Greenish black
115 - 122.5	10GY 3/1 Dark greenish gray
1[illegible] 5 - 175	5GY 3/1
175 - 179	N1.5/0 and 2.5GY 3/1 layers
179 - 220.5	5GY 3/1 and 10Y 3/2 layers
220.5 - 230	7.5Y 3/2
230 - 239	2.5GY 3/1
239 - 246	5GY 3/1
246 - 257.5	2.5GY 3/1 and 5GY 3/1
257.5 - 259	marl layer
259 - 260	5GY 3/1
260 - 261	marl layer
261 - 262	5GY 3/1

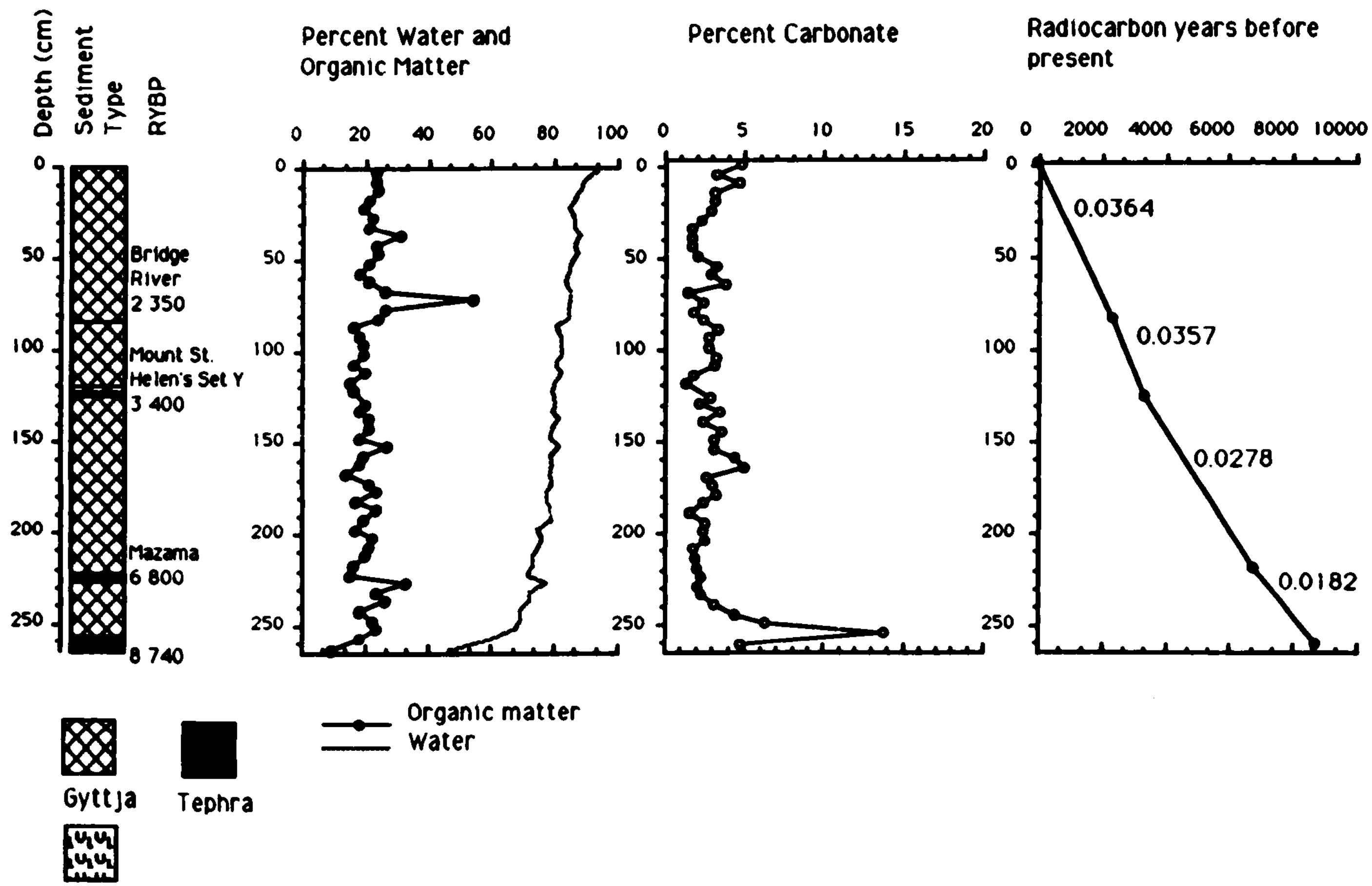


Figure 2 Percent water, organic matter, carbonate content and estimated sedimentation rates (using the tephra dates and the basal date of 8, 740 RYBP) for core LOR3.

the lake. LOR5 is 261.5 cm in length with a marl - clay layer (Lc3 Ag1 [test. (moll.) 2] from the base to 246 cm (Figure 3). From 246 to 227 cm, the composition of the sediment changes slightly to Lc2 Ld1 Ag1[test. (moll.) 1]. Plant macrofossils are abundant from the base of the core to 227 cm. The top 227 cm of sediment is gyttja (Ld^{o}4), with little change in sediment structure and texture throughout (strf4, elas3, sicc3). The colour of the sediment is presented in Table 3. The nature of the contact between the calcareous layer and the gyttja is diffuse. The tephras are found at: 204 to 207.5 cm (Mazama), 102 to 103 cm and 96 cm (Mount St. Helen's Set Y), and 66.3 to 66.5 (Bridge River).

Core LOR4 , 270 cm in length is divided into four stratigraphic units (Figure 4). The basal unit is sand (Ga4) occuring up to 242 cm. A small tephra deposit is embedded in the sand at 260 cm. A 10 cm layer of gray inorganic silt and clay with charred macroremains overlays the sand layer (As1 Ag1 Dl2; strf 0, elas 0, sicc 3, calc 0). The lower boundary between the clay and sand is very sharp. A marly unit (Lc2 Ag2; strf 4, elas 1, sicc 3, calc 4), with mollusc shells, leaves and wood fragments overtops the clay to 217 cm. The upper boundary is sharp. The top 217 cm is gyttja (Ld^{o}4; elas 3, sicc 3, strf 4, calc 0), with no colour change on exposure to air. Four tephras are present in the gyttja: 195 to 200 cm (Mazama), 123 to 125.5 cm and 118.5 to 118.8 cm (Mount St. Helen's Set Y), and Bridge river at 85.5 cm, 2.5 mm in width. Table 4 presents the sediment colours and Figure 5 presents the sediment stratigraphy.

Table 3. Munsell sediment colours for core LOR5.

Sediment depth (cm)	Sediment colour
0 - 47	10Y 4/2
47 - 55	7.5Y 4/2, olive colour
55 - 66.	10Y 4/2, grey
66 - 101	7.5Y 4/2, olive colour
103 - 160	10Y 4/2, olive grey and black layers, with 2,4 mm layers, 2.5Y 4/2 at 157 and 158 cm
160 - 162.5	5Y 4/2
162.5 - 165	10Y 4/2 and 5Y 2/2, grey and black layers
165 - 176	5Y 4/2 and 5Y 2/2, brown with blackish layers
176 - 180	8 laminations, 5Y 3/2 and 2.5 Y 4/2
180 - 186	10Y 3/2 and 5Y 2/2
186 - 190	2.5Y 4/2 and 5Y 2/2
190 - 197	10Y 3/2 and 5Y 2/2
197 - 204	5Y 4/2, 5Y 3/2
207 - 224.5	5Y 3/2
224.5 - 227	10Y 5/1
227 - 234	7.5Y 8/2 and 10Y 4/2
234 - 237	7.5Y 8/2 and 7.5Y 5/2

Table 3. cont'd

237 - 242	7.5Y 4/2 and 7.5Y 3/2
242 - 250	5Y 4/2
250 - 259	7.5Y 5/2 and 7.5Y 4/2
259 - 261.5	5Y 5/3

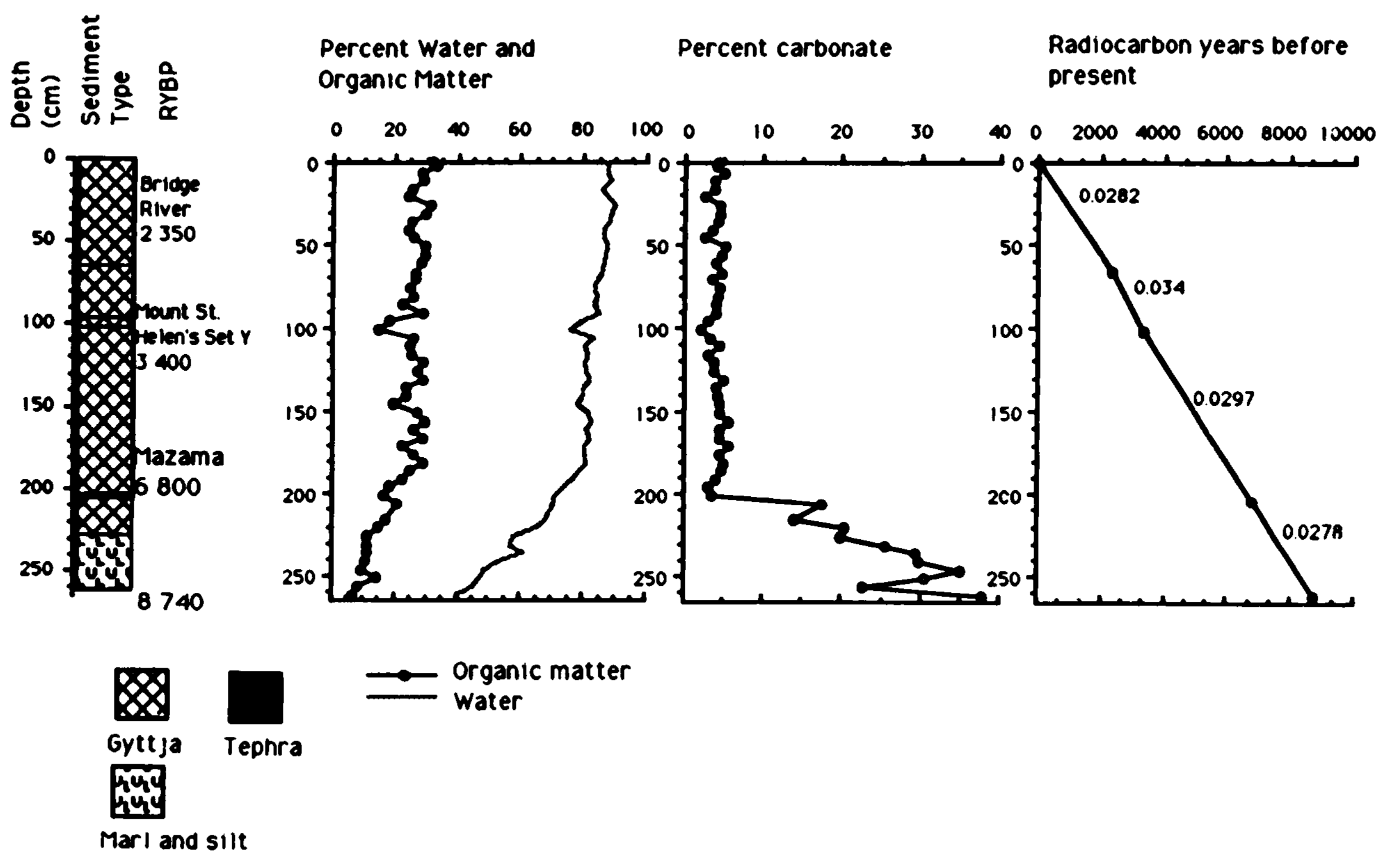


Figure 3. Percent water, organic matter, carbonate content and estimated sedimentation rates (using the tephra dates and the basal date of 8, 740 RYBP) for core LOR5.

Table 4. Munsell sediment colour for core LOR4.

Sediment depth (cm)	Sediment colour
0 - 24	greyish olive, 7.5Y 2/0
24 - 61	7.5Y 4/2
61 - 64	7.5Y 5/2
64 - 132	7.5Y 5/1
132 - 195.5	10Y 4/1 and 7.5 Y 2/1
200 - 201.5	7.5Y 4/1
201.5 - 210	7.5Y 5/2
210 - 215	7.5Y 6/2
215 - 217	7.5Y 5/1
217 - 232	7.5Y 5/2, 7.5Y 6/2
232 - 242	N 6/0
242 - 270	sand and clay

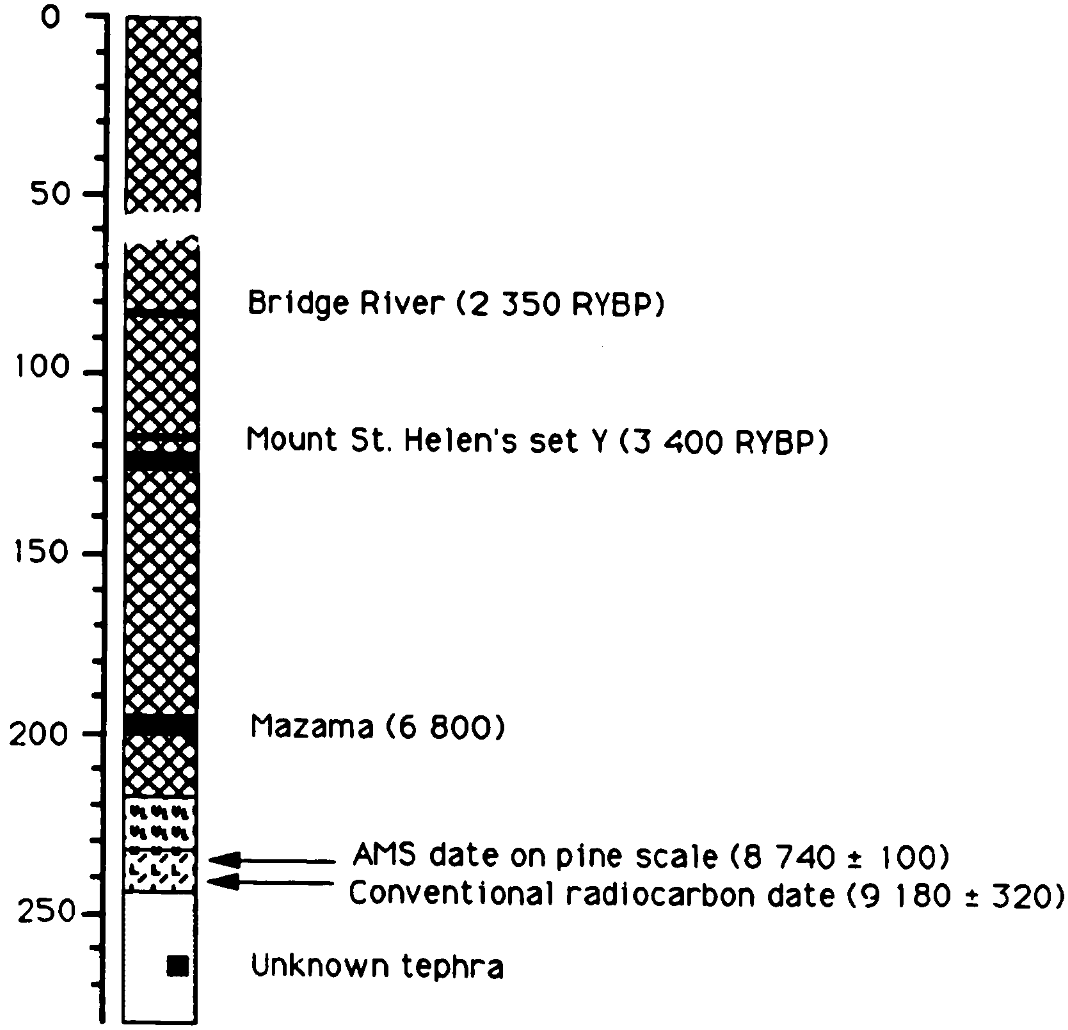


Figure 4. Sediment stratigraphy of core LOR4.

Cores LOR1 and LOR2 are incomplete and were not considered for analysis.

3.1.2. Percent Water, Organic Matter, Carbonate, and Sedimentation Rates.

3.1.2.1 Core LOR3

The sediment water content generally decreases with increasing core depth (Figure 2). There are no apparent trends in the organic matter content. The values fluctuate between 19% and 25% throughout the core, except for an isolated peak at 70 cm (55%). Percent carbonate content is greatest at the base (14%), then fluctuates between 2% and 5% throughout the rest of the core. Estimated sedimentation rates (cm/yr) were generated using the AMS date of 8 740 ± 100 RYBP, and tephra dates (Figure 2). Sedimentation rates increase as sediment depth decreases, with the lowest rates pre - Mazama.

3.1.2.2 Core LOR5

The percent water content, organic matter, and carbonate content of core LOR5 is presented in Figure 3. Sediment water content decreases with increasing sediment depth. Percent organic matter increases steadily from the base (7%) to 180 cm (28%). From 180 cm to the top, the organic matter fluctuates between 14% and 30%, but there are no apparent trends. Carbonate content reaches maximum percentages from 261.5 to 202 cm. It drops sharply thereafter, and fluctuates between 2% and 5% throughout the rest of the core. Estimated sedimentation rates (Figure 3) are calculated as for LOR3. The rates are generally lower that those of LOR3 and they remain relatively unchanged throughout the core.

3.1.3 Radiocarbon dating

The radiocarbon dates, depths and dated materials from cores LOR3 and LOR4 are presented in Table 5. The radiocarbon dates from gyttja which bracket the Mount St. Helen's Set Y tephras are too old, and it is likely that the basal date from the deep water core is also incorrect. Contamination from groundwater carbon is likely a problem at this site. The radiocarbon dates from the macroremains and the pine scale in core LOR4 overlap and suggest that ca. 9, 000 RYBP the basin started to fill and that an established vegetation existed at this time.

3.2 Pigment Analysis

In general, the total a pigment (TAP) stratigraphies from cores LOR3 and LOR5 show the same trends (Figures 5 and 6). Low concentrations are present at the base of both cores, then increase to ca. 6, 800 RYBP. Following the deposition of the Mazama tephra concentrations in both cores remain fairly stable with a small increase in LOR3 ca. 5, 000 to 4, 000 RYBP and in LOR5 ca. 5, 000 to 4, 500 RYBP. At about 3, 400 RYBP values begin to increase to the top of the core. The carotenoid stratigraphies (Figures 5 and 6) also show that the pigment concentrations have increased in the latter part of the Holocene. Percent undegraded chlorophyll stratigraphies

Table 5.

Radiocarbon dates for Lorraine Lake.

Core	Interval (cm)	Material	Laboratory Number	Radiocarbon age (years B.P.)
LOR3	115 - 118	gyttja	AECV-429C	6, 310 ± 190
LOR3	127 - 131	gyttja	AECV-430C	6, 980 ± 150
LOR3	262 - 268	gyttja	AECV-431C	12, 350 ± 440
LOR4	232 - 242	wood	AECV-501C	9, 180 ± 320
LOR4	232 - 242	pine scale	BETA-25916 ETH-4134	8, 740 ± 100

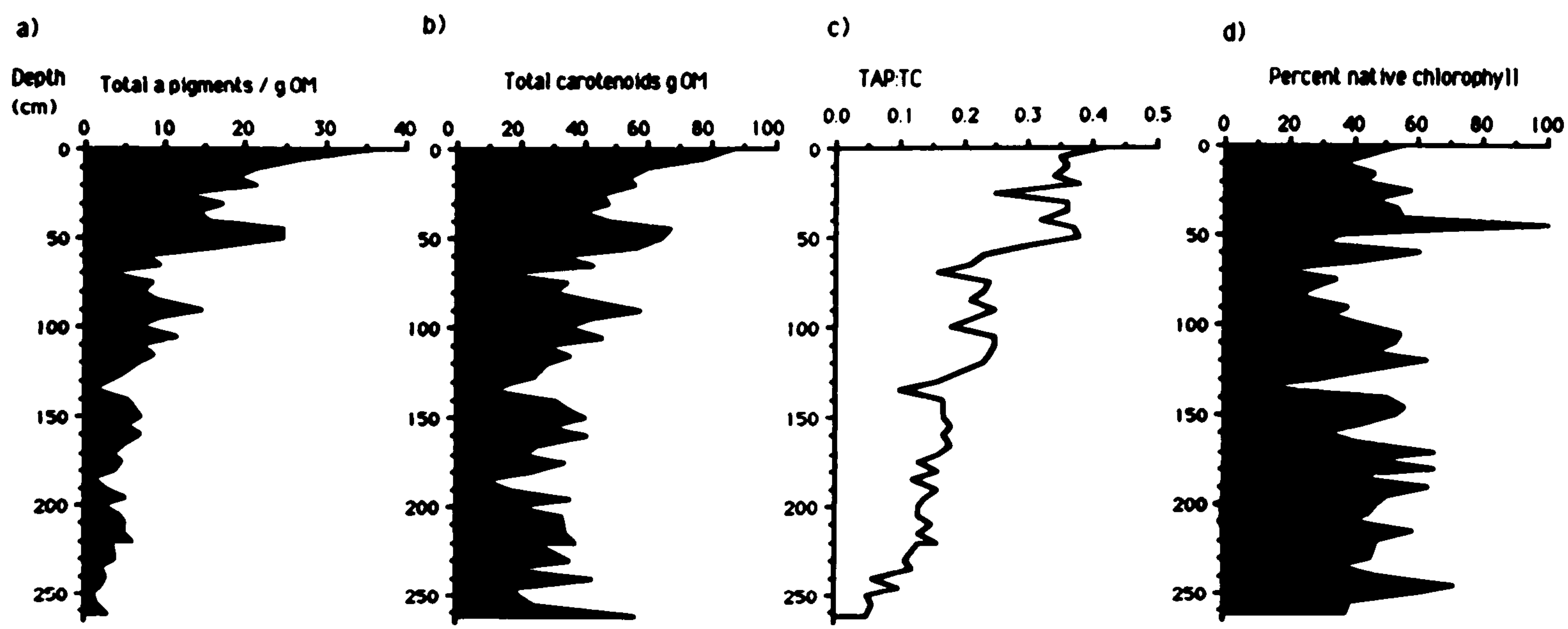


Figure 5. a) Total a pigments (SPU per g organic matter), b) carotenoids (SPU per g organic matter), c) TAP:TC ratio for core LOR3, d) percent undegraded chlorophyll.

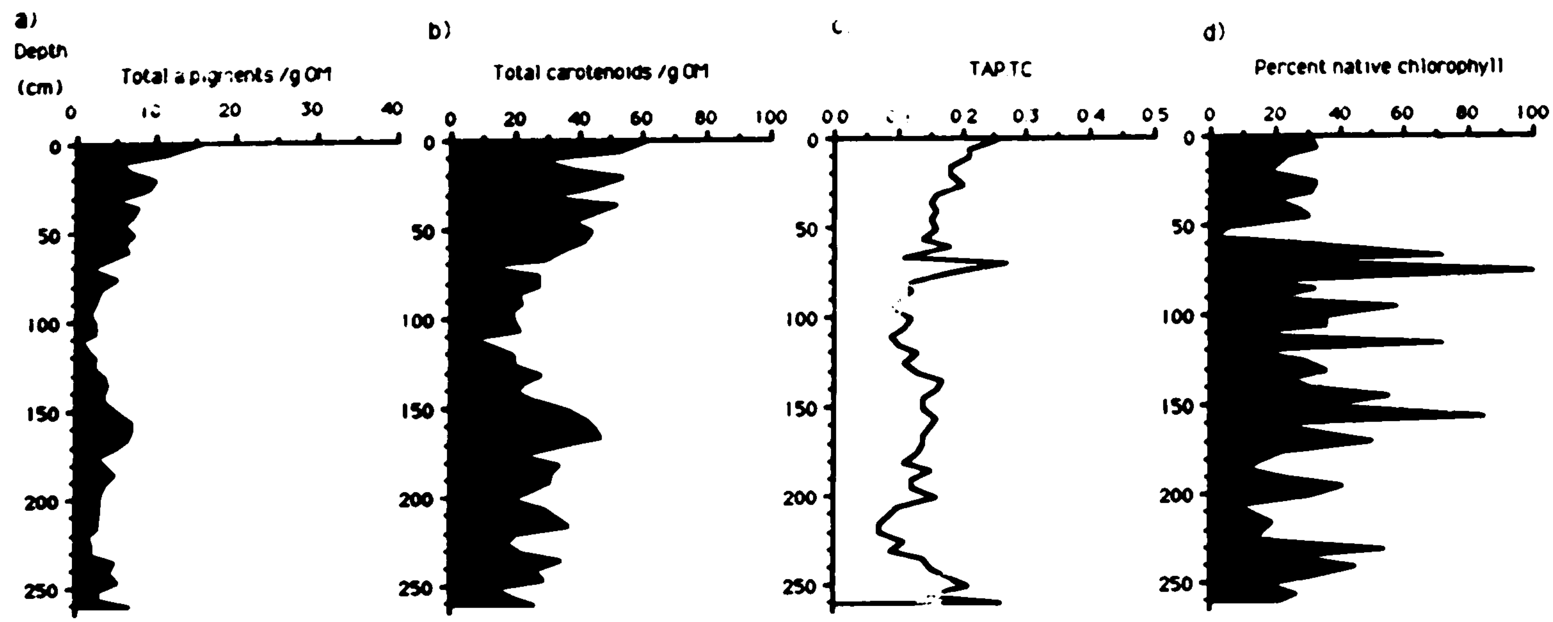


Figure 6. a) Total a pigments (SPU per g organic matter), b) carotenoids (SPU per g organic matter), c) TAP:TC ratio for core LOR5, and d) percent undegraded chlorphyll.

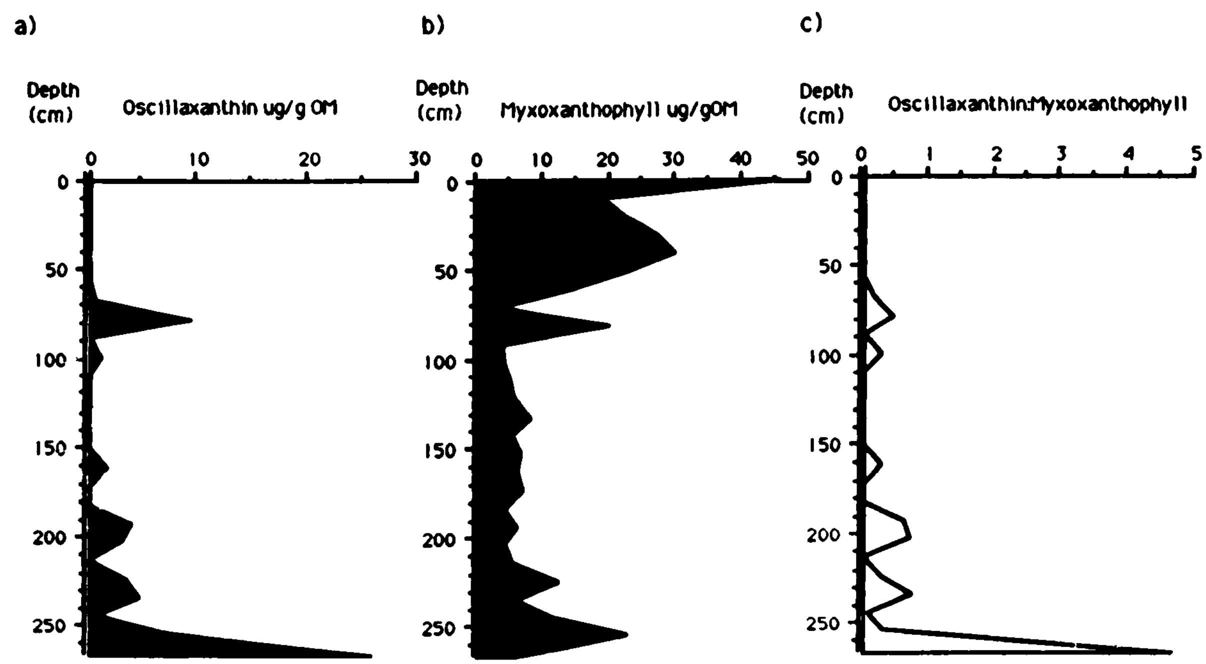


Figure 7. a) Oscillaxanthin content (ug/g OM), b) Myxoxanthophyll content (ug/ g OM), and c) Oscillaxanthin:Myxoxaanthophyll ratio for core LOR3.

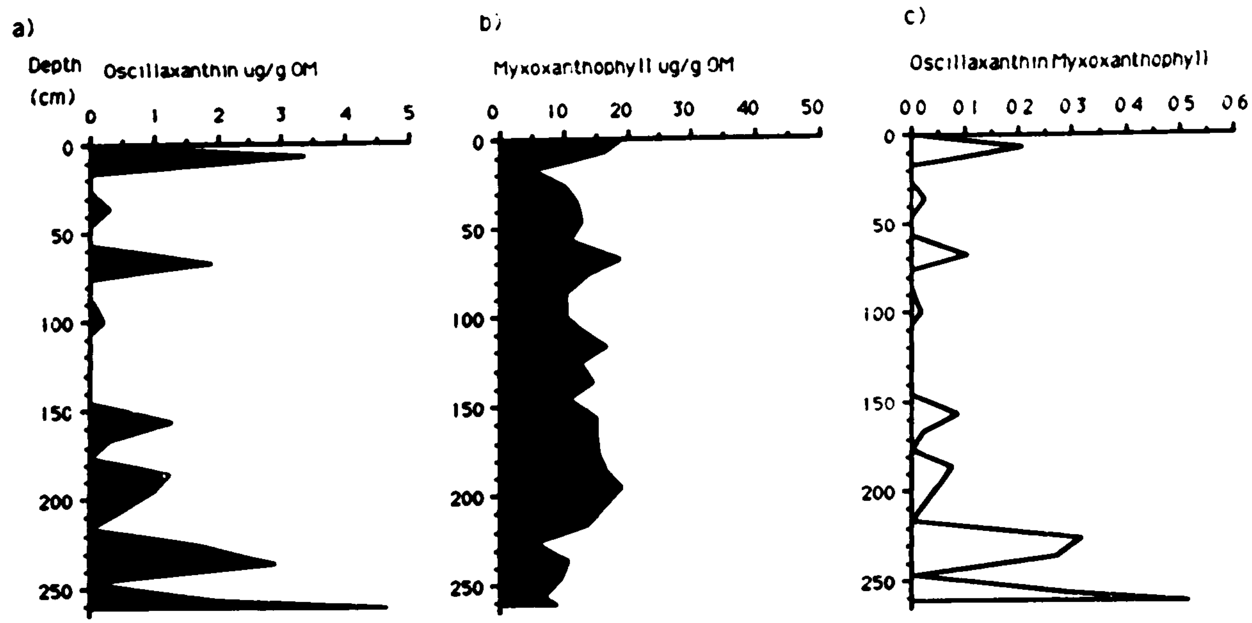


Figure 8. Oscillaxanthin content (ug/g OM). b) Myxoxanthophyll content (ug/g OM), and c) Oscillaxanthin:Myxoxanthophyll ratio for core LOR5.

show that more native chlorophyll is present in core LOR3, which may account for the greater pigment concentrations in this core.

The oscillaxanthin stratigraphies closely follow one another (Figures 7 and 8). Pigment concentration values peak at the base of each core, decrease then three successive peaks are observed. The upper pigment peaks in LOR5 are not present in LOR3. The myxoxanthophyll stratigraphies (Figures 7 and 8) do not follow one another closely. Myxoxanthophyll values peak at the base of LOR3 and then decrease ca. 6, 800 RYBP, and remain low during the mid - Holocene. Values begin to increase following the deposition of the Bridge River tephra (2, 350 RYBP). In core LOR5 myxoxanthophyll concentrations are lowest at the base of the core, and begin to increase ca. 7, 500 RYBP. Following the increase, values remain relatively stable and there are no further observable trends. Oscillaxanthin and myxoxanthophyll concentration values are greatest in LOR3.

3.3 Tephrostratigraphy

The depths and thicknesses of the tephras varied slightly in the five cores. The samples for electron microprobe analysis were taken from core LOR5. The Bridge River tephra was orange in colour, 2 mm in width at a sediment depth of 66.5 cm. The two St. Helen's tephras in the mid section of the core were 6 cm apart, the upper one was diffuse, and only 0.5 mm in width, while the lower one was 1 cm in width. The Mazama tephra was at 204 cm, and was 3.5 cm in width. The tephras were tentatively identified on the basis of their stratigraphic positions and by their characteristics in the freshly split core.

The results from electron microprobe analysis are presented in Table 6, and graphically in Figures 9 through 12. To identify the tephras, the data for CaO, FeO and K_2O were normalized to 100%, and the analyses were plotted against ranges previously reported for the tephras (Westgate and Gorton 1981). The glass chemistry data indicates that sample ATC - 0629 is Mazama, ATC - 0630 is St. Helen's Set Y, and ATC - 0632 is Bridge River (see Figures 9 through 11). No definitive correlation can be made for sample ATC - 0631 (Figure 12). This sample had multiple glass populations, some of which appeared to resemble the Mazama tephra, some of which resembled the St. Helen's Set Y and some of which appear to represent the Bridge River tephra. Petrographic examination of the sample showed that the glass shards were highly oxidized (Beget, personal communication). The shard morphology of Mazama and Bridge River resembles that reported by other workers (Westgate and Gorton 1981, Reasoner and Healy 1986) (see Plates 8 through 11). The shard morphology of St. Helen's Y ash has not been well documented, and is presented in Plates 12 and 13.

The samples' mineral content was examined as well. Cummingtonite, an unusual amphibole rarely described in tephras was found in ATC - 0630. Cummingtonite is a well known phase in St. Helen's tephras including Set Y. Biotite was present in ATC - 0632. Biotite is a mica rarely found in tephras, but has been previously described as an accessory phase in Bridge River (Westgate 1977, Westgate and Gorton 1981).

Table 6. Average glass composition of tephras in core LOR5, as determined by the electron microprobe technique[a] .

	Bridge River	Mount St. Helen's set Y		Mazama
	ATC632	ATC631	ATC630	ATC629
SiO_2	75.45 ± 0.59	75.36 ± 0.38	75.43 ± 0.48	3.19 ± 0.20
TiO_2	0.31 ± 0.06	0.23 ± 0.08	0.14 ± 0.04	0.41 ± 0.04
Al_2O_3	13.49 ± 0.36	14.01 ± 0.52	14.30 ± 0.25	14.43 ± 0.11
FeO[b]	1.45 ± 0.09	1.38 ± 0.17	1.42 ± 0.26	2.06 ± 0.09
MgO	0.28 ± 0.05	0.31 ± 0.08	0.37 ± 0.05	0.45 ± 0.03
CaO	1.21 ± 0.16	1.53 ± 0.39	1.84 ± 0.17	1.62 ± 0.07
Na_2O	4.41 ± 0.15	4.40 ± 0.13	4.31 ± 0.10	4.97 ± 0.17
K_2O	3.30 ± 0.09	2.69 ± 0.61	2.09 ± 0.12	2.69 ± 0.08
Cl	0.11 ± 0.04	0.10 ± 0.03	0.11 ± 0.09	0.19 ± 0.04
Total	100.00	100.00	100.00	100.00
n[c]	16	18	25	20

[a] One standard deviation is given.
[b] Total iron as Fe_2O_3
[c] Number of grains analyzed.

Table 7. Correlation coefficients for tephras[a] [b].

	ATC629	ATC630	ATC631	ATC632
Mazama	**0.95**	0.84	0.85	0.82
"Old Yn"	0.86	**0.99**	**0.98**	0.87
Yn	0.86	**0.99**	**0.97**	0.86
Bridge River	0.89	0.86	0.87	**0.95**

[a] The method of calculating correlation coefficients follows Davis (1985).
[b] The old Yn and Yn data is from Westgate and Gordon (1981).

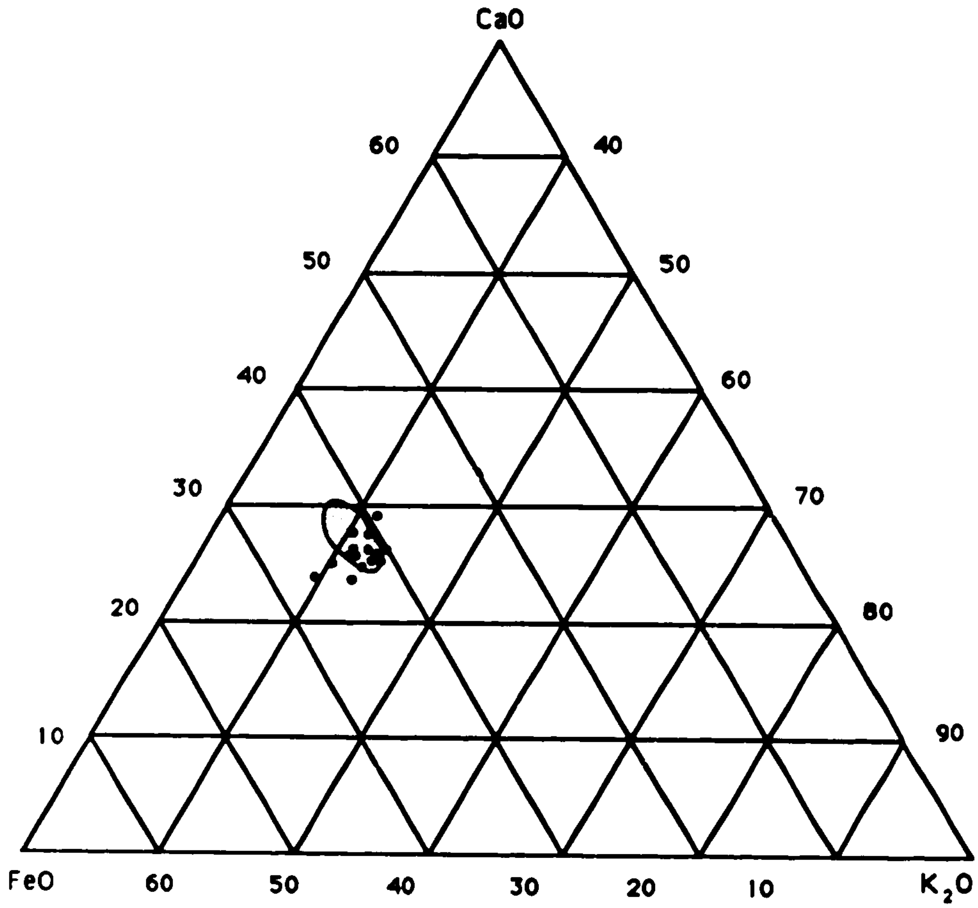


Figure 9. Ternary diagram showing the relative abundance of CaO, FeO, and K_2O in in sample ATC - 0629. The shaded area represents compositional data on the Mazama tephra from Westgate and Gorton (1981) and the circles represent data from this study.

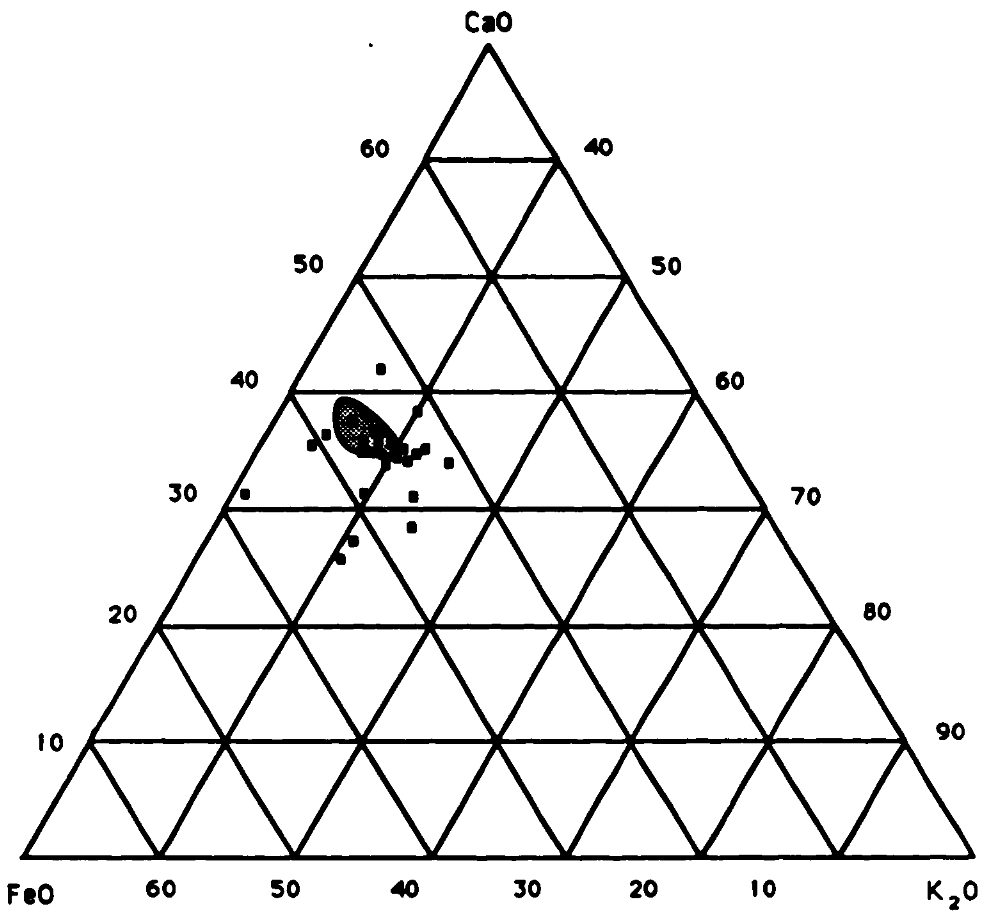


Figure 10. Ternary diagram showing the relative abundance of CaO, FeO, and K_2O in in sample ATC - 0630. The shaded area represents compositional data on the Mount St. Helen's Set Y tephra from Westgate and Gorton (1981) and the squares represent data from this study.

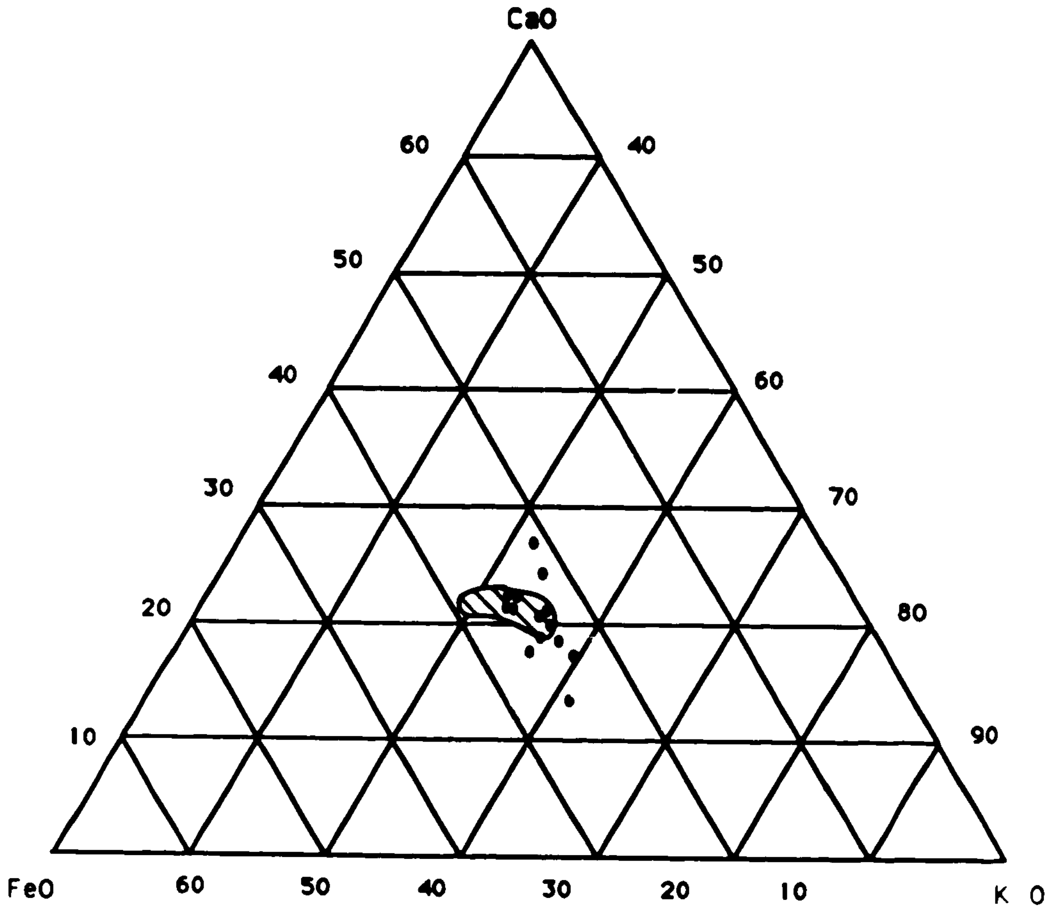


Figure 11. Ternary diagram showing the relative abundance of CaO, FeO, and K_2O in in sample ATC - 0632. The shaded area represents compositional data on the Bridge River tephra from Westgate and Gorton (1981) and the circles represent data from this study.

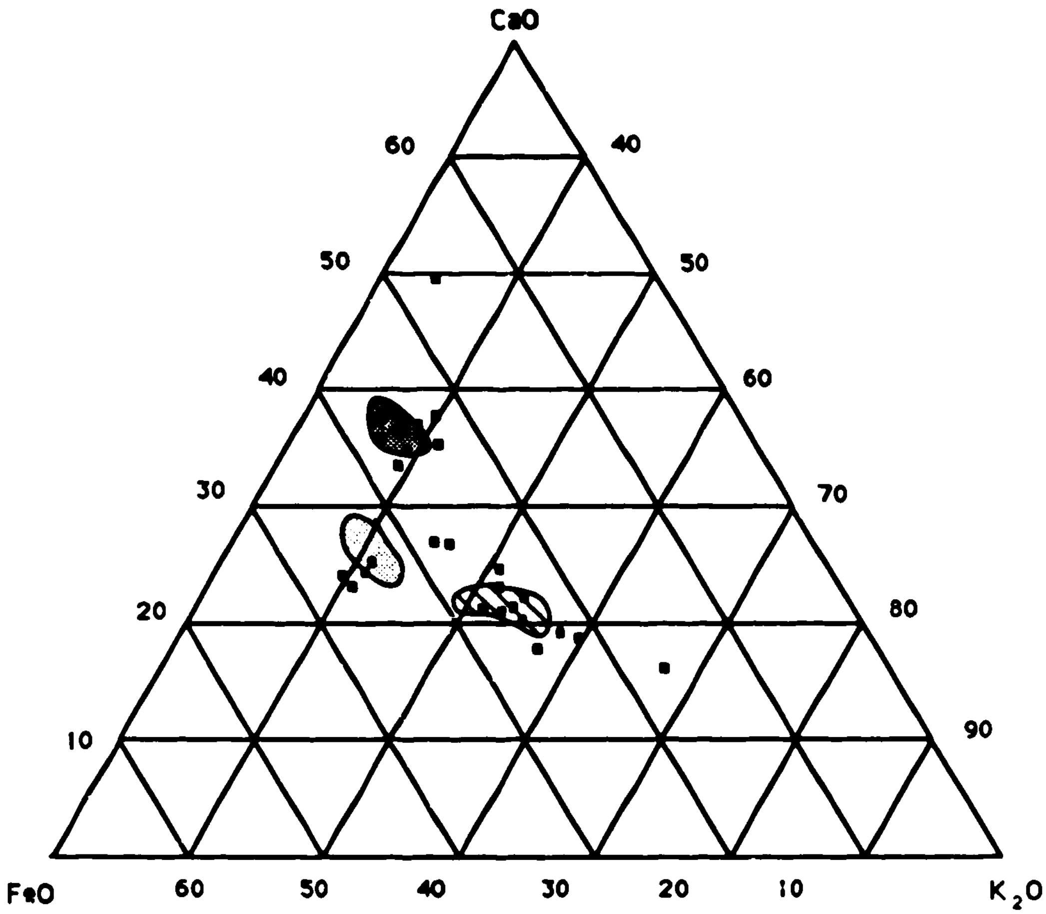


Figure 12. Ternary diagram comparing the relative abundance of FeO, K_2O, and CaO for sample ATC - 0631 (this study, represented by squares) to Westgate and Gorton's (1981) data of the relative abundances for Bridge River ▨ Mount St. Helen's Set Y ▩ and Mazama ☐ from Westgate and Gorton (1981).

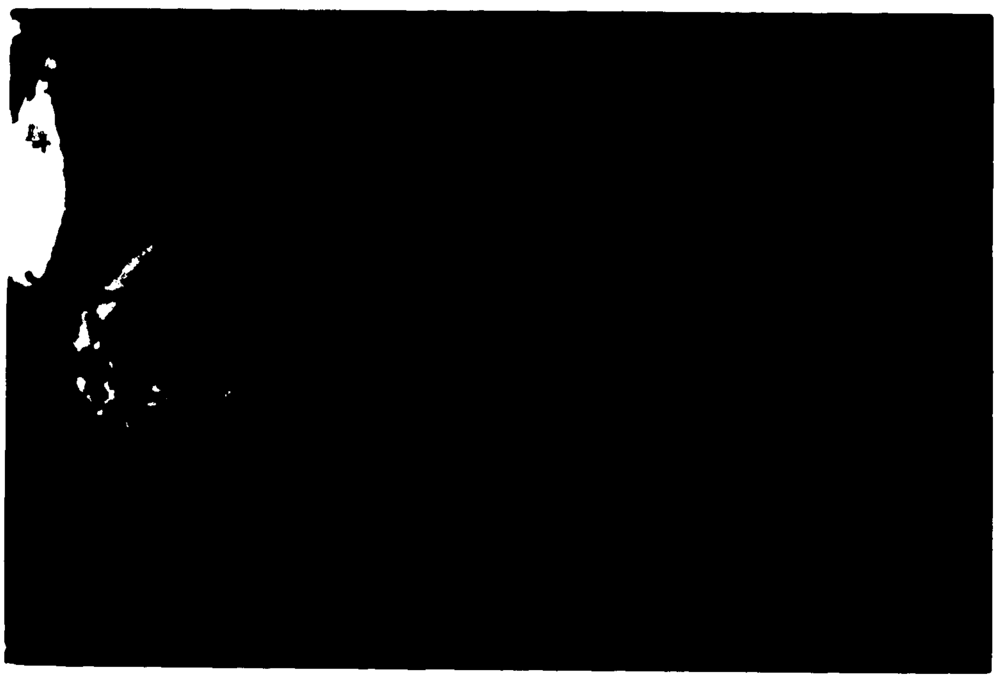


Plate 8. Light micrograph showing the morphology of a shard from the Mazama tephra.

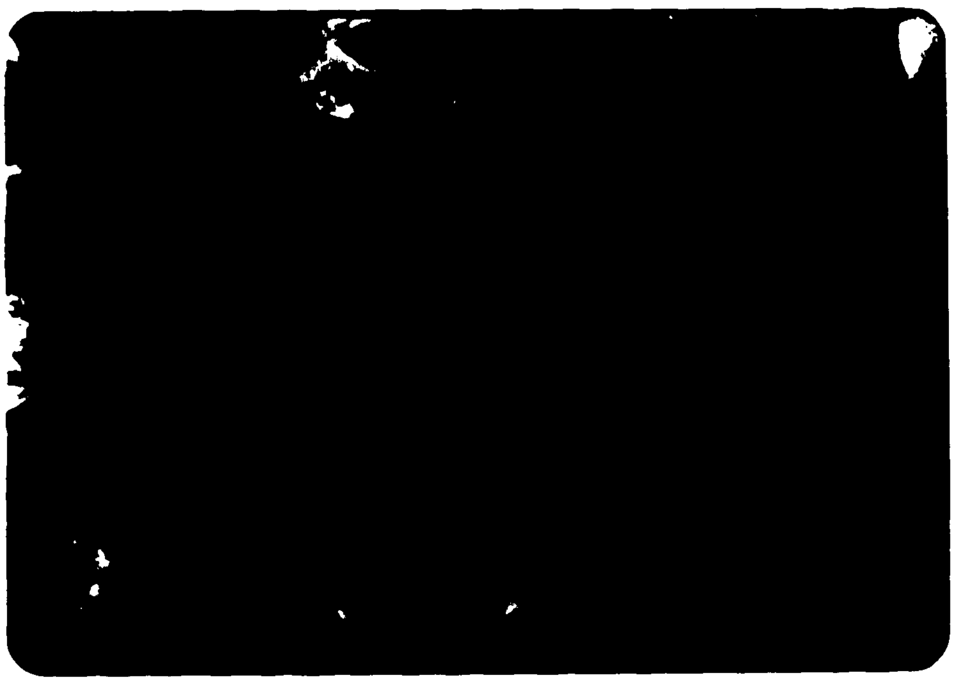


Plate 9. Scanning electron micrograph of a glass shard from the Mazama tephra.

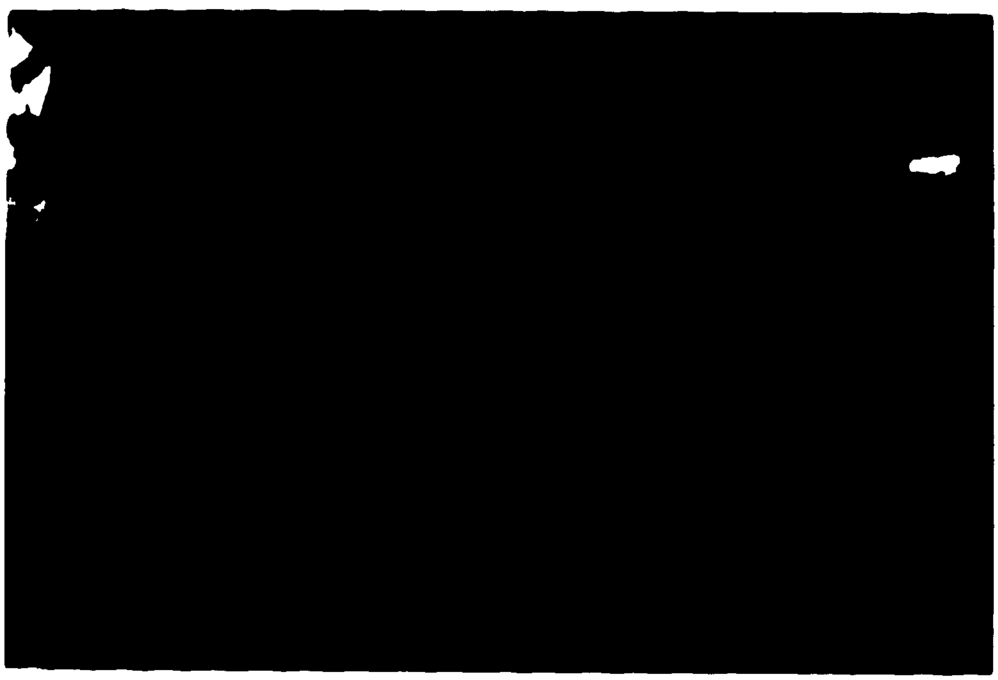


Plate 10. Light micrograph showing the typical morphology a shard from the Bridge River tephra.

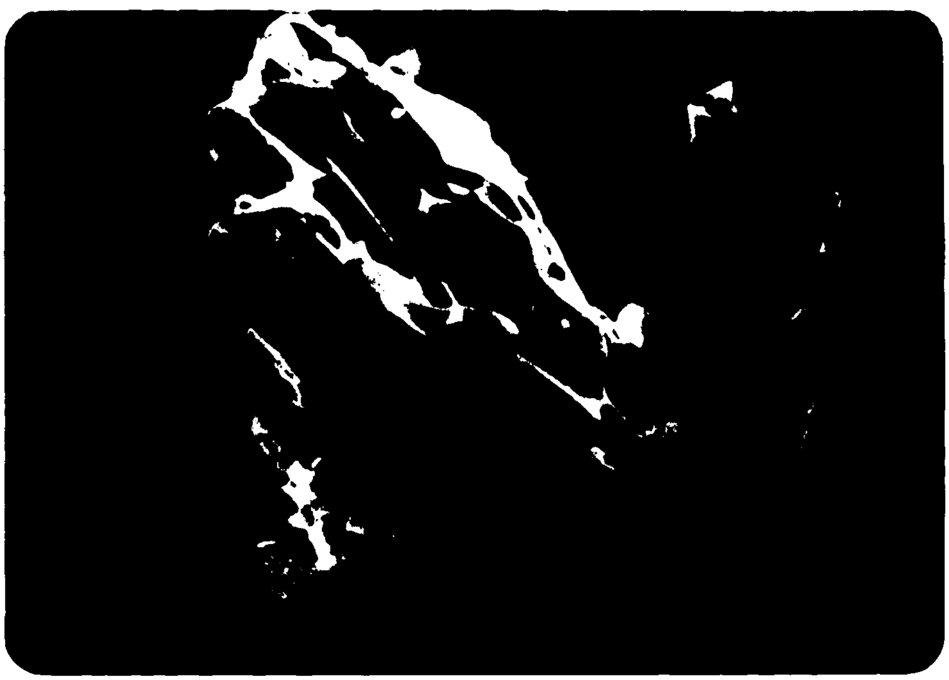


Plate 11. Scanning electron micrograph of a glass shard from the Bridge River tephra.

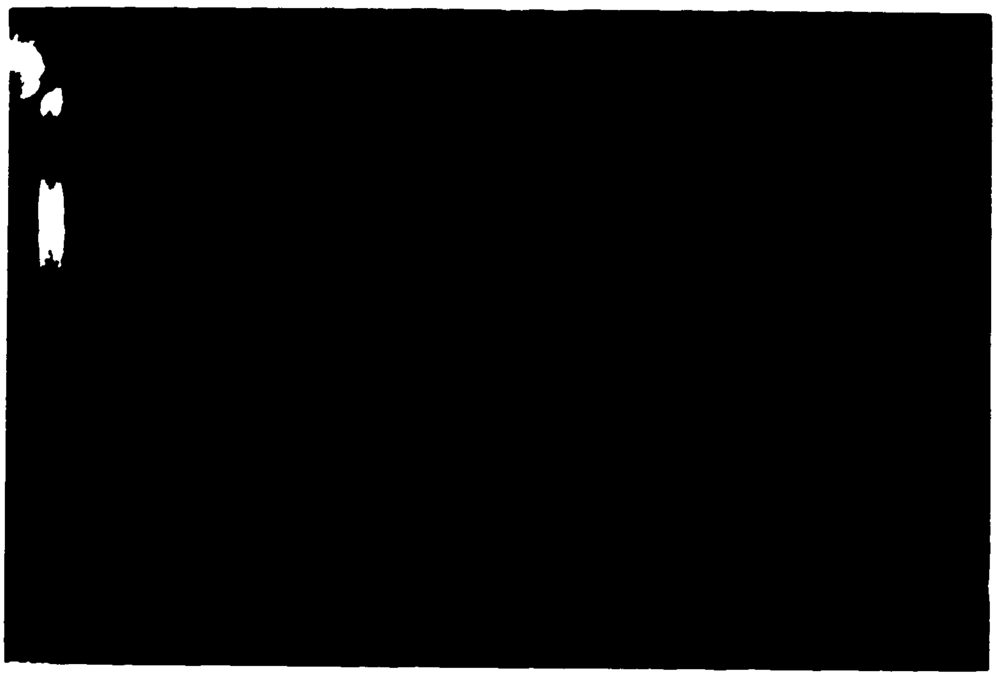


Plate 12. Light micrograph of a shard from the Mount St. Helen's set Y tephra.

Plate 13. Scanning electron micrograph of a glass shard from the Mount St. Helen's set Y tephra.

To aid in identifying sample ATC - 0631, correlation coefficients were calculated (Table 7). The data was "cleaned" by calculating compositional means and then removing points according to their statistical distance from the mean (Davis 1985). The match is made independently after the data is cleaned. A correlation is usually indicated for values of 0.95 or greater. The results show that ATC - 0629 is correlated with Mazama, ATC - 0630 and ATC - 0631 correlate with Westgate and Gorton (1981) values for Yn and for a stratigraphically older Yn and with each other, and ATC - 0632 correlates with the Bridge River Tephra.

3.4 Palynology

3.4.1 Core LOR3

The percent pollen stratigraphy of core LOR3 is presented in Figure 13. The zonation programs CONSLINK, SPLITINF, AND SPLITLSQ were applied to the data but no zones were delimited. Only *Pinus*, *Picea*, *Abies*, *Alnus*, *Betula* and occasionally *Salix* pollen percentages exceeded 2%. The percentage of arboreal pollen generally exceeds 65% while the total herbaceous pollen is not greater than 4%. *Pinus*, *Picea*, and *Alnus* pollen dominate the spectrum with little change in the profile of these pollen types, especially if 95% confidence limits are applied. *Corylus* pollen is present at the base of the core. *Picea* and *Abies* pollen reach maximum percentages in the pre - Mazama spectra (ca. 7, 300 to 7, 800 RYBP). *Picea* and *Abies* values decrease sharply following

Mazama. The initial percentages of *Pinus* pollen are high, but decline during the the period of high spruce and fir values. *Tsuga* pollen appears regularly in the pollen record ca. 4, 000 RYBP. Total pollen concentration values are highest in the pre - Mazama samples, and decrease near the top of the core. This is probably due to the high percent water content in the upper sediments.

3.4.2. Core LOR5

The percent pollen stratigraphy of core LOR5 is presented in Figure 14. The zonation programs, CONSLINK, SPLITINF, SPLITLSQ were used on these data, but no zones could be delimited. Generally, the arboreal pollen component exceeds 70%, and the herbaceous component is less than 4%. Only *Pinus*, *Picea*, *Abies*, *Salix*, *Alnus*, and *Betula* reach percentages greater than 2%. When 95% confidence limits are applied to the major taxa, it appears as if little change has taken place. *Corylus* pollen is present at the base of the core. In the pre - Mazama spectra, pine, spruce and fir pollen follow a similar pattern to core LOR3. *Pinus* values are highest at the base (80 - 85%), decrease only to increase again prior to the deposition of the Mazama tephra. When *Pinus* percent values decrease, *Picea* and *Abies* percentages increase, similarly, as values for *Pinus* increase, spruce and fir values decrease. *Tsuga* appears consistently in the pollen record at about the same time as it appears in LOR3. The percent values of *Picea*, *Abies*, and *Salix* are greater in LOR5 than in the deep water core. Total pollen concentration shows the same trend as LOR3 with decreasing concentrations in the upper samples.

Again, this is likely due to higher percent water content in the upper portion of the core compared to the lower portion.

4. Discussion

4.1 Chronology

The precise timing of deglaciation in the Canadian Cordillera is difficult to assess. In general, ice retreat began ca. 13, 000 RYBP and was nearly complete 3, 000 years later (Fulton 1984). Minimum dates of deglaciation from sites in the Rocky Mountains of Alberta and British Columbia are listed in Table 8 (modified from Osborn and Luckman 1988). Most of the dates are from valley sites or high alpine basins, there are only a few dates available from subalpine elevations in the Canadian Rocky Mountains.

Chronological control for this study is provided by tephra analysis and radiocarbon dating. The radiocarbon dates of 6, 310 ± 190 and 6, 980 ± 150 RYBP, obtained on gyttja from core LOR3 are considered to be incorrect as the dates bracket Mount St. Helen's set Y tephra, dated elsewhere in the park ca. 3, 400 RYBP (Luckman *et al* 1984). The basal date of 12, 350 ± 440 RYBP obtained on gyttja from core LOR3 is probably also incorrect as it appears contamination by carbon not in equilibrium with atmospheric sources is a problem at this site. The erroneous dates are examples of the difficulties associated with radiocarbon dating limnic sediments in areas with calcareous bedrock and soils, (Hickman *et al* 1984; MacDonald 1989, MacDonald *et al* 1984; White *et al* 1985). The radiocarbon dates from terrestrial macrofossils in core LOR4 are probably more accurate than the dates on gyttja because contamination from "old carbon" is less likely. If the standard

Table 8. Minimum dates of deglaciation from the Canadian Rockies (Alberta and British Columbia) (modified from Osborn and Luckman 1989).

Alpine and upper Subalpine sites

Site	Date (RYBP)	Reference
Wilcox Pass	9, 600 ± 305	Beaudoin (1986)
Sunwapta Pass	8,100 ± 100	Luckman *et al.* (1977)
Watchtower Basin	9 ,445 ± 375	Luckman & Kearney (1986)
Excelsior Site	8, 450 ± 170	Luckman & Kearney (1986)
Tonquin Pass	9, 660 ± 280	Kearney & Luckman (1983a)
Opabin Lake	8, 530 ± 190	Reasoner & Hickman (1989)
Lake O'Hara	10, 060 ± 160	Reasoner & Hickman (1989)

Lower Subalpine and Montane sites

Site	Date (RYBP)	Reference
Lorraine Lake	8, 740 ± 100	Bear & Hickman (in prep.)
Maligne Lake kettle	7, 560 ± 440[a]	Kearney & Luckman (1987)
Pocahantas (Athabasca Valley)[a]	11, 900 ± 200	Levson & Rutter (1985)
Elk Valley	12, 200 ± 160	Harrison (1976)
Southern Rocky Mountain Trench	11, 000 ± 180	Lowdon and Blake (1970)
Morley Flats, Kananaskis	10, 400 ± 110	MacDonald (1982)
Toboggan Lake, Kananaskis	10, 400 ± 70	MacDonald (1989)

[a] Date is from peat above base, extrapolated basal date ca. 8, 500 RYBP.

deviations from the radiocarbon dates on the terrestrial macrofossils are taken into account, the conventional radiocarbon date (9, 180 ± 320 RYBP) and the AMS date (8, 740 ± 100 RYBP) are near the same age. The deep portion of the basin may be older than the shallow portion, but it is difficult to determine unless a terrestrial fossil from the deep site is dated. If the pollen sequences are compared, core LOR3 and LOR5 appear to be close to the same age. Kearney and Luckman (1987) reported a date of 7, 650 ± 440 RYBP (BGS 466) from peat near the base of a core from a small kettle near the north shore of Maligne Lake. The date is probably too old as the Mazama tephra is absent, and radiocarbon dates from overlying sediments are very old. This again emphasizes the difficulties of radiocarbon dating in limestone areas. The basal dates from Lorraine Lake indicate when the basin filled, but do not aid in determining patterns of deglaciation for the area. The pollen records and macrofossils show that an established vegetation existed prior to the infilling of the basin, and that the area was ice free. Radiocarbon dates from high elevation sites on the northern flank of the Maligne Range, indicate that these areas have been deglaciated since 9, 500 RYBP (Luckman and Kearney 1986).

Sedimentation rates are calculated using the tephra dates and the basal date of 8, 740 RYBP. Sedimentation rates for cores LOR3, LOR4, and LOR5 are low and do not significantly differ from one another. Differences in sedimentation rates can be attributed to variations in water content of the sediments, and sediment focussing to the deep part of the basin. Pre - Mazama sedimentation rates are lowest in all cores. Reasoner (1988) also found that pre -

Mazama sedimentation rates for Lake O'Hara and Lake Opabin were lowest.

4.2 Palaeoecology

The oldest dated pollen records in the park range from 8, 450 ± 170 RYBP, to 9, 660 ± 280 RYBP (Kearney 1981; Kearney and Luckman 1983a). *Pinus*, *Picea* and *Abies* forests with a *Pteridium* dominated understory were established at higher elevations (above present levels) at this time (Kearney and Luckman 1983a). Subalpine forest was established in the Jasper Park area sometime after 9, 000 RYBP (Kearney and Luckman 1983a, 1987; Beaudoin 1984). The pollen records of LOR3 and LOR5 represent the oldest dated subalpine pollen stratigraphies in Jasper National Park (8, 740 ± 100 RYBP). The Lorraine Lake basin filled during the early Holocene. Climatic conditions were warmer and drier than present, but summer insolation values had declined to the point that basin flooding had begun in central Alberta (Schweger and Hickman 1989). The radiocarbon date from the pine scale provides a basal date for the lake as well as a minimum date for the establishment of *Pinus* at this site. The morphology of the pine scale closely resembles that of *Pinus contorta*, with a thickened scale tip and minutely curved prickle. The pollen records and the macroremains at the base of LOR4 also indicate that an established vegetation must have existed at this site by 8 740 ± 100 RYBP. The date of the establishment of pine in this area is in agreement with with the 8, 320 ± 80 RYBP date from pine wood recovered from the snout of the Athabasca glacier (Luckman 1989), and MacDonald and Cywnar

(1985) who noted the arrival of *Pinus contorta* ssp. *latifolia* at latitudes close to this site ca. 10, 000 RYBP.

There is little change evident in the Lorraine Lake pollen assemblages throughout the Holocene. Fluctuations in local timberline approached 200 m (Luckman and Kearney 1986, Kearney and Luckman 1987) but there is no evidence of altitudinal shifts in the subalpine vegetation. The apparent insensitivity of these pollen records and other subalpine records to the environment can be attributed to two main factors. Firstly, many taxa are over or under represented in the spectra. *Pinus* pollen is greatly over represented, as well as *Alnus*, while *Picea*, *Abies* and the herbaceous taxa are under represented. Secondly, in mountainous regions, a variety of microclimates are present which ultimately are responsible for the vegetation at each site (Bryson 1985). Climatic data are scarce for the subalpine, and topographical and elevational gradients, and aspect make vegetation unique at each site. Variations in aspect, topography and elevation make it difficult to draw comparisons and generalizations about the palaeoclimate and palaeoecology between subalpine pollen sites . Kearney and Luckman (1987) report that ca. 7, 500 RYBP, vegetation cover at the shallow kettle on the north shore of Maligne Lake was at its maximum extent, and due to the hydrologic and topographical characteristics at this site ca. 1, 300 RYBP, modern vegetation (discontinuous subalpine forest and wet meadows and fens) developed. It is difficult to compare these changes in this pollen record to the Lorraine Lake pollen spectra because of the inherent differences between the sites, and lack of chronological control.

Initial high values of *Pinus* in both cores conform to Luckman's and Kearney's (1986) pollen spectrum of Watchtower Basin. The concurrent increase of *Picea* pollen in both cores from ca. 7, 000 to 7, 500 RYBP may indicate that *Picea* displaced *Pinus* during this time. At Wat^htower Basin , Luckman and Kearney (1986) found that timberline elevation receded close to modern levels ca. 7, 300 to 7, 400 RYBP. The decline in timberline may correspond to the increase in *Picea* pollen in the Lorraine Lake cores due to wetter and cooler growth conditions. The *Corylus* pollen at the base of both cores may be indicative of the warmer and drier conditions of the early Holocene. *Corylus cornuta* is present in dry, sandy areas at low elevations (eg., Valley of the Five Lakes) in Jasper National Park (Kriebom, personal communication). In the pollen spectra, there is little evidence of the mid - Holocene Hypsithermal event postulated by many others Hickman *et al.* (1984), Luckman (1989), Luckman and Kearney (1986) MacDonald (1987, 1989), Ritchie (1976), Schweger and Hickman (1989) and Schweger *et al.* (1981). Small amounts ot *Tsuga* pollen appear consistently in the records after 4, 000 RYBP. Reasoner and Hickman (1989) and Beaudoin (1986) also found that *Tsuga* appears regularly about this time. *Tsuga heterophyllia* is presently found in the moist montane forest in the interior of British Columbia and in a few areas of Jasper National Park, therefore the *Tsuga* pollen is probably of regional origin. The increase in frequency about 4, 000 RYBP may indicate cooler and wetter regional conditions. Timberline elevation at Watchtower Basin had declined to modern levels ca. 4, 500 RYBP indicating cooler climatic conditions (Luckman and Kearney 1986). Beaudoin

(1986), Schweger and Hickman (1989), Schweger *et al.* (1981), Vance (1986), and Vance et al. (1983) also report regional cooler and wetter climatic conditions at about this time. After ca. 4, 000 RYBP, there is little change evident in the pollen spectra, which implies that modern vegetational trends were established at about this time.

The percent pollen stratigraphies of the deep and shallow cores are similar to one another. The pollen concentration values for the deep site were greater than the shallow site, presumably due to sediment focussing in the deep part of the basin. The zonation programs CONSLINK, SPLITLSQ, and SPLITINF were employed on each of the stratigraphies but no similar zones between the two pollen records were found. I feel this confirms my conclusion that the pollen diagrams showed no distinct zones.

The sediment stratigraphies of LOR3 and LOR5 may be indicative of warmer climatic conditions of the mid - Holocene. Precipitation of carbonate in limnic sediments is from the saturation of CO_3^- in the water column, and saturation is influenced by three factors: 1) an increase in water temperature, 2) biogenic factors, such as algal uptake of CO_2 during photosynthesis which may lead to an increased pH as CO_2 replacement from atmospheric sources is a slower process, or 3) high CO_3 input from the catchment basin (O'Sullivan 1983). The carbonate maximum at the base of LOR5 is likely a result of high input of carbonate from the watershed, combined with lower water levels. Mackereth (1965) demonstrated that lakes with active inflow and outflow streams, and catchment areas with calcareous tills that carbonate content of the sediments was highest following deglaciation. Carbonate is leached from the catchment area and marl is deposited in the shallow part of the lake where the concentration would be the greatest. In the deep part of the lake, less marl is deposited because of higher water levels and lower water temperature. The smaller carbonate peak in LOR3 is evidence of this phenomenon. The gradual change from marl sediments to gyttja in LOR5 indicates that water levels slowly changed. High values of carbonate cease abruptly following the deposition of the Mazama tephra in LOR5. The cessation of high carbonate deposition coincides with the onset of moist and cool climatic conditions reported from other sites in Alberta at this time (Hickman *et al.* 1984). The high concentration of mollusc shells in the marl sediments is another indicator that water levels were likely lower and water temperatures were warmer in the early Holocene.

In general, pigment concentrations reflect contemporary plant production so that fossil pigment stratigraphies can be used as an index of past productivity (Wetzel 1983). However, pigment diagenesis, past preservation conditions, differential degradation and varying detrital input are processes that are difficult to assess but will greatly affect the fossil pigment stratigraphies.

Enhanced pigment preservation conditions at the LOR3 site can account for the greater concentration of pigments, and the higher percent undegraded chlorophyll content. The deposition of pyrite spherules is by anoxic bacteria whose dead cells act as nuclei for pyritization (Dickman 1979). The formation of pyrite spherules is dependent upon anoxic conditions so that the bacteria will thrive, but deposition of pyrite is increased when a sudden influx of oxygenated water kills the bacteria (Dickman 1979). The irregular pyrite stratigraphy from LOR5 (Figure 14) could be indicative of changing oxygen conditions in the surficial sediments which could also affect conditions for pigment preservation. The pyrite stratigraphy from LOR3 (Figure 13) does not fluctuate as greatly as LOR5, perhaps indicating that oxygen levels at this site have remained more stable (anoxic) and pigment preservation is enhanced. The organic matter stratigraphies from LOR3 and LOR5 do not reveal any trends, although the organic matter content is higher in the shallow core. Conditions for algal growth in the shallow part of the lake are more favorable compared to the deep portion, hence more organic matter is deposited, even though pigment concentrations are lower.

The TAP stratigraphies show that the lake has slowly become more productive over the Holocene, especially following the deposition of the St. Helen's set Y tephra. The carotenoid stratigraphies also indicate that the lake is becoming more productive.

The oscillaxanthin stratigraphies of cores LOR3 and LOR5 closely follow one another. Oscillaxanthin is specific to the algal group Oscillatoriaceae, and at the lake's inception *Oscillatoria* spp. were more dominant than today. *Oscillatoria* spp. populations have fluctuated throughout the lake's history. The upper peaks of oscillaxanthin in LOR5 are not present in LOR3. This may be due to better conditions for epipelic algal growth at the shallow coring site. Myxoxanthophyll concentrations peak at the base of LOR3 following the large oscillaxanthin ·   During the mid - Holocene myxoxthanophyll concentrations i   are low which may be indicative of smaller populations. Myxoxanthophyll levels increase ca. 2, 400 RYBP and remain high, which follows the TAP trends of higher production in the late Holocene. The myxoxanthophyll stratigraphy from LOR5 is unlike LOR3. Low concentrations are present at the base of the core. During the mid - Holocene production levels remain fairly stable and do not decrease as in LOR3. Favorable growth conditions at the shallow water site may account for the difference. In the late Holocene, myxoxanthophyll concentrations increase, which seems to follow the trends of the TAP and LOR3 myxoxanthophyll.

4.3 Tephrostratigraphy

The widespread distribution of many tephras in northwestern North America provide stratigraphic and chronologic indicators for research in Holocene limnology, pedology, palynology, geology, and archeology. The identification of volcanic ash layers can be done by stratigraphic position, radiometric dating, shard morphology, measuring the refractive index of the glass, bulk chemistry, or grain discreet methods such as electron microprobe analysis, x - ray diffraction and neutron activation techniques. The occurrence of four ashes in one core has not previously been documented for the Jasper park area. Prior to this study, the Mazama tephra, and Mount St. Helen's Set Y ashes have been reported from the Maligne Valley by Luckman and Kearney (1986) and Kearney and Luckman (1987). Bridge River has not yet been reported from this area.

The source area for the Bridge River tephra is the Plinth - Meager Mountain area, British Columbia. The tephra occurs within a narrow belt through central British Columbia and some parts of southwestern Alberta (King 1986). The Bridge River tephra has been previously reported in the Sunwapta Pass region of Jasper National Park (Bowyer 1977). The Lorraine site lies on the outer edge of the plume, and the deposit is not thick. Despite this, the average glass compositions of this study show a general similarity to those reported by Westgate and Gorton (1981). The presence of biotite also confirmed the identity of this tephra. In the freshly split core, the tephra was an orange colour, which could mean that the tephra was reworked and in varying states of oxidation. Scanning electron micrographs of the tephra showed that the diatom frustules, and chrysophyte cysts were abundant in this layer. This implies that the ash settled slowly to the bottom sediments.

The Mazama tephra originates from Mount Mazama, Oregon, and is the most widespread Holocene tephra in northwestern North America. It is dated at about 6, 800 RYBP (Bacon 1983). Blinman *et al.* (1979) found several layers which comprise the tephra set at Wildcat Lake, Washington. In the southern Canadian Rockies only one ash layer is present, although in some lakes in southern Alberta two ash layers have been reported (Hickman, personal communication). In the Lorraine Lake cores, one Mazama layer was found which varied in thickness from 3 to 6 cm. An unknown tephra is present in LOR4, embedded in the basal sand layer, 70 cm below Mazama tephra. The shard morphology resembles that reported for the Mazama tephra (plates 8 and 9 a-b). There are three possible explanations which may account for the presence of this ash: 1) contamination from further up in the core barrel, 2) the tephra is a reworked deposit of Mazama from upper sediments, or 3) this may be an earlier eruption from Mazama, or an unknown source. The first explanation is unlikely as the unknown tephra was embedded in the basal sand layer, indicating that contamination from the core barrel was not a factor. Either of the latter two explanations are more likely. As previously stated, Blinham *et al.* (1979) have reported more than one ash belonging to the Mazama set, but it is thought that the eruptions took place only hundreds of years apart. The ash may belong to the Glacier Peak tephra set, but the tephra has not yet been found in Alberta. The settling of volcanic ash on surficial lake sediments is not well documented, and the ash may have sunk through the unconsolidated fluid sediments at the lake bottom.

Members of the Mount St. Helen's Y tephra set have been reported from many sites in the foothills, and Rocky Mountains in Alberta (Kearney and Luckman 1987; Dumanski 1970, Dumanski *et al.* 1980, Bowyer 1977; King 1984; Beaudoin 1984). There are three ashes in this set, Yn (3, 510 ± 200 years B.P.), Yb (3, 900 ± 250 years B.P.), and Ye (3, 350 ± 250 years B.P.) (Mullineaux *et al.* 1975). There has been some confusion as to the dates of the set Y ashes found in Alberta. The available dates can be divided into two groups, ca. 4, 300 RYBP and ca. 3, 400 RYBP (Westgate and Gorton 1981). Two ashes are present in all of the Lorraine Lake cores. Dumanski *et al.* (1980) reported two members of the St. Helen's Yn tephra set near Hinton, Alberta. He found that the older member (ca. 4 300 RYBP) was generally well preserved, and that the younger Yn member was poorly preserved and only was identifiable upon occasion. This description matches the tephrostratigraphy at Lorraine Lake. The thicker set Y unit has been identified as the Yn layer. Electron microprobe analysis show that the composition of the thiner Y ash layer seems to have characteristics of Mazama, Mount St. Helen's and Bridge River. Correlation coefficients show this tephra to be most closely related to set Y. There are two possible explanations which account for the presence of this ash: 1) erosion of the shore disturbed a section of St. Helen's and Mazama ash which became mixed and was deposited in the basin or, 2) this unit represents a separate eruption which incorporated considerable amounts of non - juvenile material. The glass shards are oxidized and this would be consistent with either redeposition of ash from [illegible] or incorporation of older material during the eruption. The latter explanation is unlikely as the plume from the youngest eruption of the set Y group, Ye, is documented as travelling southeast. Therefore, if the thin Lorraine Lake tephra is Ye, the plume had a different trajectory. If the ash layers at this site correspond to those of Dumanski *et al.* (1980) then the radiocarbon date of 3 400 RYBP from the large Y ash from Tonquin Pass (Luckman *et al.* 1984) does not conform with the Lorraine Lake tephrostratigraphy. The former explanation is the most probable one, although it also seems unlikely as the ash is found in all sediment cores across the lake basin, and no other evidence of a major erosional event is found.

5. Conclusions

1) The basal date from Lorraine Lake does not aid in determining patterns of deglaciation for this area; however the area was likely ice - free as an established vegetation existed prior to the infilling of the basin, 8, 740 ± 100 RYBP.

2) The estimated arrival of *Pinus contorta* corresponds to the AMS date on the ovuliferous scale (8, 740 ± 100 RYBP).

3) Generally, subalpine pollen spectra show little change, but with careful observation, conclusions can be reached which pertain to regional climate. No different trends are observed between the pollen spectra from deep versus shallow sites in this lake.

4) Warmer climatic conditions existed in the early Holocene as evidenced by the sediment stratigraphy and the presence of *Corylus* pollen.

5) The regular appearance of *Tsuga* in the pollen spectra ca. 4, 000 RYBP may indicate that climatic conditions became cooler and wetter at this time. This corresponds to findings by other researchers in the Jasper area and for the rest of Alberta.

6) The fossil pigment stratigraphies indicate that the lake has become more productive in the latter Holocene.

7) The blue - green alga *Oscillatoria* spp. seems to have invaded the lake initially, then fluctuated sporadically throughout the Holocene.

8) Fossil pigment concentrations and fluctuations at each coring site was ultimately determined by the oxygen concentration in the surficial sediments.

9) This is the first time that two possible Mazama ash members have been reported in northern Alberta, but further work is necessary to confirm the identity of the unknown ash.

10) The presence of two St. Helen's tephras in Alberta is strengthened by this study, but more glass chemistry and radiocarbon dating must be done before any positive conclusions can be drawn.

11) The trajectory of the Bridge River ash can be extended further northwest.

BIBLIOGRAPHY

Aaby, B. 1979. Characterization of peat and lake deposits. In: Palaeohydrological changes in the temperate zone in the last 15 000 years, Vol. II, Bjorn E. Berglund (Editior). pp. 77 - 98.

Bacon, C.R. 1983. Eruptive History of Mount Mazama and Crater Lake Caldera, Cascade Range, U.S.A. In: S. Aramaki and I. Kushiro (Editors), Arc Volcanism. Journal of Volcanology and Geothermal Research, 18 57 - 115.

Bear, R.B., Hickman, M., and J. Beget (1989). Electronmicroprobe analysis of four Holocene tephras in a limnic sediment core, from Lorraine Lake, Jasper National Park, Alberta (In preparation).

Bear, R.B., and M. Hickman. (1989). The Holocene pollen stratigraphy from a subalpine lake in Jasper National Park, Alberta. (In preparation).

Beaudoin, A.B. 1984. Holocene environmental change in the Sunwapta Pass area, Jasper National Park. Ph.D. Thesis. University of Western Ontario, London, Ontario 487 pp.

Beaudoin, A.B. 1986. Using *Picea/Pinus* ratios from the Wilcox Pass core, Jasper National Park, Alberta, to investigate Holocene timberline fluctuations Géographie Physique et Quaternaire XL: 145 - 152.

Bengtsson, L. 1979. Chemical Analysis. In: Palaeohydrological changes in the temperate zone in the last 15, 000 years, Vol. II, Bjorn E. Berglund (Editior). pp. 113 - 132.

Birks, H.J.B. 1979. Guideline for numerical treatment of data. In: Palaeohydrological changes in the temperate zone in the last 15 000 years, Vol. I, Bjorn E. Berglund, (Editior). pp. 99 - 120.

Birks, H.J.B. and H.H. Birks. 1980. Quaternary Palaeoecology. Edward Arnold, London 289 pp.

Birks, H.J.B. and A.D. Gordon. 1985. Numerical Methods in Quaternary Pollen Analysis. Academic Press, London 317 pp.

Blinman, E.C., Mehringer, P.J., Jr., and J.C. Sheppard. 1979. Pollen influx and the depositon of Mazama and Glacier Peak tephra. In: Volcanic Activity and Human Ecology, P.D. Sheets and D.K. Grayson (Editors), pp. 393 - 425. Academic Press, New York.

Bombin, E.R. 1982. Holocene paleolimnology of Mary Gregg Lake, Foothills of the Alberta Rocky Mountains, Canada. M Sc. Thesis. University of Alberta. 147 pp.

Bowyer, A. 1977. The Sunwapta section: Composition and development of a complex stratigraphic section from Sunwapta Pass, Jasper National Park. M.Sc. Thesis. University of Alberta.

Brown, S., and B. Coleman. 1963. Oscillaxanthin in lake sediments. Limnology and Oceanography, 8: 352 - 353.

Bryson, R.A. 1985. On Climatic Analogs in Paleoclimatic Reconstruction. Quaternary Research, 23: 275 - 286.

Crandell, D.R., and D.R. Mullineaux. 1973. Pine Creek volcanic assemblage at Mount St. Helens, Washington. U.S. Geological Survey Bulletin, 1383 - A, 23p.

Crandell, D.R., and D.R. Mullineaux. 1978. Potential hazards from future eruptuins of Mount St. Helens volcano, Washington. U.S. Geological Survey Bulletin, 1383 - C, 26 p.

Davis, J.O. 1985. Correlation of late Quaternary tephra layers in a long pluvial sequence near Summer Lake Oregon Quateranry Research, 23: 38 - 53.

Dickman, M.D. 1979. A possible varving mechanism for meromitic lakes. Quaternary Research, 11: 113 -124.

Dumanski, J. 1970. A micromorphological investigation on the genesis of soils developed on calcareous aeolian material. Thesis, University of Alberta.

Dumanski, J., Pawluk, S., Vucetich, C., and D.J. Lindsay. 1980. Pedogenesis and tephrochronology of loess derived soils, Hinton, Alberta. Canadian Journal of Earth Sciences, 17: 52 - 59.

Faegri, K., and J. Iverson. 1975. Textbook of Pollen Analysis. Blackwell, Oxford. 295 pp.

Fulton, R.J. 1984. Quaternary glaciation, Canadian Cordillera. In: Quaternary Stratigraphy of Canada - A Canadian Contribution to IGCP Project 24. Geological Survey of Canada, Paper 84-10, pp. 39 - 48.

Guilizzoni, P., Bonomi, G., Galanti, G., and D. Ruggiu. 1982. Basic trophic status and recent development of some Italian lakes as revealed by plant pigments and other chemical components in sediment cores. Memorie dell'Istituto Italiano di Idrobiologia, 40: 79 - 98.

Habgood, T. 1978. Key to principal pollen types of Alberta. Photocopied manuscript, 18 pp.

Hansen, H.P. 1949a. Postglacial forests in west - central Alberta, Canada. American Journal of Botany, 36: 54 - 65.

Hansen, H.P. 1949b. Postglacial forests in west - central Alberta, Canada. Bulletin of the Torrey Botanical Club. 76: 278 - 289.

Hansen, H.P. 1952. Postglacial forests in the Grande Prairie - Lesser Slave Lake region of Alberta, Canada. Ecology, 33: 31 - 40.

Harrison, J.E. 1976. Dated organic material below Mazama (/) tephra: Elk Valley, British Columbia. Report of Activities, Part C. Geological Survey Canada, Paper 76-1C, pp. 169 - 170.

Heusser, C.J. 1956. Postglacial environments in the Canadian Rocky Mountains. Ecological Monographs, 26: 263 - 302.

Hickman, M., and G. Klarer 1981. Palaeolimnology of Lake Isle, Alberta, Canada (including sediment chemistry, pigments and diatom stratigraphy). Archive für Hydrobiologie, 91: 490 - 508.

Hickman, M., Schweger, C., and T. Habgood. 1984. Lake Wabamun, Alberta: a paleoenvironmental study. Canadian Journal of Botany, 63: 1438 - 1465.

Holland, W.D. and G.M. Coen. 1982. Ecological (Biophysical) Land Classification of Banff and Jasper National Parks, Vol. II: Soil and Vegetation Resources. Alberta Institute of Pedology Publication SS-82-44, 540 pp.

Kearney, M.S. 1981. Late Quaternary Vegetation and Environmental History of Jasper National Park, Alberta. Ph.D Thesis. University of Western Ontario, London, Ontario, 318 pp.

Kearney, M.S. and B.H. Luckman. 1981. Evidence for late Wisconsin/early Holocene climatic change in Jasper National Park. In: W.C. Mahaney (Editor), Quaternary Paleoclimate. Geo Abstracts Ltd., Norwich, England. pp. 85 - 106.

Kearney, M.S. and B.H. Luckman. 198 Holocene vegetational and climatic history of Tonquin Pt--, British Columbia. Canadian Journal of Earth Sciences, 20: 776 - 786.

Kearney, M.S. and B.H. Luckman. 1983b. Holocene timberline fluctuation in Jasper National Park, Alberta. Science, 221: 261 - 263.

Kearney, M.S. and B.H. Luckman. 1987. A mid - Holocene vegetational and climatic record from the subalpine zone of the Maligne Valley, Jasper National Park, Alberta (Canada). Palaeogeography, Palaeoclimatology, Palaeoecology., 59: 227 - 242.

King, R. 1984. Quaternary paleosols in the Canadian Rockies and their significance for establishing a chronologic framework. In: Correlation of Quaternary Chronologies, W.C. Mahaney (Editor), pp. 243 - 258. Geo Books, Norwich, U.K.

King, R. 1986. Weathering of Holocene airfall tephras. In: Rates of Chemical Weathering of Rocks and Minerals, S.M. Colman, and D.P. Dethier (Editors), pp. 239 - 264. Academic Press, Inc., Orlando.

Kuchar, P. 1972. Ecological impact study of the Maligne Lake area, Jasper National Park. Canadian Department of Indian Affairs, and Northern Development. Natural History Parks Branch.

Levson, V. and N. Rutter. 1985. The last glacial in Jasper National Park. Program with Abstracts, Canqua Symposium on the Paleoenvironmental Reconstruction of Late Wisconsin Deglaciation and the Holocene, p. 39.

Lichti - Federovich, S. 1970. The pollen stratigraphy of a dated section of late Pleistocene lake sediments from central Alberta. Canadian Journal of Earth Science, 7: 938 - 945.

Lichti - Federovich, S. 1972. Pollen stratigraphy of a sediment core from Alpen Siding, Alberta. Geological Survey of Canada, Paper 72-1, Rept. of Act. Part B: 113 - 115.

Litherland, A. 1985. The promise of atom counting. In: Dating methods of Pleistocene deposits and their problems, N.W. Rutter (editor). Geoscience Canada, Reprint series 2.

Lorenzon, C.S. 1967. Determination of chlorophyll and pheo - pigments: spectrophotometric equations. Limnology and Oceanography, 12: 343 - 346.

Lowdon, J.A., and W. Blake. 1970. Geological Survey of Canada Radiocarbon Dates IX. Geological Survey of Canada, Paper 70-2B, 41 pp.

Luckman, B.H. 1989. 8, 000 year old wood from the Athabasca Glacier, Alberta. Canadian Journal of Earth Sciences., 25: 148 - 151.

Luckman, B.H., and M.S. Kearney. 1986. Reconstruction of Holocene changes in alpine vegetation and climate in the Maligne Range, Jasper National Park, Alberta. Quaternary Research, 26: 244 - 261.

Luckman, B.H., Kearney, M.S., King, R.H., and A.B. Beaudoin. 1984. Revised ^{14}C age for St. Helen's Y tephra at Tonquin Pass, Britsh Columbia. Canadian Journa of Earth Sciences, 23: 734 - 736.

MacDonald, G.M. 1982. Late Quaternary paleoenvironments of the Morley Flats and Kananaskis Valley of southwestern Alberta. Canadian Journal of Earth Sciences, 19: 23 - 35.

MacDonald, G.M. 1987a. Postglacial development of the subalpine - Boreal transition forest of western Canada. Journal of Ecology., 75: 303 - 320.

MacDonald G.M. 1987b. Postglacial vegetation history of the MacKenzie River Basin. Quaternary Research, 28: 245 - 262.

MacDonald G.M. 1989. A fossil pollen, charcoal, and plant macrofossil record form the subalpine forest-grassland ecotone of Alberta: new insights on the postglacial vegetation and climate history of the Canadian Rocky Mountains and adjacent foothills. Palaeogeography, Palaeoclimatology, Palaeoecology, (in press).

MacDonald, G.M., and L.C. Cwynar. 1985. A fossil pollen based reconstruction of the late Quaternary history of lodgepole pine (*Pinus contorta* ssp. *latifolia*) in the western interior of Canada. Canadian Journal of Forest Research. 15: 1039 - 1044.

MacDonald, G.M., Beukens, R.P., Kieser, W.E. and D.H. Vitt. 1987. Comparative radiocarbon dating of terrestrial plant macrofossils and aquatic moss from the ice - free corridor of western Canada. Geology, 15: 837 - 840.

Mackereth, F.J.H. 1966, Some chemical observations of lake sediments and their interpretation. Philosphical Transactions of the Royal Society of London, 161 pp.

Marker, A.F.H. 1972. The use of acetone and methanol in the estimation of chlorophyll in the presence of phaeophytin. Freshwater Biology, 2: 361 - 385.

Mathewes, R.W., and J. Westgate. 1980. Bridge River tephra: revised distribution and significance for detecting old carbon errors in radiocarbon dates of limnic sediments in southern British Columbia. Canadian Journal of Earth Sciences, 17: 1454 - 1461.

Mosimann, J.E. 1965. Statistical Methods for the Pollen Analyst. In: B. Kammel and D. Raup (Editors) Handbook of Paleontological Techniques, pp. 636 - 673.

Moss, B. 1980. Ecology of Freshwaters. Blackwell Scientific Publications, Oxford.

Mullineaux, D.R., Hyde, J.H., and Rubin, Meyer, 1975. Widespread late glacial and postglacial tephra deposits from Mount St. Helens volcano, Washington. U.S. Geological Survey Journal of Research, 3: 329 - 335.

Mullineaux, D.R., Wilcox, R.E., Ebaugh, W.F., Fryxall, R. and M. Rubin. 1978. Age of the last major scabland flood of the Columbian Plateau in eastern Washington. Quaternary Research, 10: 171 - 180.

O'Sullivan, P.E. 1983. Annually - laminated lake sediments and the study of Quaternary environmental changes - a review. Quaternary Sciences Reviews, 1: 245 - 313.

Osborn, G., and B. Luckman. 1988. Holocene glacier fluctuations in the Canadian Cordillera (Alberta and British Columbia). Quaternary Science Reviews, 7: 115 - 128.

Olsson, I.U. 1979. Radiometric Dating. In: Palaeohydrological changes in the temperate zone in the last 15, 000 years, Vol. II, Bjorn E. Berglund, (Editior). pp. 1 - 33

Oyama, M., and N. Takehara. 1970. Revised standard soil colour charts. 2nd ed. Research Council for Agriculture, Forestry, and Fisheries, Tokyo.

Porter, S.C. 1978. Glacier Peak tephra in the north Cascade Range, Washington; stratigraphy, distribution and relationship to late - glacial events. Quaternary Research, 10: 30 - 41.

Reasoner, M.A. 1986. An inexpensive, lightweight percussion core sampling system. Geographie physique et Quaternaire, XL: 217 - 219.

Reasoner, M.A. 1988. The late Quaternary lacustrine record from the upper Cataract Brook Valley, Yoho national Park, British Columbia. 201 pp.

Reasoner, M.A., and R.E. Healy. 1986. Identification and significance of tephras encountered in a core from Mary Lake, Yoho National Park, British Columbia. Canadian Journal of Earth Sciences, 24: 1991 - 1999.

Reasoner, M.A., and M. Hickman. 1989. Late - Quaternary environmental change in the Lake O'Hara region, Yoho National Park, British Columbia. Palaeogeography, Palaeoclimatology, Palaeoecology, (in press).

Ritchie, J.C. 1976. The late - Quaternary vegetational history of the western interior of Canada. Canadian Journal of Botany, 54: 1793 - 1818.

Roed, M.A. 1964. Geology of the Maligne Valley. Alberta Research Council, open file report 1964 - 1.

Sanger, J.E. and E. Gorham. 1972. Stratigraphy of fossil pigments as a guide to the postglacial history of Kirchner Marsh, Minnesota. Limnology and Oceanography, 17: 840 - 854.

Sarna - Wojcicki, A.M., Champion, D.E., and J.O. Davis. 1983. Holocene volcanism in the conterminous United States and the role of silicic volcanic ash layers in correlation of latest - Pleistocene and Holocene deposits. In: Late Quaternary Environments of the United States, Vol. 2, H.E. Wright Jr. (Editor), pp. 52 - 77.

Sarna - Wojcicki, A.M., Meyer, C.E., Woodward, M.J., and P.J. Lamothe. 1981. Composition of airfall ash erupted on May 18, May 25, June 12, July 27, and August 7. In: The 1980 eruptions of Mount St. Helens, Washington, U.S.. Geological Survey Professional Paper 1250, pp. 667 - 682.

Schweger, C. 1976. Late Quaternary Paleoecology of the Onion Portage Region, Northwestern Alaska. Ph.D. Thesis. 180 pp.

Schweger, C., and M. Hickman. 1989. Holocene paleohydrology, central Alberta: Testing the general circulation model climate simulations. Canadian Journal of Earth Sciences, (in press).

Schweger, C. Habgood, T. and M. Hickman. 1981. Late - glacial Holocene climatic changes of Alberta. The record from lake sediment studies. In: The impacts of climatic fluctuations on Alberta resources and environment. Environment Canada Report No. WAES 181, pp. 47 - 59.

Smith, D.G.W., and J. Westgate. 1969. Electron probe technique for characterizing pyroclastic deposits. Earth and Planetary Science Letters, 5: 313 - 319.

Stockmarr, J. 1971. Tablets with spores used in absolute pollen analysis. Pollen et Spores, 13: 615 - 621.

Swain, E.B. 1985. Measurement and interpretation of sedimentary pigments. Freshwater Biology: 15: 53 - 76

Troels - Smith, J. 1955. Karakterisering af løse jordarter. Danmarkes Geoløgiske Undersøgelse, Series IV, 3 (10).

Vallentyne, J.R. 1955. Sedimentary chlorophyll determinations as a paleobotanical method. Canadian Journal of Botany, 35: 304 - 313.

Vance, R.E., Emerson, D. and T. Habgood. 1983. A mid - Holocene record of vegetative change in central Alberta. Canadian Journal of Earth Sciences, 20: 364 - 376.

Vance, R.E. 1986. Aspect of postglacial climate of Alberta: calibration of the pollen record. Geographie physique et Quaternaire, 40: 153 - 160.

Westgate, J.A. 1977. Identification and significance of late Holocene tephra from Otter Creek, southern British Columbia, and localities in west - central ALberta. Canadian Journal of Earth Sciences: 14: 2593 - 2600.

Westgate, J.A. and R.W. Mathewes. 1980. Bridge River tephra: Revised distribution and signifcance for detecting old carbon errors in radiocarbon dates of limnic sediments in southern British Columbia. Canadian journal of Earth Sciences, 14: 1454 - 1461.

Westgate, J.A. and M.P. Gorton. 1981. Correlation techniques in tephra studies. In: Tephra Studies. S. Self and R.S.J. Sparks (Editors). North Atlantic Treaty Organization Advanced Studies Institute Series C, D. Reidel Publishing Company, pp. 73 -94.

Wetzel, R.G. 1970. Recent and postglacial production rates of a marl lake. Limnology and Oceanography, XV no. 4: 491 - 503.

Wetzel, R.G. 1983. Limnology. Saunders College Publishing, Philadelphia.

White, J.M., Mathewes, R.W., and W.H. Mathews. 1985. Late Pleistocene chronology and environment of the "Ice - Free Corridor" of northwestern Alberta. Quaternary Research, 24: 173 - 186.

White, J.M. and R.W. Mathewes. 1986. Postglacial vegetation and climatic change in the upper Peace River district, Alberta. Canadian Journal of Botany, 64: 2305 - 2318.

Züllig, H. 1981. On the use of carotenoid stratigraphy in lake sediments for detecting past developments of phytoplankton. Limnology and Oceanography, 26: 970 - 976.

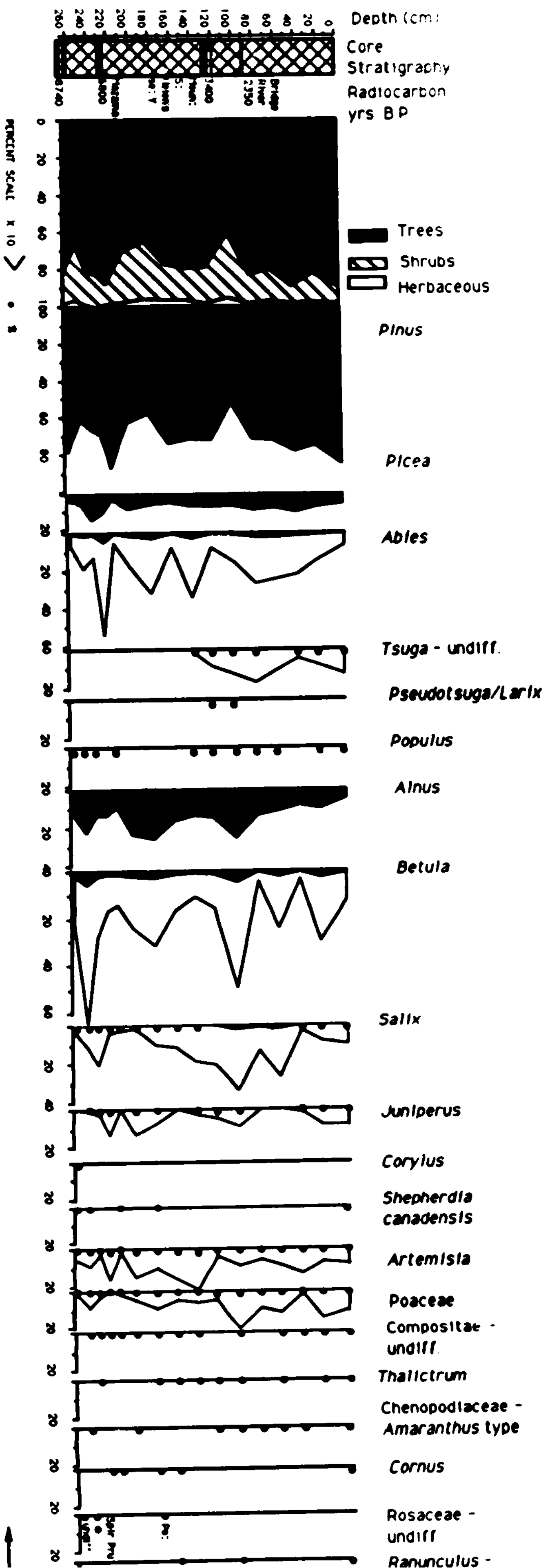


Figure 13. Relative percent diagram from core LCR3. Arc = *Arceuthobium*, Cru = Cruciferae, Eri = Ericales, Pot = *Potentilla*, Rum = Rumex, Spir = *Spiraeae*, Umb = Umbelliferae, Undiff = undifferentiated, Val = *Valerania*, Ver = *Veronica*.

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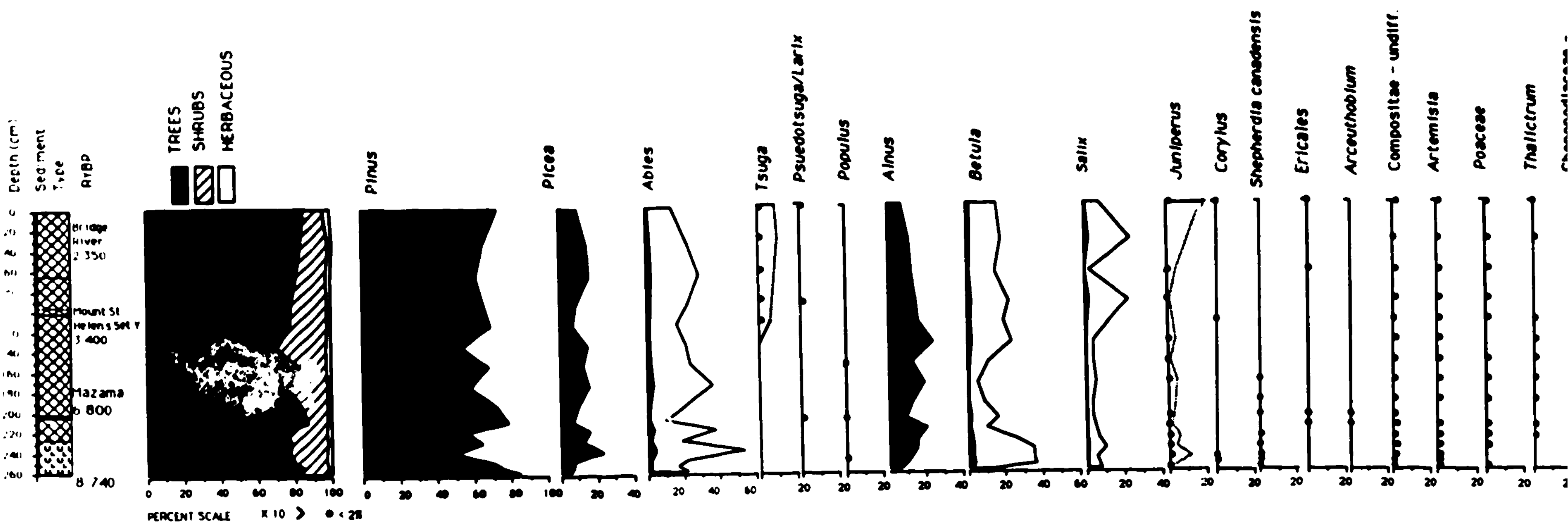

Figure 14 Relative percent pollen diagram for LORS. Cru = Cruciferae, Lig = Liguliflorae, Sax = Saxifragaceae, Umb = Umbelliferae, Val = *Valeriana*, Ver = *Veronica*.