University of Southampton

Faculty of Engineering, Science and Mathematics

School of Geography

The interaction of fire, climate and vegetation in the boreal forest of Alaska-Yukon during the Holocene

by
Leanne Franklin-Smith, B.Sc., M.Sc.

A Thesis presented
in partial fulfilment of the requirements for the degree of
Doctor of Philosophy

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Dedicated in loving memory to Derek Franklin
he took care of the land all his life

I've stood in some mighty-mouthed hollow
That's plumb-full of hush to the brim;
I've watched the big, husky sun wallow
In crimson and gold, and grow dim,.....
.....It's the great, big broad land 'way up yonder,
It's the forests where silence has lease;
It's the beauty that thrills me with wonder,
It's the stillness that fills me with peace.

'Extract from: The Spell of the Yukon by Robert W Service

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ABSTRACT

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The interaction of fire, climate and vegetation in the boreal forest of Alaska-Yukon, during the Holocene

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The circumpolar boreal forest covers large areas of the high northern latitudes, and is important for feedbacks between land surface and atmosphere. Large carbon stores, fires and a dynamic ecotone with tundra mean that altered disturbance regimes in the boreal system are likely to have major impacts on climate via feedback mechanisms affecting the surface energy budget and atmospheric chemistry.

Using radiocarbon-dated sedimentary records with the temporal focus on two key vegetative transitions (deciduous-*Picea* and *Picea-Pinus*) and a climatic transition (cold and dry to moister), high resolution time series of charcoal-peak frequencies from lake sediments are used as a proxy of the local fire regime.

Modern observations lead to multiple hypotheses as to how a major climate change might interact with the regional vegetation and fire regime. Warm, dry conditions might be expected to encourage more frequent fires, and above some threshold frequency, deciduous forest may predominate. Thus, a moister climate regime would lead to a reduction in fire frequency and subsequent conifer dominance. Conversely, modern observations indicate that overall flammability of *Picea* is greater than that of boreal hardwoods. If climate controlled the vegetation distribution, and a moister climate favoured the spread of *Picea* alterations in the fire regime would be expected to follow the vegetation change, and that vegetation controlled the fire regime.

The regional vegetation transition from deciduous- to coniferous-dominated forest at ~10ka BP displays a clear sequence where the climate shift precedes the alteration in vegetation composition, to which the fire regime responds. The deciduous vegetation experienced low levels of burning, with a lower fire frequency than when *Picea* became dominant on the landscape, suggesting that *Picea* was excluded from the landscape due to moisture limitations rather than high fire return frequencies.

In the Yukon Territory, *Pinus contorta* (lodgepole pine) is migrating northwards and westwards towards Alaska, and is considered a potential invasive species to the northern boreal forest of Alaska under global warming. Lodgepole pine is a fire-dependent species that appears to thrive and spread when fires are intense and frequent. Analysis of stomata reveals lodgepole pine was present in the Southern Yukon forests, at least in low numbers, by ~6 ka BP, much earlier than conventional pollen records suggest. The main population expansion (represented by increased *Pinus* pollen from <5 to >15%) was regionally asynchronous, and occurred over 3 ka after the first appearance of *Pinus*. Contrary to expectations derived from flammability estimates and modern observations that pine stands burn particularly frequently, there is no clear, sustained increase in charcoal peak frequency in the late-Holocene *Pinus* zone; *Pinus-Picea* forests appear to have burned under a regime similar to that of the preceding *Picea*-dominated forests.

Contents

List of Tables

List of Figures

List of Plates

List of Abbreviations

AL	Aleutian Low
CHAR	Charcoal Accumulation Rate
LLNL	Lawrence Livermore Laboratory
LOI	Loss on Ignition
LPP	Lodgepole Pine
NERC	Natural Environment Research Council
NPI	North Pacific Index
Poz	Poznan Radiocarbon Laboratory
TBP	Time Between Peaks
WRA	White River Ash tephra

Site Abbreviations:

LHLZ	Little Harding Lake, core Z
BL97b	Birch Lake, core 97b
SLA	Salmo Lake, core A
DFLA	Dragonfly Lake, core A
HCLA	Haircut Lake, core A
MLA-Z	Marcella Lake, amalgamated core A (Anderson, 2005) and core Z (LMFS, 2004)
JBL	Jellybean Lake, core C

Table of Analysts

ABBREVIATION	ANALYST	AFFILIATION
LMFS	Leanne Franklin-Smith	University of Southampton
NHB	Nancy Bigelow	University of Alaska, Fairbanks
APK	Andrea Krumhardt	University of Alaska, Fairbanks
LA	Lesleigh Anderson	University of Massachusetts, Amhurst
JRB	Joanna Baker	University of Southampton, undergraduate
SMH	Sian Hill	University of Southampton, undergraduate
KJG	Katherine Gallagher	University of Southampton, undergraduate

Table of Analyses

LAKE	CORED BY	LOI	MAGNETIC SUSCEPTIBILITY	POLLEN	BLACK/WHITE SPRUCE	CHARCOAL	DATING
Little Harding	LMFS	LMFS / APK	LMFS	LMFS	n/a	LMFS	LMFS / APK
Birch	NHB	NHB / APK	NHB	NHB / APK / LMFS	n/a	LMFS	LMFS / NHB
Salmo	LMFS	LMFS	LMFS	LMFS / SMH	KJG	LMFS	LMFS / APK
Dragonfly	LMFS	LMFS	LMFS	LMFS	KJG	LMFS	LMFS
Haircut	LMFS	LMFS / APK	APK	LMFS / JRB	n/a	LMFS	LMFS / APK
Marcella	LMFS / LA	APK / LA	APK / LA	LMFS / APK	KJG	LMFS	LMFS / LA
Jellybean	LA	LA	LA	n/a	n/a	LMFS	LA

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Chapter 1. Introduction

1.1 Background

The extreme environmental events occurring around the world over just the last few years, such as Hurricane Katrina in the Gulf Coast of the United States, the heat waves and prolonged periods of drought across Europe, and the most recent severe fire seasons in Alaska and the Yukon, serve as a reminder of just how vulnerable the human race is to variations in the natural environment. Predicted climate change indicates that global temperatures are set to rise. Boreal and arctic ecosystems are particularly sensitive to global warming, and they also play an important part in the feedback pathways of the Earth system that determine the dynamics of the climate system. Disturbance events, such as fire, are especially important in the dynamics of these ecosystems, and alterations in climatic conditions, either human-induced, or due to natural variability, have an important influence on the disturbance regime. As the boreal zone now increasingly becomes inhabited and exploited, the role of fire as a threat to infrastructure and modifier of the resource base is becoming a major concern (Chapin *et al.*, 2003), especially in the light of two consecutive severe fire years.

The circumpolar boreal forest, the second largest forest ecosystem on the planet, represents a wood resource of global significance, both environmentally and economically. During the past ~10,000 years the boreal forest of Alaska-Yukon has experienced dramatic shifts in climate (e.g. Bartlein *et al.*, 1998; Bigelow *et al.*, 2003) and vegetation (e.g. Ager & Brubaker, 1985; Anderson, P.M & Brubaker, 1994; Edwards & Barker, 1994; Bigelow *et al.*, 2003). In more recent times climatic warming has been associated with thawing of permafrost (e.g. Viereck, 1973; Brown, 1983), modified growth rates of dominant tree species (e.g. Barber *et al.*, 2000), increases in area burned (e.g. Chapin *et al.*, 2000; Gavin *et al.*, 2003) and insect outbreaks (Schweger, 1997). The causal links between these changes and their implications for the functioning of the boreal forest are not completely understood. The structure and function of the boreal forest both determines and is influenced by its disturbance regime, which is driven by fire, insect outbreaks, logging and flooding. The extent and distribution of these disturbances is changing rapidly as the climate warms (Chapin *et al.*, 2003; Chapin *et al.*, 2004a; Chapin *et al.*, 2004b). To understand the current and future structure, diversity and functioning of the boreal forest, it is important to understand how climate and disturbances interact and the implications of these interactions for the ecological if not economic, and cultural sustainability of the boreal forest.

The understanding of vegetation development across Alaska-Yukon has advanced extremely rapidly over the past few decades (P.M. Anderson, *et al.*, 2004). However, our understanding of the role of ecosystem disturbance in vegetation development has lagged far behind, even in the light of several studies noting the relationship between vegetation type and fire frequency, the long-term relationships between disturbance, climate and vegetation are at best incompletely understood.

The boreal forest is important for feedbacks between land surface and atmosphere (Bonan *et al.*, 1992; French *et al.*, 2000; Chapin *et al.*, 2000), as large carbon stores, fires and a dynamic ecotone with tundra mean that alterations of the disturbance regime are likely to have major impacts on climate via feedback mechanisms affecting the surface energy budget and atmospheric chemistry (e.g., Levis *et al.,* 1999).

Model-based studies have addressed possible futures for the boreal ecosystem under global warming. While global/hemispheric-scale simulations, which currently do not take into account disturbance regimes, indicate considerable expansion of boreal forest at its northern limit (e.g., Kaplan *et al.* 2003), ecologically based models highlight the possibility of unexpected switches from evergreen conifers to deciduous or even open vegetation (e.g., Starfield and Chapin 1996), in part due to shifts in the disturbance regime that are climatically controlled – the primary disturbance type being fire. It is generally assumed that the large climate change anticipated with rising atmospheric CO_2 will influence boreal fire regimes (Kasischke *et al.,* 2000; Stocks *et al.*, 2000). However, climate changes similar to those that may affect fire regimes under future global warming have not occurred within recent experience, and the nature of disturbance regimes under a greatly changed climate remains conjectural. Palaeodata can throw light on changes in vegetation cover and fire regime during the Holocene, which has experienced much larger climate shifts than the recent decades covered by the instrumental record and detailed fire data.

In interior Alaska and Yukon Territory, (*Picea*) spruce-dominated (flammable) boreal forest became established in place of deciduous scrub and woodland (least flammable woodland type) ~ 10 ka BP, coincident with a major climate shift from warm and dry to moister conditions (Barber and Finney 2000; Edwards *et al.* 2005). In the Yukon, invasion of *Pinus contorta* (*lodgepole* pine), most flammable species, has further altered forest composition (Johnstone and Chapin 2003) over the past 1-3 ka and is a possible model for the modification of Alaskan spruce forest under global warming.

Deciduous woodland and spruce and pine forest can all occur as end-successional vegetation within their current geographic ranges on the contemporary landscape, and these communities are associated, as far as it is possible to determine, with different mean fire return intervals. Differences appear in part due to the varying flammability of the different species: *Pinus*>*P. mariana*>*P. glauca*>deciduous taxa (Johnson, 1992). However it appears that the most drought- and fire-prone sites are currently dominated by deciduous taxa. These are fast growing and can sprout vegetatively, which better adapts them to a rapid fire return cycle (Viereck *et al.*, 1992). Furthermore, fire-season weather determines regional patterns of fire occurrence. Thus, both vegetation and climate have controls on the fire regime.

Such major climate and vegetation transitions provide long-term 'natural experiments' with which to observe the interaction of fire and vegetation. As both climate and vegetation (flammability, fuel loads etc) have controls on the fire regime, the first key question concerns the hierarchy of control: does climate affect fire and hence vegetation, or does vegetation, responding directly to climate, determine fire regime? Outcomes of a given climate shift for ecosystem structure, composition, and function, may differ, depending on how these various factors interact.

Although the current climatic influences on fire behaviour in the boreal forest have been well studied and documented (Johnson, 1992; Flannigan *et al.*, 2003), the influences on fire occurrence and extent in Alaska and the Yukon Territory are complex (Kasischke *et al.*, 2002). The limited data on palaeo-fire occurrence are contradictory in part because of this. There are studies that suggest higher fire episode frequencies occurred in the early Holocene (Earle *et al.*, 1996) when some models suggest climate was warmer and drier than today (Kaufman *et al.*, 2004), while others document higher frequencies for the mid- to late-Holocene (Hu *et al.*, 1996; Lynch *et al.*, 2002; Lynch *et al.*, 2004) when climate became cooler and wetter (R.S Anderson,. *et al.*, 2006).

The second key question concerns the extent to which the invasion of the highly flammable pine (predicted to accelerate under global warming) alters the fire regime of an area: does it necessarily induce an increase in fire return frequency, or are the prevailing climate and site factors important?

This project builds on previously obtained regional records of vegetation and climate change in interior Alaska and the Yukon as part of LTER VI (Long Term Ecological Research Project) a National Science Foundation (USA) initiative. It will provide reconstructions of long-term fire regimes across key vegetation and climate transitions in order to assess the relative importance of climatic versus species composition factors in determining fire history during a period of change. It will also draw conclusions in association with a palaeohydrologic reconstruction of the area and contribute to the growing knowledge base for high latitude interactions with climate, which are projected to be among the most vulnerable to global warming (Serreze *et al.*, 2000; Moritz *et al.*, 2002).

1.2 Aims and Objectives

The main aim of this project is to test hypotheses about how climate, fire, and vegetation interact to bring about large-scale change in the boreal forest ecosystem of Alaska-Yukon using palaeodata.

- *H1. Climate change drives fire regime, which alters vegetation type*

This is tested against the transition to spruce from deciduous vegetation. A climatic shift (equivalent to a 20-35% increase in precipitation; Barber and Finney, 2000) at ~ 10 ka BP suppressed burning causing a switch in dominance from deciduous taxa to spruce. For this hypothesis is to be supported, a reduction of the fire regime should be observable across the transition.

- *H2. Climate driven vegetation change alters the fire regime.*

This is tested against both vegetation transitions (deciduous-spruce, spruce-pine). The dominant vegetation types have differing flammabilities (deciduous < spruce < pine), and therefore a shift in vegetation type prior to a change in the fire regime would be expected. In this instance there should be little change or an increase of fire across the vegetation shift, based on contemporary flammability observations. The transition from spruce to pine is not associated with a clear climatic shift, and therefore provides a form of control, as any observed changes in the fire regime can therefore be solely attributed to vegetative drivers.

1.3 Thesis Structure

The thesis consists of 9 main chapters; a brief outline of each is presented below from Chapter 2 onwards:

- Chapter 2 reviews the available literature regarding the main palaeoenvironmental method employed in this project – charcoal analysis. In addition it provides a synthesis of available palaeoclimatic and palaeoenvironmental data from Eastern Beringia relating to vegetation composition and structure, and it discusses climatic reconstructions throughout the Holocene and their implications to fire regime.
- Chapter 3 presents the methodologies undertaken and/or developed during the research.
- Chapter 4 provides descriptions of the study sites and their surrounding areas.
- Chapter 5 develops the chronology for each site that informs the interpretation of all the proxies used in the study.
- Chapter 6 presents and interprets the vegetation history of each site based on pollen analysis and physical sedimentary properties.
- Chapter 7 presents and discusses the charcoal record, and the charcoal model outputs, comparing and contrasting the fire history for the major vegetation types present at each site.
- Chapter 8 develops a synthesis of the site-specific data and provides a regional interpretation and discussion of the results in comparison with other known studies in the area.
- Chapter 9 presents the main findings of the project and covers the problems within the current areas of palaeoecological understanding along with recommendations for future research in an attempt to bridge these gaps.

Chapter 2. Previous Research

2.1 Introduction

This Chapter aims to orientate the reader towards the study area and the region of east Beringia. To those ends, the chapter is divided topically, to describe the study area, outline the current and past climate regimes, and reconstruct the vegetation history of the area. The majority of the Chapter outlines the main topics necessary to understand how charcoal can be used to interpret and reconstruct past fire regimes. Modes of production, transportation and deposition within a lake catchment are discussed, and the methods utilized to produce a charcoal record are explored. Fire regime reconstructions from the area are compared and contrasted.

2.2 Site Specifics and Physical Geography

2.2.1 Alaska

Alaska occupies the extreme northwestern region of the North American continent (fig 2.1), and is separated from Asia by the 51-mile wide Bering Strait. It is the largest of the United states, with a total area of 663,267 miles^2, a large proportion of which lies within the Arctic Circle.

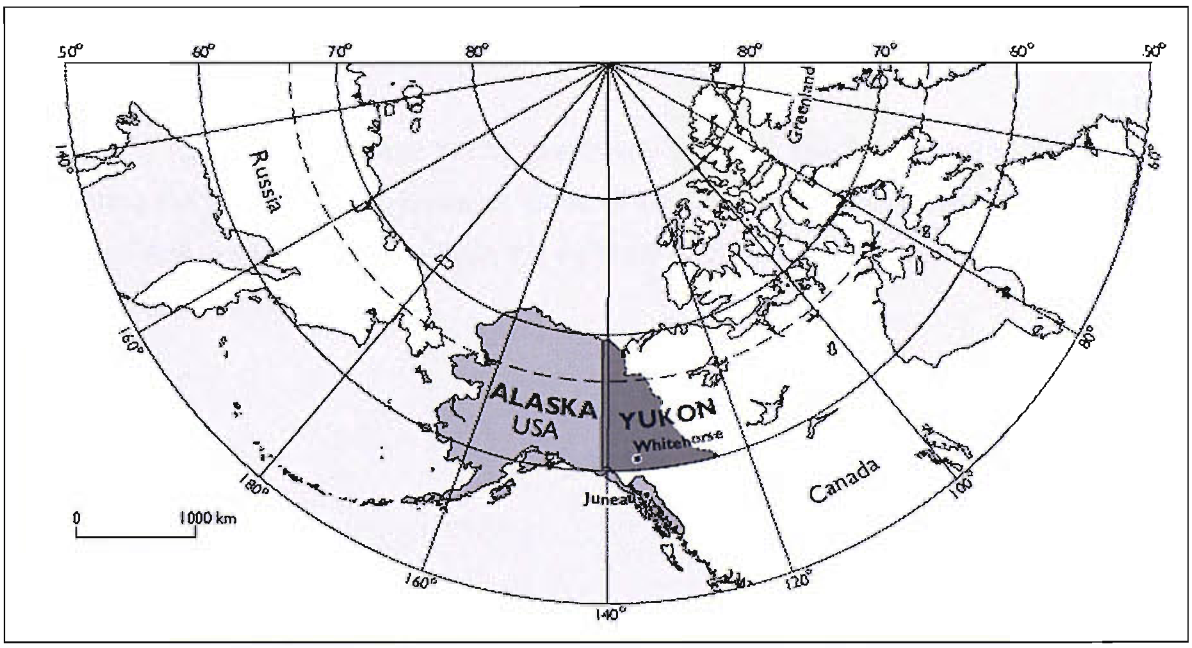


Figure 2.1: Location map of Alaska and the Yukon Territory in the circumpolar belt

Alaska has a greater area of lakes and rivers than any other US state, with a high concentration in the interior, where this study is based. The Alaskan Interior is bounded by the Alaska Range to the south, the Brooks Range to the north, the border with the Yukon Territory on the east, and the Bering Sea on the west (Figure 2.4). The glaciated Brooks Range separates Interior from Arctic Alaska.

The topography of interior Alaska is a mixture of flat alluvial plains with high mountains and rolling hills. The alluvial plains, particularly the Tanana and Yukon basins, are underlain primarily by fine-grained silt and sand (Bigelow, 1997). In the lowland areas these thick deposits contain abundant ice-rich permafrost, which impedes drainage, and leads to widespread paludification (Viereck *et al.*, 1986). In general the hills are mantled with wind-deposited loess. South-facing hill slopes receive abundant solar radiation and are not underlain with ice. The north-facing slopes however, are much colder and can contain significant amounts of ice. At the higher elevations the loess mantle is thin or completely absent, as is the ice rich permafrost.

2.2.2 Yukon Territory

The Yukon Territory is the westernmost territory of Canada (fig 2.1), bordering Alaska, lying north of 60 ^{o}N and partly within the Arctic Circle, covering an area of 186,270 miles^2. The entire territory belongs to the physiographic province of Canada called the Cordilleran Region, or Cordillera. The topography away from the river valleys, is generally rugged and rolling with an average elevation of about 1,200 m (about 4,000 ft) above sea level (Phillips, 1990).

The Yukon Territory was the most easterly largely unglaciated area of Beringia (fig 2.2) during the last glaciation; however lobes of the Cordillerian and Laurentide ice sheets advanced and retreated repeatedly in the extreme south-east of the territory until ~10,000 yrs BP (Anderson, 2004).

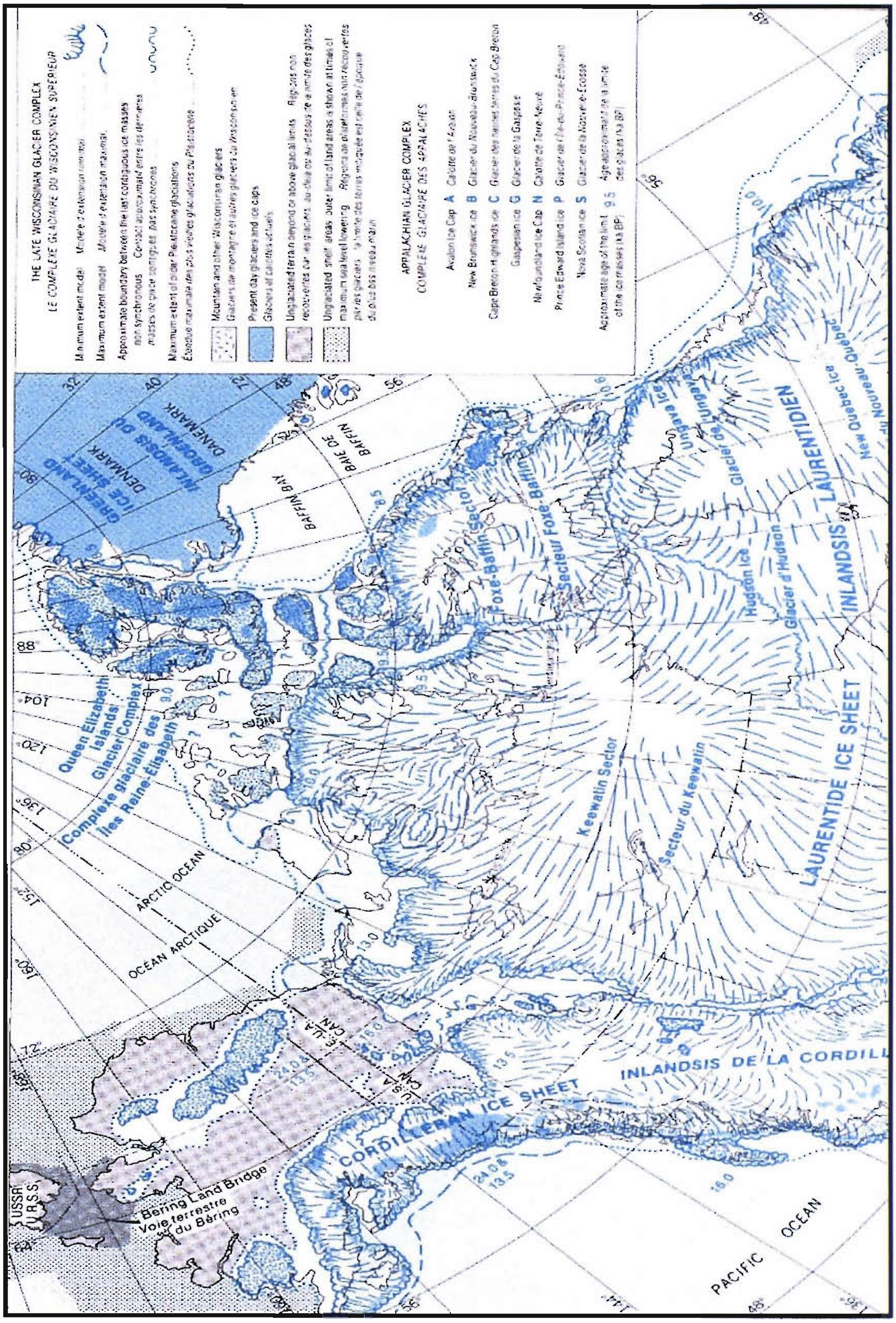


Figure 2.2: Glacier positions in relation to study area, at maximum glacial extent. (Source: Energy, Mines and Resources, Canada (Energy, 2004))

2.3 Modern Climate

2.3.1 Alaska

North and central Alaska is dominated by boreal forest and tundra (Viereck & Little, 1975; Viereck *et al.*, 1986). Global distribution of these biomes (fig 2.3) is related to climatic factors such as average seasonal temperatures (Larsen, 1965, 1980), air masses (Larsen, 1971; Hare & Ritchie, 1972), potential evapotranspiration (Hare, 1950), net radiation (Hare & Ritchie, 1972), and combined temperature-moisture-air mass indices (Tuhkanen, 1984). The relationship between climate and vegetation patterns differs in Alaska compared to other arctic/sub-arctic regions of North America (Anderson & Brubaker, 1994). For example the general north-south distribution of boreal forest and tundra in North America is modified in western Alaska by east-west vegetation gradients that reflect maritime influences (Hare and Ritchie, 1972). During the winter, there is a reversal of the normal pattern of decreasing temperature with elevation (Phillips, 1990), which results in a shallow, intense inversion being present almost continuously from late October, to early March and on occasion in the summer. Thunderstorms are more frequent than in most other high latitude areas, with a long season lasting from May to September.

Figure 2.3: Global location map of Taiga/ Boreal Forest biome. (Source: WWF Canada (WWF Canada, 2006)).

Interior Alaska is broadly categorised as the area north of the Alaska Range, south of the Brooks Range and east of the Seward Peninsula (Figure 2.4a), and is classified as sub-arctic, with a predominately cold continental climate. Summers are short and hot, and the winters are long and cold. Mean annual temperature at Fairbanks is -3.5°C, with mean January temperature of -22.4°C, and mean July temperature of 14.7°C (Bigelow, 1997), and the annual temperature extremes are great (e.g., +38 to -60°C). Mean annual precipitation is ~280.5mm, with most falling as rain during the summer and early autumn (Reiger *et al.*, 1979) (Figure 2.4b). The region is classified as semi-arid due to the fact that it experiences moisture deficits resulting from the distance from the ocean, and the presence of large topographic barriers such as the Alaska and Brooks ranges (Barber & Finney, 2000). These characteristics at lower latitudes would classify Central Alaska as a desert; however, due to the presence of widespread permafrost, the surface of which melts in summer, much of the low elevation areas are in fact poorly drained. In contrast, hillsides (especially south facing slopes), can be moisture-limited because of the high solar radiation and low precipitation (Wesser & Armbruster, 1991).

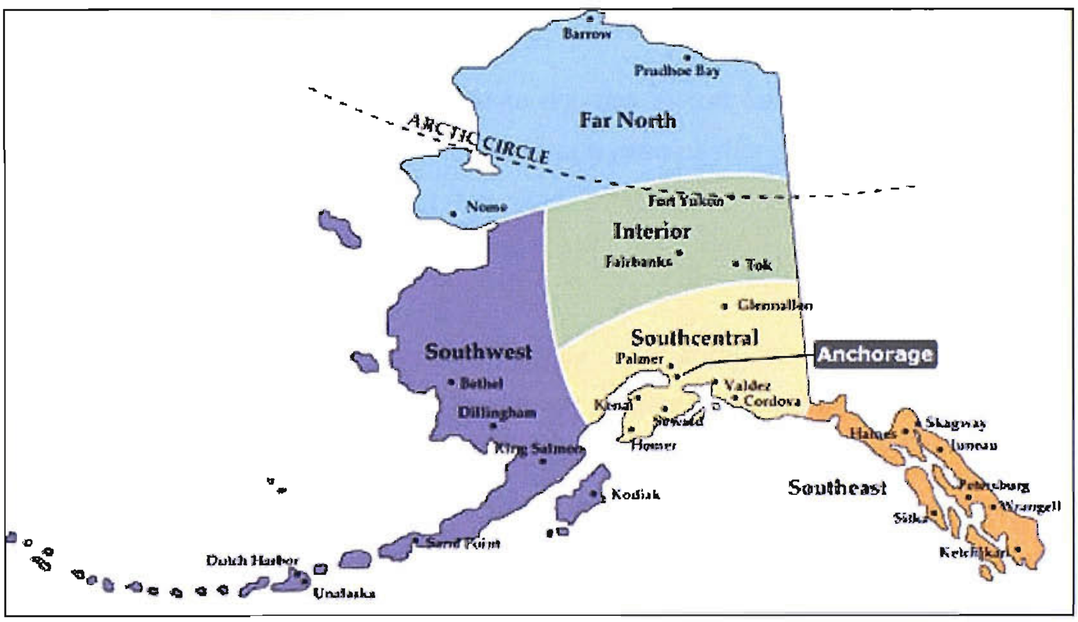


Figure 2.4a: Location map of the main Alaskan regions, with the Interior region highlighted in green (Source: (AFD, 2006)).

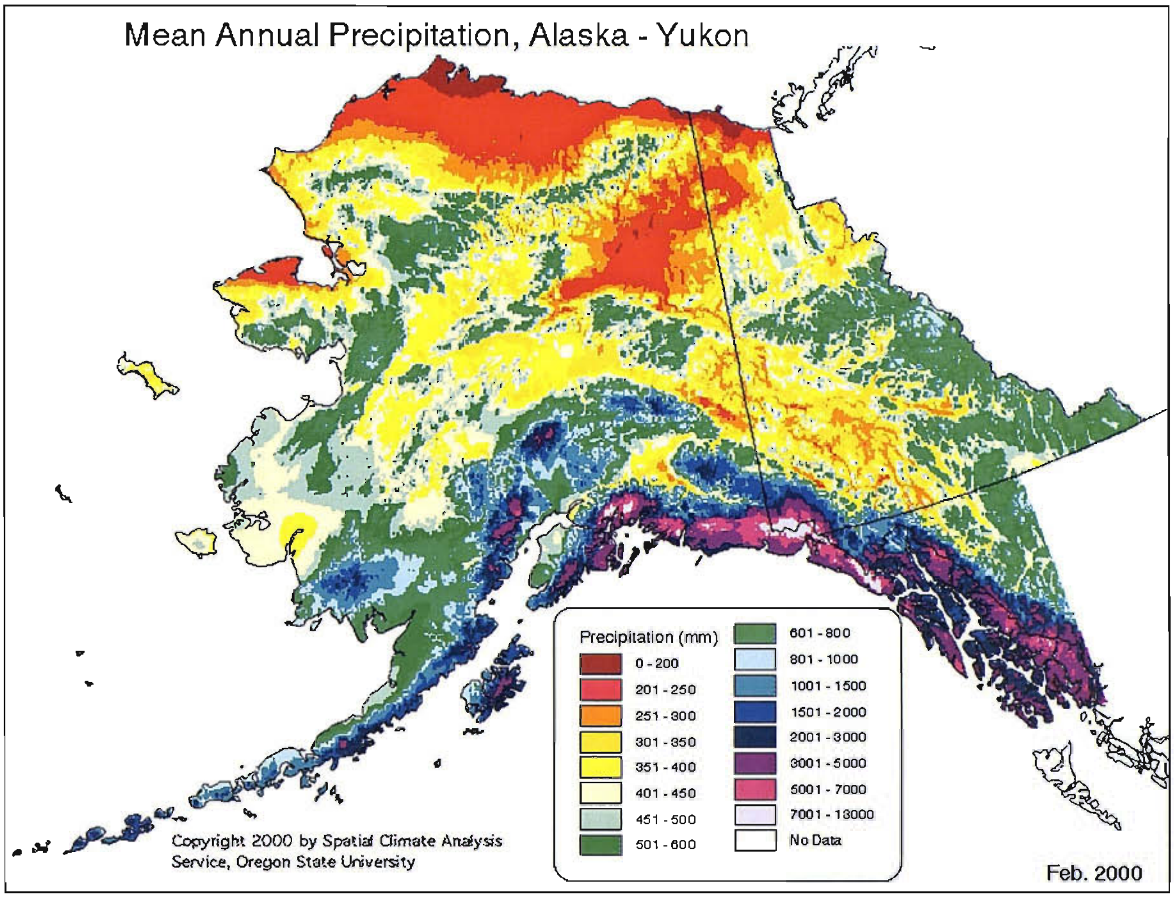


Figure 2.4b: Precipitation map of Alaska and the Yukon Territory, showing the Interior region as semi-arid (<300mm annual average precipitation). (Source: SCAS (2000))

2.3.2 Yukon Territory

The climate of the Yukon is generally continental, the Beaufort Sea is ice-covered for long enough periods to negate moderating effects on air temperatures, and the Pacific Ocean is cut off to the south by the St. Elias Mountains. Average January temperatures range from -20°C in the south to -30°C in the north, with lower values in the interior valleys, average July temperature is ~15°C. (Phillips, 1990). As in Alaska, during the winter, there is a shallow, intense temperature inversion present almost continuously from late October, to early March and on occasion in the summer. Precipitation throughout the Yukon varies considerably from north to south and from valley bottom to mountain peak. The southwest corner, exposed to storms from the Pacific, receives ~2000mm annually. The southeast of the Territory, where the study sites are located receives ~250mm annually (Figure 2.4b). Again as in Alaska, Territory-wide, thunderstorms are more numerous than expected at high latitudes with the storm season lasting from late May to early September.

2.4 Modern Vegetation

North and central Alaska and the Yukon Territory are dominated by boreal forest and tundra (Viereck and Little, 1975; Viereck et al 1986). The vegetation composition is strongly affected by topography, successional stage, and forest fire history (Viereck, 1970, 1983; Viereck *et al.*, 1992; Hulten, 2000).

The boreal forests of Alaska are characterised by three conifer (Yukon has four) and three hardwood tree species that have markedly different life-history characteristics, soil requirements and responses to disturbance (Anderson & Brubaker, 1994). The interaction of species with variations in topography, soils and disturbance results in distinct forest mosaics (e.g. Viereck *et al.*, 1986; Bonan & Shugart, 1989; Van Cleve *et al.*, 1991).

Picea mariana (black spruce, Figure 2.5a and b) primarily grows in pure stands on poorly drained clays and clay loams on north-facing upland sites and on thick organic soils in low-lying muskegs (Viereck et al, 1986). It has the lowest nutrient requirements of all the boreal forest species (Chapin, 1986; Van Cleve & Yarie, 1986). Black spruce is well adapted to endure disturbance regimes through a flexible reproductive strategy, establishing either by vegetative layering or sexually reproducing, and semi-serotinous (requires increases in heat/fire to open and release seeds) cones which provide an on-site seed source for reestablishment as long as 8 years post fire (Viereck, 1983). This allows it to occupy areas with a variety of site conditions, including recent burns (Black & Bliss, 1978), thick organic soils (Viereck et al, 1986) and deep snow patches(Payette *et al.*, 1985).

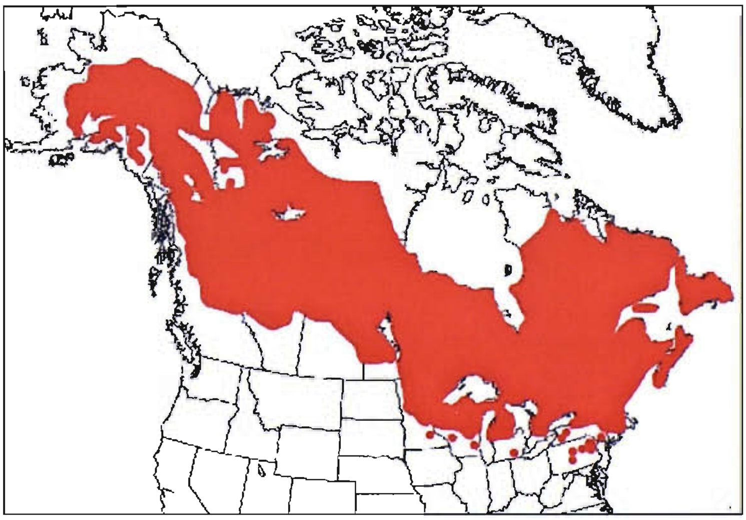


Figure 2.5a: Modern geographic range of P. mariana (Source: http:www.nearctica.com/trees/p mariana.jpg)

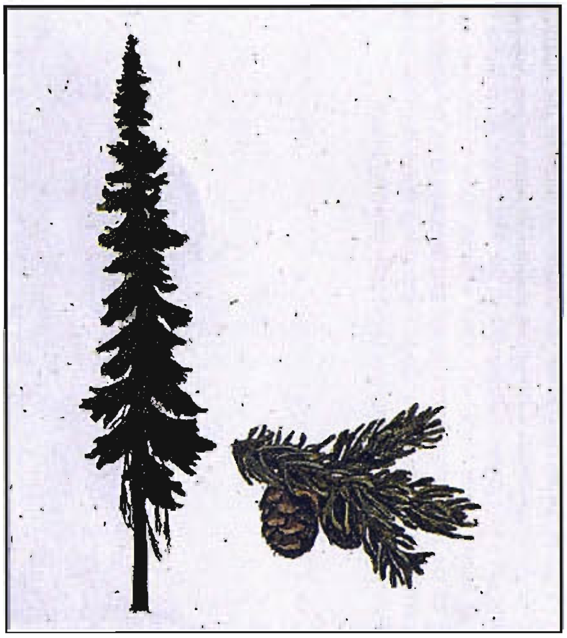


Figure 2.5b: Picture of black spruce to show needle and cone orientation, and tree morphology. (Source: Forest Management Division, Environment and Natural Resources (Resources, 2004)).

P. glauca (white spruce, Figure 2.6a and b) generally grows either as single-species stands or in mixtures with *B. papyrifera,* on warm, well drained-sites (Anderson and Brubaker, 1994). It is often restricted to south-facing slopes and alluvial soils (Viereck *et al.*, 1986; Ritchie, 1987) and is less tolerant of cold poorly drained soils than black spruce. White spruce attains the largest sizes and is most abundant along elevated levees and old meander banks of major rivers, especially in the Yukon flats and along the Porcupine River (Yarie & Van Cleve, 1983). It is not fire adapted but can re-colonise burnt sites from nearby populations (Van Cleve et al, 1991), although seed establishment and germination is best on mineral soils exposed after fires or floods (Zasada, 1986). All stages of the life history favour warm summer temperatures (Garfinkel and Brubaker, 1980).

Figure 2.6a: Modern geographic range of P. glauca (Source http://www.nearctica.com/trees/conifer/pglauca.jpg)

Figure 2.6b: Picture of white spruce to show needle and cone orientation, and tree morphology. (Source: Forest Management Division, Environment and Natural Resources, (Resources, 2004)).

Larix laricina (Alaskan Larch, fig 2.7) is the only deciduous conifer in the region, and common only in south-central Alaska, where it occurs on poorly drained sites with black spruce (Viereck and Little, 1975). It grows best on well-drained loamy soils alongside streams and lakes, and on mineral soils with a shallow organic overlay. However, it is uncommon on these sites in the southern portion of its range due to shade intolerance and competition from other trees. Numbers are also kept low due to its inability to tolerate fire. Mature trees are killed easily due to thin bark and shallow roots. Seeds have no endosperm to protect them from high temperatures, and so are usually destroyed by fire. Post-fire regeneration is therefore dependent on surviving trees as dispersal of seed is over short distances and thus not adapted to rapid reseeding of large burnt areas. Therefore larch usually is often found as isolated pockets of surviving trees in boggy/swampy terrain.

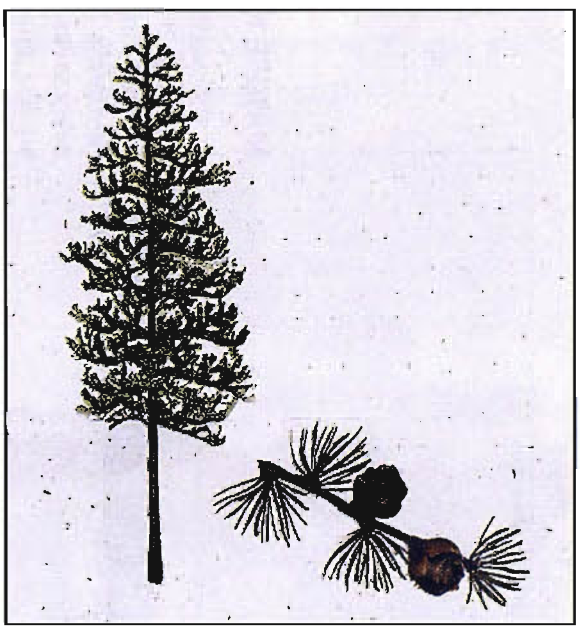


Figure 2.7: Picture of Alaskan Larch to show needle and cone orientation, and tree morphology. (Source: Forest Management Division, Environment and Natural Resources, (Resources, 2004)).

B. papyrifera (paper birch, fig 2.8) grows in pure stands, or in mixtures with white spruce on light textured soils, typically on south-facing slopes and alluvial sites (Viereck et al, 1986). In eastern upland areas, mosaics of paper birch and white spruce are common (Anderson and Brubaker, 1994). Paper birch like most hardwood species requires warmer soils and higher nutrient regimes than spruce, but regenerates vigorously after fire by stump sprouts in addition to dispersal of abundant winged seeds (Viereck and Little, 1975). It is commonly found with *Alnus crispa* (mountain alder) and willow spp. on riparian sites and active floodplains throughout the boreal forest (Viereck et al, 1986).

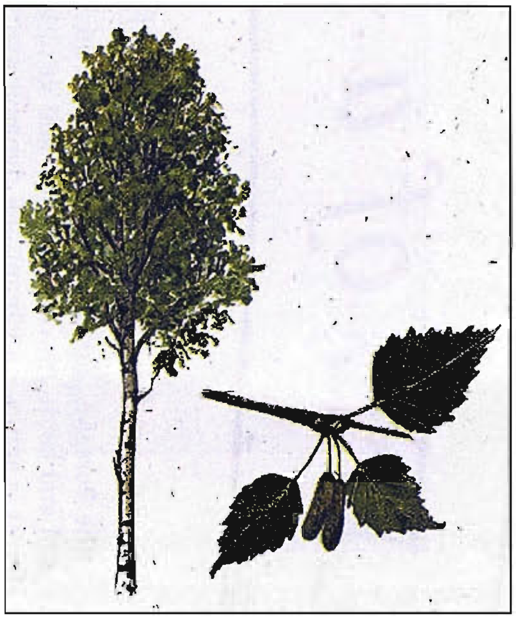


Figure 2.8: Picture of paper birch to show leaf and bud orientation, and tree morphology. (Source: Forest Management Division, Environment and Natural Resources, (Resources, 2004)).

Populus tremuloides (trembling aspen, fig 2.9) is most common on warm dry upland sites (Viereck and Little, 1975). It is shade intolerant and regenerates by wind-dispersed seeds and vigorous root suckers (Viereck and Little, 1975). Trembling aspen regenerates abundantly after fire and without disturbance, is typically replaced by white spruce (Viereck et al. 1986).

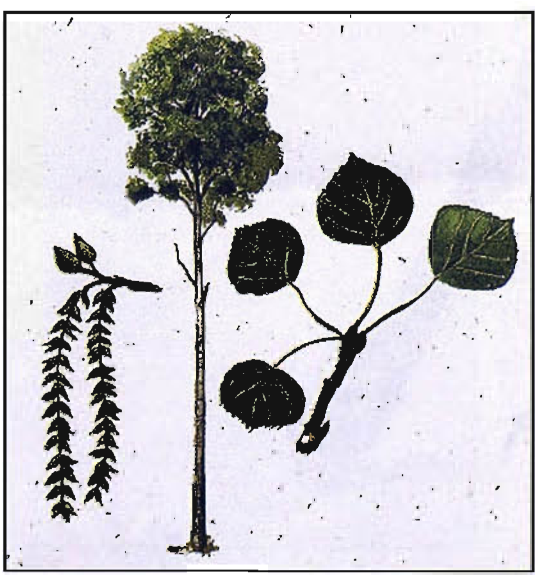


Figure 2.9: Picture of trembling aspen to show leaf and bud orientation, and tree morphology. (Source: Forest Management Division, Environment and Natural Resources, (Resources, 2004)).

Populus balsamifera (balsam poplar, fig 2.10) is common in river valleys including sandy bottoms and gravely flood plains, terraces and coarse alluvial fans (Viereck and Little, 1972). In forests, especially openings and clearings, it is associated with white spruce, birch and aspen. Rare hybrids with quaking aspen have been recorded (Viereck and Little, 1972). Balsam poplar grows rapidly to approx. 80-100ft before being successionally replaced by white spruce.

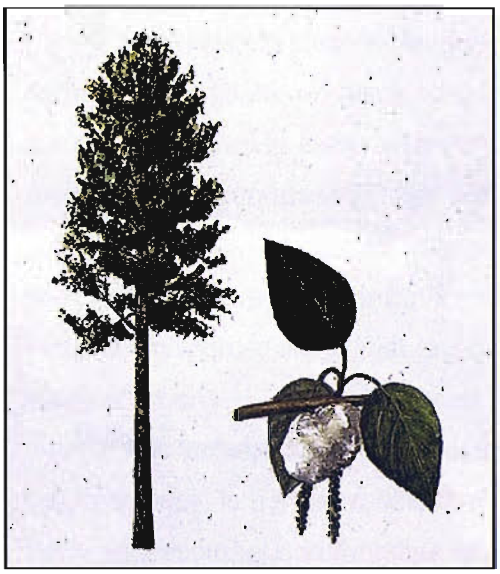


Figure 2.10: Picture of balsam poplar to show leaf and bud orientation, and tree morphology. (Source: Forest Management Division, Environment and Natural Resources, (Resources, 2004)).

Pinus contorta var. latifolia (lodgepole pine, fig 2.11) is currently only found in the Yukon Territory, although it is predicted to spread into Alaska in response to global warming. It grows both in extensive pure stands, and in association with the typical boreal conifers. Lodgepole pine is a minor seral species in moist habitats and a dominant seral species in warm, well-drained habitats, although it can persist even on cool sites. The cones require extreme heat (e.g. from fire) to open and release the seeds (i.e. are fully serotinous), and cones may remain on the tree for many years until a fire occurs.

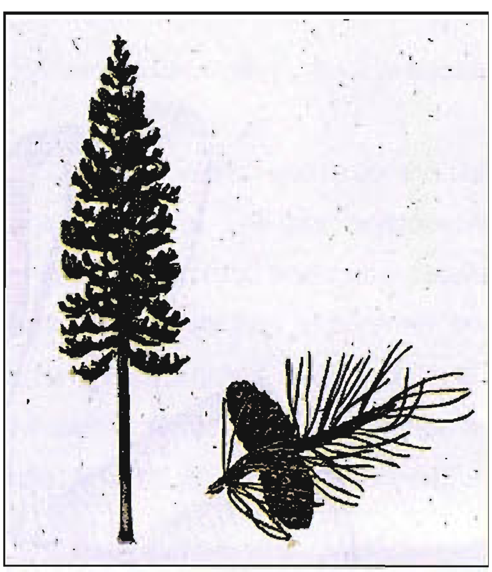


Figure 2.11: Picture of lodgepole pine to show needle and cone orientation, and tree morphology. (Source: Forest Management Division, Environment and Natural Resources, (Resources, 2004)).

Hillsides are vegetated with mixed stands of aspen, birch, white spruce and black spruce. Microsite temperature and moisture variations, in addition to successional history strongly affect species distributions (Viereck, 1970). South facing slopes are dominated by aspen, birch and white spruce. At these sites fire history plays an important role, and localities which are burned repeatedly may remain aspen and birch dominated (Quirk and Sykes, 1971), while areas which burn more rarely eventually succeed to a white spruce forest.

North-facing slopes, because of inadequate drainage and shallow permafrost, are dominated by scattered black spruce, larch, shrub birch and willow (Bigelow, 1997). Trees are rare or absent in areas with very poor drainage, and the vegetation is dominated by *Carex* spp., cotton grass (*Eriophorum* spp.), labrador tea (*Ledum groenlandicum*), and Cassandra (*Chaemadaphne calyculata*). Steep south-facing slopes are droughty sites and are often either treeless, or have scattered aspen, with vegetation dominated by grasses, sedges, and forbs, along with low-lying shrubs (e.g. *Artemisia frigida*) and *Juniperus* in some locations.

The vegetation on the alluvial plains, depending on successional stage and forest fire frequency, is a mosaic of willow, balsam poplar, white and black spruce, and larch. Early successional communities are dominated by willow, alder and balsam poplar. As the balsam poplar dies, leaving openings in the canopy, white spruce begins to gain dominance. On older alluvial deposits, the white spruce develops into a black spruce forest and eventually to bogs (Viereck, 1970). This final transition is thought to be controlled by a thickening of the organic mat, resulting in cooler soil temperatures, higher permafrost table, and impeded drainage, which promotes the development of black spruce bog. This successional sequence is strongly affected, however, by repeated burning, as well as flooding, so that areas with a high flood frequency may remain a white spruce forest indefinitely (Mann et al., 1995).

2.5 Late Quaternary Climate History

This section will cover the climatic reconstructions for East Beringia using models and proxy data. The late Pleistocene and Holocene climate history of the Yukon is only broadly understood and many details remain unknown (Anderson, 2004). The climatic history of Alaska has, in contrast, been more widely researched. Here a regional climate history is presented, with more local details for the two areas discussed where possible. All dates mentioned in this section are calibrated years BP, if original source cited uncalibrated ages, these were calibrated using IntCal04 (Reimer *et al.*, 2004). Climate model simulations, based on glacier dynamics and palaeoecological- and geomorphological data are used to describe significant changes since the last glaciation.

Climate simulations

Climate forcing mechanisms during the glacial-interglacial transition include solar insolation, ice-sheet decay and feed-backs resulting from rising sea-level (COHMAP, 1988). Barnosky et al (1987) reviewed the major climate changes for Alaska and the Pacific Northwest as predicted by GCM simulations for deglaciation and showed that the general climatic trends predicted are compatible with the fossil pollen data. Simulations for 21 ka BP produced warmer than present January temperatures, (Schweger, 1997), and colder than present July temperatures in East Beringia,

Bartlein *et. al* (1998) looked specifically at model outputs for Alaska, which defined probable critical thresholds of change in climate systems related to summer insolation values and the extent of the Laurentide Ice Sheet. At 18 ka BP the Laurentide Ice sheet was extensive and had a strong effect on circulation. Anticyclonic circulation over the ice sheet in central Canada displaced most of the eastward flowing jet stream to the south, which resulted in cold and dry winter conditions in unglaciated Alaska (Bartlein et al, 1998).This resulted in increased southerly flow into Alaska and a strengthened Aleutian Low. Summer and winter temperatures were significantly colder at the ice sheet edge, but simulations suggest that much of interior Alaska may not have been significantly cooler than today (Bartlein *et al.*, 1998). The climate would have been more severe in the east, especially the Yukon as it is closer to the Laurentide and Cordilleran glaciers and within the rain shadows of the Alaska-St. Elias ranges.

Moisture simulations

Simulations for 21 ka BP discussed in Barnosky et al, (1997) indicate that July precipitation was less than present and eastern areas were drier than the west. Subsequent studies simulated that winter and summer effective moisture was significantly less than present, possibly half of modern values (COHMAP, 1988; Kutzbach *et al.*, 1993; Kaplan *et al.*, 2003).

The presence of large ice sheets would have significant effects on albedo, airflow, and energy exchange over the ice sheets, and also regionally, with GCM simulations indicating very low air temperatures occurring downstream (eastward) of the Laurentide ice sheet (up to 32°C colder in winter, and 8°C lower in summer south of Greenland) (Manabe & Broccoli, 1985). By ~14 ka BP this Ice Sheet effect had disappeared, and circulation patterns would have been more similar to present, with climate model simulations indicating a weakened Laurentide anticyclone, with a diverted/split jet stream only in summer. This reduced the southerly flow into Alaska and increased the westerly flow from Asia. However, ice-sheet cooling effects may have remained in parts of Alaska until 12 ka BP (Bartlein *et al.*, 1998). Lakes in interior Alaska were either seasonally dry or dessicated prior to ~13 ka BP (Abbott et al, 2000). Lake-level reconstructions at ~13 ka BP indicate that precipitation was 35-75% less than modern values (Barber and Finney, 2000). Pollen records indicate that vegetation was dominated by herb tundra, which implies much lower rates of evapotranspiration than present.

From 15 to 8 ka BP, orbital variations resulted in increased seasonality at 65°N, reaching maximum values at ~10 ka BP (Edwards & Barker, 1994). By this time the Laurentide ice sheet had considerably receded, and only had minor temperature or circulation effects on central Alaska (Bartlein et al, 1991). According to Milankovitch theory, maximum summer solar radiation at 60°N occurred between 10 and 9 ka BP. At this point, summer insolation was approximately 9% greater than present, and winter insolation was approximately 25% less than present (Berger, 1978, Bartlein et al, 1991). At 9 ka BP simulated summers were warmer than present due to high summer insolation and a strongly developed Pacific High. Lake-level reconstructions suggest that precipitation was 25-45% less than modern (Barber and Finney, 2000). Evaporation was possibly higher than modern due to greater summer insolation, but this alone cannot account for lower lake levels (Barber and Finney, 2000).

Climate model simulations at 9 ka BP suggest that summer temperatures were significantly higher than at present, however, winter temperatures were slightly cooler (Kaufman *et al.*, 2004).

By 7 ka BP, summer insolation at 65°N was still greater than modern, resulting in significantly warmer simulated temperatures (Berger, 1978; Kutzbach et al, 1993). Lake-level reconstructions estimate precipitation for this period as 10-20% less than modern, with lake levels recorded as intermediate to high (Barber and Finney, 2000). After 7 ka BP, orbital effects relaxed (Bartlein et al, 1991), and oxygen isotope studies from lake sediments in northern Alaska indicate a transition to cooler, more mesic conditions after ~5.5 ka (Anderson et al, 2001). The middle Holocene saw development of peatlands and permafrost (Cwynar and Spear, 1995; Vardy et al, 1997), which points to a cooler and wetter climate.

From 5 ka BP the climatic conditions converged on those of present (Bartlein et al, 1998). Some palaeoclimatic records from northwestern British Columbia, southwestern Yukon, and Alaska suggesting that Late Holocene (ca. 3.0 – 2.0 ka BP) climate may have been influenced by specific air mass circulation dynamics (Spooner *et al.*, 2003; Anderson *et al.*, 2005; Anderson *et al.*, In Press). This is supported by evidence of increases in far-travelled (exotic) western hemlock (*Tsuga heterophylla*) pollen, which have been attributed to long-term changes in air mass circulation (Cwynar 1993; Spooner et al, 1997; Spooner et al, 2002).

Lake-sediment records of diatom inferred salinity and biogenic silica abundance have been used to identify high frequency climate variability (Hu et al, 2003). Presently there is no widely accepted proxy that characterises and describes late-Holocene climate variability in the northwest arctic and sub-arctic, although some recent studies are using isotopic techniques (Darby *et al.*, 2001; Anderson, 2004; Anderson *et al.*, 2004), and tree rings (Briffa, 2000; Rayback & Henry, 2005, 2006).

2.6 Glacial History

Variations in the controls of regional climate, particularly insolation, for East Beringia have been reconstructed for the late and middle Pleistocene, a period of ~800 ka, and compared to the marine isotope record (Bartlein et al, 1991). This reconstruction shows that the climate has been in a constant state of flux, cooler than the Holocene for nearly all of the past 800 ka and for five time periods similar to full-glacial conditions during the period from 20-14ka B.P. (oxygen isotope stage 2). For the Yukon, glaciers formed and advanced during some of these cold periods. During the Late Wisconsinan McConnell Glaciation (~20 ka BP), the Whitehorse area was glaciated by ice lobes originating in the Coast Mountains of southern Yukon (Bond, 2004). Glacial maximum, ice retreat and the re-establishment of vegetation in the Whitehorse area was complete by 10.7 ka BP (Anderson *et al.*, 2002) according to terrestrial and aquatic macrofossils from sediments in Marcella Lake (Kettlehole Pond (Cwynar, 1988)) and Dragonfly Lake (this study) in the southern Yukon. However, the ice-free corridor between the Cordilleran and Laurentide ice sheets may have been open by ~14ka BP (MacDonald & Mcleod, 1996).

During the last glacial maximum (~21 ka BP), central Alaska was predominantly unglaciated as the region was too arid for glacial growth. Glaciers were limited to the mountain valleys of the Brooks Range and Alaska Range, and did not extend beyond the foothills (Hamilton, 1986, 1986a; Péwé, 1975; Ten Brink, 1983). By 19 ka BP, ice was retreating in all mountain systems and reaching near-Holocene volumes by about 17 ka BP. Subsequent brief glacial advances in the Brooks Range and Alaska Range date to approximately 15.5 to 16.5 ka BP (Hamilton, 1986). Small advances occurred in the Alaska Range about 14.5 ka BP and between 12.4 and 11ka BP (Ten Brink and Waythomas, 1985).

Although the earliest dates of cirque glacier expansion in response to minor climatic fluctuations following early Holocene retreat in the Brooks Range and Alaska Range are vague, it is thought to have been between 5.8 and 5 ka BP (Hamilton, 1986a; Calkin, 1988). Ages on valley glacier expansion on the northern flank of the St. Elias Mountains in southern Alaska and Yukon Territory are younger, ~3 ka BP (Denton and Karlen, 1977). Since then there have been glacier and tree-line fluctuations culminating in so-called "Little Ice Age" advances between ~AD 1500 and the early 20th Century. Tree-ring studies from the Yukon do not extend beyond 1000 years and do not yield distinct climate signals (Szeicz and MacDonald, 1995; Szeicz and MacDonald, 1996). Barber et al, (2000) show that tree-ring width in central Alaska is controlled by temperature induced drought-stress, illustrating again, the importance of considering moisture availability for this region.

2.7 East Beringian Vegetation History

Last Glacial Maximum

Full glacial pollen assemblages from eastern Beringia are dominated by *Artemisia,* Poaceae, Cyperaceae and other herbaceous types (Lamb and Edwards, 1988; Anderson and Brubaker, 1994). Important topographic and longitudinal moisture gradients, and a mosaic of vegetation types in an arid environment have been identified from pollen influx, plant macrofossil and fossil insect remains (Schweger, 1997). Mesic Graminoid tundra occupied lower elevations of western areas and more xeric, sparse tundra communities occupied the east and higher elevations (Anderson and Brubaker, 1994). The exact nature of this vegetation remains controversial, however, because there is no close modern analogue for vegetation with high abundances of these taxa, especially *Artemisia* (Lamb and Edwards, 1988; Bigelow, 1997). One theory (e.g. Guthrie, 1990) suggests this grassland as highly productive, with similarities to temperate steppe. Abundant faunal remains from large herbivores (mainly grazers e.g. *Mammuthus* (mammoths), *Equus* (horse), *Bison* (bison) and *Rangifer* (reindeer), Guthrie, 1968, 1982) during this period indicate a productive landscape must have existed to support them. Another theory (e.g. Ritchie, 1982; Cwynar, 1982) suggests this landscape was barren and unproductive Arctic tundra, as comparisons between fossil pollen from discontinuous tundra on xeric substrates in the northern Yukon (Cwynar, 1982) and pollen from the high Arctic islands of Canada (Anderson et al, 1989) produce similar spectra to those of full-glacial vegetation.. The application of a plant functional group approach (Bigelow et al., 2003) recently assigned pollen taxa from the last glacial maximum (LGM) to one or more plant functional types, which were then assigned to biomes using a rule-based algorithm (Prentice *et al.*, 1996). Most Alaskan spectra were classified as either prostrate shrub tundra or grass-forb tundra.

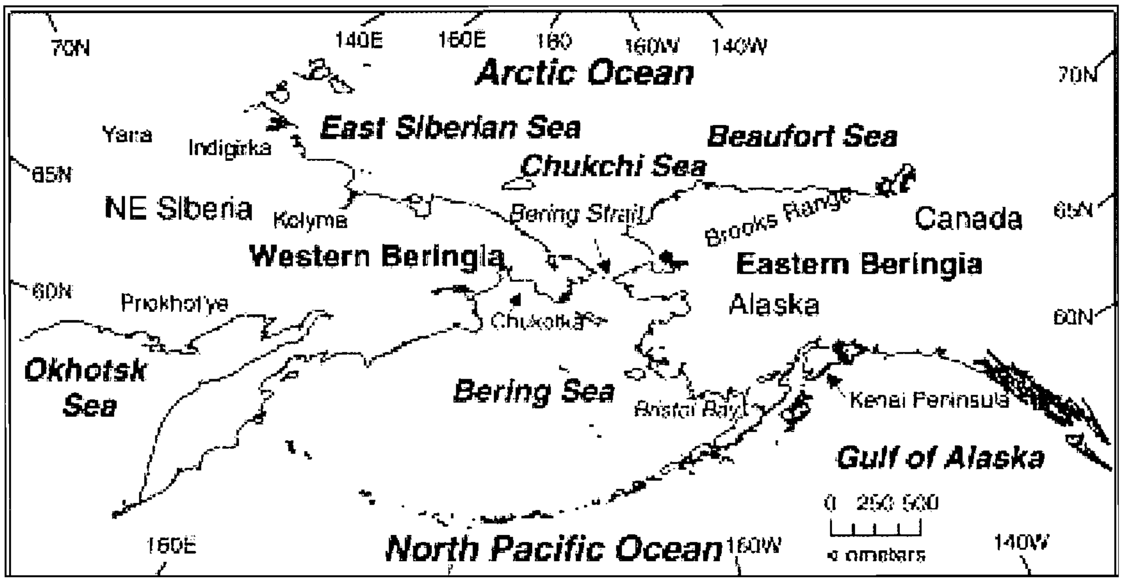


Figure 2.12: Location map of Beringia. Eastern Beringia consists of Alaska and unglaciated Yukon Territory . (Source: Brubaker et. al, 2005, with permission)

The Late-glacial climatic amelioration ca 14,000-15,000 cal yr BP

The increase of fossil *Betula* (shrub) pollen in pollen records marks the end of full-glacial conditions (Schweger, 1997). It is the most widespread and significant event in the Arctic prior to the Holocene. The increase of *Betula* is registered at *ca.* 15 ka BP in the ice free areas of Beringia (Lamb and Edwards, 1988). This was presumably triggered by the removal of the ice sheet effect (see above). The importance of moisture changes is particularly clear in east-central Alaska, where the regional increase in shrub *Betula* correlates with times of rapid lake-level rise (Bigelow & Edwards, 2001; Edwards *et al.*, 2001). It is probable that this shift is related in some way to the submergence of the Bering land bridge (Bartlein *et al.*, 1998), which probably brought milder conditions to the coast when inland areas were possibly still dominated by the continental ice-sheet (Lamb and Edwards, 1988).

The exact timing is not easy to pinpoint (e.g. Bigelow, 1997), and pollen influx studies suggest that while *Betula* (presumed shrub species *B. nana* or *B. glandulosa*) became an important component of the vegetation, most herbaceous taxa from the previous period remained abundant, indicating that the new vegetation was most likely a mosaic, with *Betula* occupying the mesic sites and herbaceous taxa occupying the more arid and exposed sites (Ritchie, 1982, Anderson and Brubaker, 1994).

Vegetation response to the early-Holocene thermal maximum ca 11,000 cal yrs BP

A vegetation response in eastern Beringia to exceptionally warm summers associated with the insolation-driven thermal maximum at high latitudes in Alaska and western Canada (Kaufman et al, 2004; Ritchie et al., 1983) can be seen in a combination of vegetation changes (Edwards *et al.*, 2005). Various taxa extended north of their present limits in ice-free areas of the Arctic during the early Holocene including the tree species - *Picea, Larix, Betula* and *Pinus,* (Lamb and Edwards, 1988). Between 13 and 10 ka BP, shrub tundra continued to dominate most upland sites, however, *Populus* forest or woodlands were common in central Alaska, although the interpretations of forest extent vary from relatively restricted populations along slopes or floodplains to rather extensive, dense forests covering lower elevations (e.g. Ager, 1983; Brubaker *et al.*, 1983; Hu *et al.*, 1993; Anderson *et al.*, 1994). This increased abundance of *Populus* is consistent with warmer than present summer temperatures in central and northern Alaska (e.g. Brubaker *et al.*, 1983; Anderson & Brubaker, 1994; Edwards & Barker, 1994), although lake-level evidence suggests that the climate remained drier than modern (Edwards *et al.*, 2001).

Vegetation changes from ca 10,000 ka BP

Picea glauca (white spruce) populations expanded between 10 and 9 ka BP, possibly from source areas in northwestern Canada (Ager, 1984; Anderson and Brubaker, 1994), or small local refugia (Brubaker *et al.*, 2005, see below). This was followed by the expansion of *Alnus* (alder) shrubs first in south-western Brooks Range between 10 and 9 ka BP, then throughout the entire region between 8 and 7 ka BP (Anderson and Brubaker, 1994). In western Alaska, *Alnus* preceded spruce (Ager, 1984). *P. mariana* (black spruce) became a dominant tree species *ca.* 6 ka BP (Anderson and Brubaker, 1994). The modern community distribution was achieved in central Alaska between 6 and 4 ka BP (Anderson and Brubaker, 1994).

In the Yukon Territory black spruce expanded regionally, coincident with an expansion of alder populations at ~6 ka BP, thought to be linked to mid-Holocene increased moisture (Cwynar & Spear, 1995). *Pinus* is thought to have arrived in the southeast Yukon by ~3 ka BP as part of a migration north-west from British Columbia, reaching its current northern limit as recently as ~150 years ago.

The transitions between deciduous and spruce vegetation in Alaska and the Yukon Territory, and the further shift to pine dominance in the Yukon is discussed in more detail below, distinguished by region.

The early-Holocene expansion of spruce in eastern Beringia

The appearance of coniferous- forest environments is evidence of the general trend towards warmer climates (Anderson and Brubaker, 1994), and increasing effective moisture (Szeicz & MacDonald, 1996; Mann *et al.*, 2002; Kaplan *et al.*, 2003). This transition will be used to investigate the hierarchy of controls on the fire regime outlined in H1, Chapter 1.

Hopkins *et al* (1981), investigating the possibility of a refugium for trees in Beringia during the glacial period, concluded that there is no evidence that any woody taxa survived *in-situ*, but suggested an east-west migrational pattern exists for *Picea*, with the oldest dates occurring in the east. It was suggested that *Picea* migrated into the region from locations south of the ice-sheet, via the Canadian 'ice-free corridor' (Rutter, 1980), reaching the Mackenzie delta region by ~9ka BP (Ritchie & MacDonald, 1986), and quickly spreading into central Alaska and the south-central Brooks Range (Lamb & Edwards, 1988). The rapid migration was possibly facilitated by strong south-easterly winds related to the Laurentide ice-sheet, transporting seeds down the Mackenzie river itself, in addition to the optimal establishment conditions in an open landscape (Ritchie and MacDonald, 1986). This is supported additionally by it's diachronous appearance across Alaska with the earliest sites in the east. According to palynological data, *Picea glauca* appears to be the first spruce species to appear in the region (Lamb and Edwards, 1988), probably colonising well-drained sites where *Betula* may not have been abundant. Pisaric et al (2003) found that pollen morphology indicated that *P. mariana* accompanied the arrival of *P. glauca* at some sites in northern British Columbia, although this is not found regionally.

More recently, Brubaker *et al* (2005) readdressed the refugium theory by assembling 154 pollen records for the period 21 to 6 ka BP. They found that in eastern Beringia, *Picea* pollen is present in low percentages (<2%) at most sites, and by 12 ka BP the percentages are >5% at three sites in far northwestern Canada and >30% at one site in coastal southeastern Alaska. Pollen increases across central Alaska started at ~11 ka BP, and reached levels >5% at ~50% of the sites by ~9 ka BP (Anderson & Brubaker, 1994). There is also some genetic data supporting the possible survival in refugia of *Picea glauca* in Alaska (Anderson *et al.*, 2006). Brubaker *et al*, (2005) propose that the data argue more strongly for survival within Beringia than for immigration from outside regions due to the new dating compilation.

Many pollen records in central Alaska indicate a mid-Holocene decrease in *P. glauca* pollen prior to the arrival of *P. mariana* (Brubaker et al, 1983; Anderson et al, 1990; Hu et al, 1993; Anderson and Brubaker, 1994; Bigelow, 1997) between 8 and 6 ka BP, probably related to an episode of warm and dry climate in Interior Alaska (hypsithermal) which was unfavourable to *Picea* (Ager, 1983). Floodplain forests in this region probably remained relatively unchanged, but *P. glauca* woodlands on fine-textured lowland soils may have declined (Anderson and Brubaker, 1994). In eastern Alaskan and Yukon sites there is no apparent decline (Cwynar and Spear, 1995; Anderson et al, 1994b; Keenan and Cwynar, 1992; Cwynar and Spear, 1991; Anderson et al, 1988; Cwynar, 1988; Ritchie, 1982). Carlson and Finney (2004) record a small decrease in *Picea* pollen percentages at a site near Fairbanks. This decline in *P. glauca* and subsequent rise in *P. mariana* was speculated to be a successional trajectory induced by autogenic edaphic changes (Brubaker et al, 1983; Anderson et al, 1990). However, Hu et al (1993) (supported by Cwynar and Spear,1995; Carlson and Finney, 2004) propose an alternative explanation: that the oscillation in *P. glauca* pollen results from climatic rather than successional forcing. Because the decline and subsequent recovery of *P. glauca* occurs prior to the appearance of *P. mariana*, a progressive development of unsuitable soil conditions cannot be responsible for both events. Therefore climatic cooling is much more probable. This hypothesis is still in need of an independent supporting climatic record.

A second major change in forest composition occurred *ca.* 6 ka BP when *P. mariana* replaced *P. glauca* as the dominant forest species in central and eastern areas. It is probable that *P. mariana* entered *P. glauca* woodlands on non-riparian sites in central and eastern areas. The *Picea* treeline was near its present location by *ca.* 4 ka BP, in the foothills of the Brooks range (Viereck, 2000).

Holocene vegetation development in the southern Yukon

In the southern Yukon, alder (*Alnus)* and birch (*Betula)* increased in abundance simultaneously, prior to the appearance of (most likely white) spruce (*P. glauca*) (Cwynar, 1988; Schweger, 1997). Black spruce (*P. Mariana*) and alder expanded over southern and central Yukon during the mid Holocene, probably in response to a cooler, more mesic climate, and the development of organic terrain and ice-rich permafrost (Cwynar and Spear, 1995).

Lodgepole pine is a key species in the boreal forest of Canada, although it was the latest tree species to migrate northward into the region to its current range (Wheeler & Guries, 1982; MacDonald & Cwynar, 1991). Its modern range was glaciated until ~10 ka BP (MacDonald & Cwynar, 1985; Anderson, 2004). Lodgepole pine is traditionally thought to have appeared first in southern Yukon records ~3 ka BP (MacDonald and Cwynar, 1985), and reached its present northern limit at Gravel Lake approximately 120 years ago (Cwynar and MacDonald, 1987; Schweger et al, 1987). It has been proposed that lodgepole pine is still migrating northwards, due to the lack of climatic limitation at its present northern limit, and its ability to grow healthily further north and west of its current range (Johnstone & Chapin, 2003).

There is considerable debate concerning possible refugia for lodgepole pine during the Wisconsin glaciation. Some consider it possible for lodgepole pine to have survived in unglaciated areas of northern Yukon, for example the upper Yukon River valley, as well as south of the ice sheets (Yeatman, 1967; Terasmae, 1973; Wheeler & Guries, 1982; Brubaker *et al.*, 2005). Wheeler and Guries (1982) found genetic evidence indicating two populations could have evolved in separate refugia, and they propose the populations only recently rejoined at the border of British Columbia and the Yukon.

However, others propose that lack of fossil pollen evidence and a likely unfavourable climate both argue against the existence of lodgepole pine in the Yukon through the Wisconsin glaciation, (MacDonald & Cwynar, 1985). MacDonald and Cwynar (1985) reconstruct a northward migration, based on fossil pollen, and find the earliest arrival in the south Yukon and the most recent in the central Yukon. The consensus at present is toward the northward migration theory.

2.8 Charcoal Analysis

Any carbonaceous material (wood, oil, coal etc) once burnt, produces high numbers of charcoal particles. In the right depositional conditions these can be preserved through geological time and used as indicators for previous vegetation types, fire regimes and human activities (Tolonen, 1986). When combined with pollen data, charcoal analysis may contribute to understandings of long- or short-term vegetation dynamics, lake and mire development and other ecological processes (Tolonen, 1986).

Sediment-charcoal studies allow for high-spatial precision in the reconstruction of burned vegetation, with macro-charcoal allowing for the identification of local burning events, and source area (Ohlson & Tryterud, 2000). Often, charcoal fragments are identifiable to the family, genus or species level (Carcaillet, 2001), which allows for the identification of not only the source material, but also when combined with palynology, an estimation of source area. It must be noted that the reconstruction of past fire regimes and vegetation relies heavily on the assumption that charcoal and indeed pollen are time-stratified in soil and sediments (Whitlock and Millspaugh, 1996 and Carcaillet, 2001).

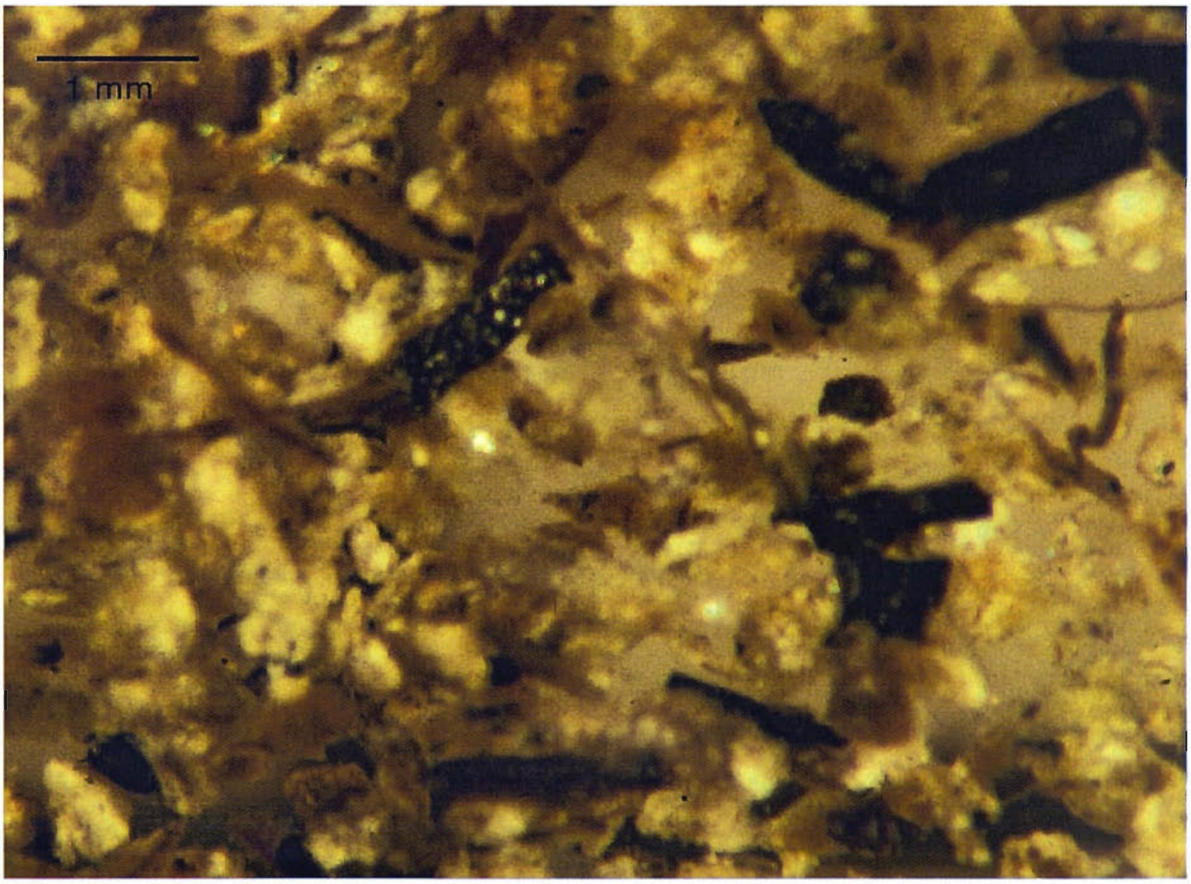


Plate 2.1: Macrocharcoal sample under a 40x magnification. Photograph: LMFS

2.8.1 Charcoal Production

The type of material burnt influences the amount of charcoal produced, with variations depending upon hardness, compactness, and the initial moisture content (Patterson *et al*, 1987). Fire characteristics (spatial extent, intensity and fuel type) impact on charcoal production and deposition (MacDonald *et al*, 1991) in addition to fire duration and temperature (Patterson *et al*, 1987). It is known for example, that particulate emissions are significantly higher in smouldering fires than in the flaming phase (MacDonald *et al*, 1991). Charred particles can vary in size from sub-microscopic to centimetres in diameter, which has implications for transport and deposition within the landscape.

It is evident that the controls over charcoal transport and deposition are complex due to the interaction of factors such as fire regime, vegetation type, mode of transport and taphonomy. These in turn influence the size distribution of particles in the sediment, but such relationships are not well understood at present (Earle *et al*, 1996). Wind and water are the primary agents transporting charcoal from source to depositional site, although there is a possibility of other mechanisms imposing local importance (e.g. human dumping for more recent sediments) (Patterson *et al*, 1987). Therefore, it is necessary to use charcoal analysis carefully, considering the prevailing local physiographic and other conditions, and in conjunction with as many other palaeoecological analyses as possible in order to gain a true picture of the past environment and ecological processes.

2.8.2 Theoretical basis of charcoal analysis

Direct evidence of burning in the environment is the presence of particulate charcoal in the sediment (Scott, 2000). This particulate charcoal is sub-divided into three size categories. Macroscopic charcoal (generally >100 µm, although there is no standardisation within the published literature) is assumed to represent the local charcoal component that results from biomass burning within the watershed. Microscopic charcoal (generally <100 to ~10µm) represents the regional or extra-local fraction, which indicates fire in the surrounding areas, but not necessarily within the watershed. The final category, sub-microscopic, particles <10µm, are thought to represent global or continental sources of fire. Studies generally concentrate on either micro- or macroscopic charcoal depending on the objective of the investigation. Few studies utilise both types, and those that do are generally comparing the methodologies and preparation procedures to assess the accuracy and merits of one versus the other.

Charcoal accumulation rates (CHAR) are a time series of slowly varying background charcoal levels and high prominent peaks that represent the occurrence of fire events via airborne deposition (Clark, 1988; Clark & Royall, 1995; Clark & Hussey, 1996; Hallett *et al.*, 2003). Background levels are made up of regional (microscopic) to global (sub-microscopic) scale charcoal particles that are continuously deposited in the environment. Charcoal abundance has been shown to increase in lakes downwind of a fire event (Gardner and Whitlock, 2001) and therefore CHAR peaks must represent a complex combination of local to extra-local fires in the watershed.

Several assumptions underlie the use of sedimentary charcoal in lakes as an indicator of past fire (see Patterson *et al*, 1987; MacDonald *et al*, 1991). Whitlock and Millspaugh (1996) summarised the most critical of the assumptions as:

1) sedimentary layers with charcoal abundances above a background level are indications of a past fire event,
2) the majority of sedimentary charcoal is from primary fall out and consequently secondary or redeposited charcoal is only a minor component,
3) large particles are only transported a short distance and are thus indicators of local fires,
4) charcoal stratigraphy is broadly similar across a lake, and
5) small lakes provide a better record of local fires as they have a smaller collection area than large lakes.

Whitlock and Millspaugh (1996) tested these assumptions by monitoring macroscopic charcoal particles preserved in lake sediments following a modern fire event. They found that within a few years (~5 years) charcoal abundance in burned and unburned sites differed enough to indicate local burning. Pitkanen et al (1999) suggest that height alone of microscopic charcoal abundance peaks above a background level is not a reliable index of a local fire. MacDonald et al (1991) compared the peak height for various micro-charcoal measurements with ages of fires and found no consistent correspondence between charcoal deposition and the presence of local fires, concluding that the abundance of micro-charcoal is influenced by variations in regional fire regimes.

2.8.3 Charcoal Transportation and Deposition

If we propose to interpret palaeoenvironments by use of charcoal assemblages in sediments, it is imperative that the behaviour of charcoal during transport and deposition is understood (Figure 2.13). Factors affecting integration of charcoal into lake sediments include transportation rates, site of deposition (directly on the lake surface, or to land with subsequent transportation to the lake), the hydrodynamic behaviour of charcoal particles and the mobility of particles within lake systems.

It is unavoidable that there will have been some transport involved prior to deposition of charcoal in lake sediments. Consequently, there is considerable potential for sorting, separation and breakdown of charred material during its transport phase, which will inevitably result in a sample that contains some sort of bias (Nichols *et al*, 2000). While inputs may be quantified by sampling aerially and fluvially transported material, losses will be a function of turnover time for the water in the basin. In closed basins there should be no losses normally for a biologically inert substance such as charcoal (Patterson *et al*, 1987).

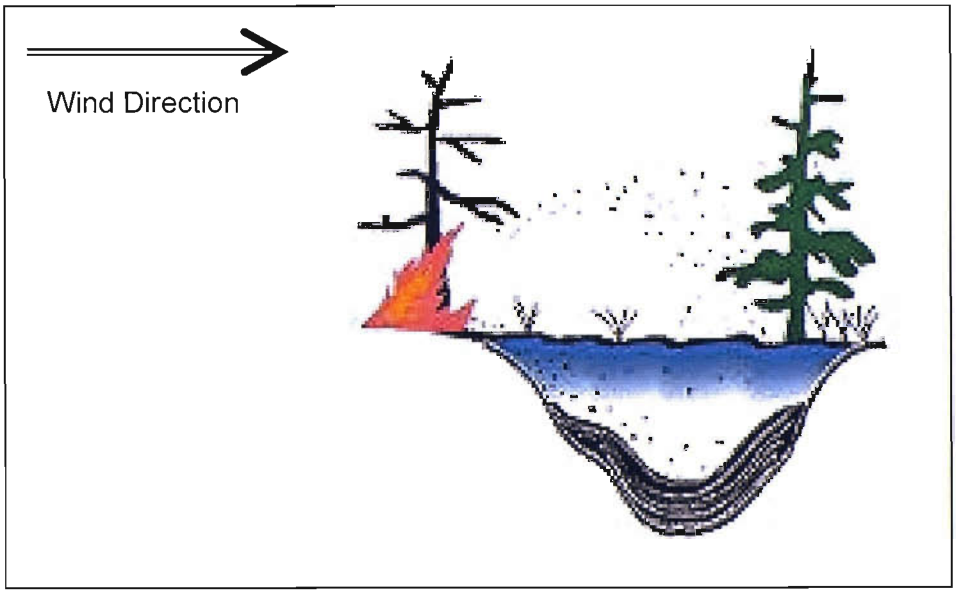


Figure 2.13: Sketch diagram of charcoal deposition within a lake system. Adapted from/www.ncdc.noaa.gov)

Surface runoff can be increased as a result of burning, which leads to increased erosion that can carry charcoal from the source area and contribute to a particle deposition rate that temporally lags behind the actual fire event (Patterson *et al*, 1987). It is possible that sediment values can be elevated above that of airborne deposition by means of surface flow and secondary deposition within lakes, although these are usually found to be of minor importance. This secondary deposition has been documented in lake sediments of eastern North America, where average accumulation rates of charred particles are higher than would be expected from current understanding of fuel loadings, burn efficiencies and emission factors (Clark and Royall, 1994).

An important control on the distribution of charcoal in sediments is the hydrodynamic behaviour of charred material (Nichols *et al*, 2000). Freshly formed charcoal has a bulk density less than that of water, and initially floats. The rate of water-logging may be an important factor in determining where charred material is deposited

(Nichols *et al*, 2000). Water-logging is affected by the state of agitation of the water body, and the temperature of charring. Vaughan and Nichols (1995) found that charcoal formed at temperatures above 300°C became waterlogged more quickly due to fracturing within the tissue structure than particles charred at temperatures below 300°C. They also found a negative relationship between rate of water-logging and charred particle size. Therefore, it is conceivable that the size range of charcoal particles may be a reflection of transport processes (Nichols *et al*, 2000).

Whitlock and Millspaugh (1996) found that particles were still mobile in their lake systems four years after the fire and that abundance in sediments had not stabilised, denoting significant disturbance of particles and subsequent secondary deposition. Tinner *et al* (1998) support this finding by identifying a time lag between the increase in numbers of forest fires (from historical databases) and the charcoal influx in their study area, possibly caused by deposition in the littoral zone of lakes, and subsequent redeposition in the deeper areas. Clark *et al* (1998) found a similar trend, which if it turns out to be a general theme, will support the interpretation that airborne fallout patterns are subsequently altered by surface flow in catchments, or secondary deposition within lakes. It can be assumed that this observable process in modern lakes and charcoal particles is similar to that taking place previously in geological time.

It can be concluded from the published literature that secondary deposition of charcoal is an important process affecting the amount of particles that are incorporated in the sediment record. This does pose interpretive difficulties via smoothing of the fire history, with charcoal peaks from a single fire event spanning a decade or more (Patterson *et al*, 1987; Kangur, 2002).

Aerial Transport and Deposition

Many studies have been conducted to try and understand the way that charred particles are moved in air. Particle movement has been demonstrated by the observation of samples from experimental burns with plume heights in excess of 100m. Larger particles only had an atmospheric residence time of minutes, suggesting the presumption that larger particles are deposited close to source, and thus represent local fire events is correct. However, turbulence may maintain some of the particles in suspension to make extended transport a possibility (Clark *et al*, 1998). Practically no large (>0.5mm) particles were distributed outside burn areas. This experiment illustrates that the transport of macroscopic particles is limited to the burn site, or within short distances of it, probably due to large particle size/weight impeding extended transport in air. The smallest particles typically counted (10µm) can be lofted to low heights from nearby sources and/or lofted to great heights from distant sources (Clark *et al*, 1998). Aerial movement of charcoal is likely to be influenced by the strong convective currents associated with fire (Patterson *et al*, 1987). To date the consensus is that microscopic charcoal particles that are capable of extended aerial transport due to size and weight are diagnostic of regional fire events. The reverse is agreed in consideration of macroscopic particles (or those with a high volume to surface area ratio), which will be deposited more quickly and within a shorter distance from the source (e.g. Patterson *et al*, 1987; Clark and Royall, 1995), and thus represent local events.

MacDonald *et al* (1991) found that microscopic charcoal accumulation rates were greatest during a period when the burning activity occurred between 40 and 120km from the sample site (closest burn occurred 20 km from lake). This suggests that variations in microscopic charcoal are indeed sensitive to changes in regional fire activity. This may not be the case for all lakes, however, as Patterson *et al* (1987) suggest that regional fire activity is unlikely to be represented in microscopic charcoal records from smaller lakes. It is now thought that regional variations will only be identifiable in microscopic charcoal of small lake sediments if the regime is dominated by spatially large fires, with high amounts of annual variation in activity. It is also necessary to ensure the sampling interval is less than the time between peak fire years in order to identify that a peak in the charcoal record signifies a fire event rather than the accumulation of charcoal particles from many fires over several years (MacDonald *et al*, 1991).

Pitkänen *et al* (1999) found that peaks in microscopic charcoal series coincided with fire years based on fire-scar data, suggesting charcoal particles usually counted on pollen slides can originate from local fires, or fires within a few kilometres of the sampling site. This is contrary to the theories of charcoal particle transport (Clark, 1988). Regional sources do contribute to the microscopic charcoal record, but Pitkänen *et al* (1999) suggest that increases in the proportion of largest particles on pollen slides, and a decrease in particle sizes with diameters less than 10 μm indicate local burning. They have linked this result to the fact that fires in their study area burned with low intensity. The convective air column produced by low-intensity burning is low in height, and a proportion of the smoke moves along the surface depositing microscopic charcoal particles in the immediate vicinity of the fire. In the boreal forests of North America high-intensity fires dominate (Johnson, 1992), but the same feature of surface smoke movement can be observed (Pyne, 1984); therefore this feature is not thought to be indicative of fire intensity. In any case, if the concentration of particles is high enough, this smoke can potentially produce a fire record in the sediment.

Sub-microscopic particles should be eliminated from charcoal records that aim to reconstruct local or regional fire activities due to the universal appearance of this category in sediments. Clark (1988) proposes that particles smaller than 5-10μm in diameter can be transported in the air for potentially thousands of kilometres after they have been drawn up into convective columns. Andreae (1983) demonstrated that particles of between 2 and 5 µm in diameter could have continental or even global sources.

As the combustion process depends on the fuel type, fire intensity and weather conditions (Pyne, 1984), the charcoal production may differ in fire regimes dominated by severe or low-intensity fires. Ward and Hardy (1991) found that charcoal particle production decreases with increasing combustion in wildfires, but that particle size increases with fire intensity, due to the rapid movement of particles beyond the combustion zone by turbulent winds.

Gardner and Whitlock (2001) found the most defined macroscopic charcoal peaks occur when the watershed of the lake being sampled burns; less well defined peaks occurred in nearby unburned sites. In addition, charcoal abundance in burned sites was statistically greater than that in unburned sites. They also found that winds during extreme fire events can, and do transport macroscopic charcoal beyond the perimeter of the fire; thus it is possible that wind can blur the distinction between burned and downwind sites by enlarging the area over which macroscopic charcoal is deposited. Whitlock and Millspaugh's (1996) data showed that in all the lakes they sampled a range of charcoal particle sizes was contained within the surface sediments following the modern fire event.

2.8.4 Charcoal Behaviour in Lakes

Charcoal deposition within lakes is another complex area that influences our interpretation of the record, and thus the reconstruction of palaeo-fire. Post-depositional mixing of sediments may affect the spatial distribution of charcoal particles. Mixing and downward migration of charcoal can be a more serious analytical problem than for pollen analysis as charcoal usually indicates events occurring over much shorter timescales (Patterson *et al*, 1987). To identify a specific fire, it would probably be necessary to restrict studies to undisturbed sediments (e.g. annually laminated deposits) or to areas where fires occur infrequently. However, bioturbation is not considered a problem when fire return intervals are long in relation to sedimentation rates (Patterson *et al*, 1987).

In relation to the sample site and its characteristics, Whitlock and Millspaugh (1996) found that the deepest part of the lake is best for sampling. Although a lag in accumulation is probable, the centre displays a steadier pattern of accumulation than the shallower parts of the lake, due to migration of particles from the littoral zone to the deepest parts. The particles deposited close to the shore are exposed to disturbance from water currents that inhibit their integration into the sediment profile. In addition the slope of the lake floor towards the deepest point (usually near the centre of the lake) will also affect the migration and subsequent accumulation of particles. The accumulation rates are therefore more stable in the deepest parts, which will more accurately reflect fire events.

It has also been found that charcoal particles become concentrated more on the downwind shore, due to wind-induced water currents (Whitlock and Millspaugh, 1996). Sampling near to the shore, therefore, will not only lead to inaccurate estimations of charcoal abundance, but if the sampling is on the upwind shore, runs the risk of missing a past fire event all together. The work by Whitlock and Millspaugh (1996) demonstrates the need for calibration of fire around a site in order to confidently describe a charcoal peak as a fire event.

It appears that position of sampling site is more important than the actual lake size, which still remains unclear in its relationship to charcoal recruitment area (Whitlock and Millspaugh, 1996). The assumption that small lakes are better for local fire reconstruction is based on one of the underlying assumptions in pollen analysis techniques. Small lakes will have a smaller surface area in which to intercept charcoal particles carried from long distances, and will more likely be biased towards a charcoal content derived from fires in their direct vicinity. There are few published data either proving or contradicting this assumption, but from comparison with the palynological theory, it seems logical that this trend be upheld in charcoal studies.

2.8.5 Appraisal of Charcoal Methods

Palynologists and palaeoecologists have been using analyses of microscopic charcoal from lacustrine and terrestrial sediment cores for the reconstructions of extended fire histories since Iversen's pioneering work in 1941 (cited in Patterson *et al*, 1987). The number of studies utilizing this technique has increased every decade since, and consequently the methods for sample preparation and charcoal quantification have escalated accordingly. However, no single method of counting and presenting charcoal results has been widely adopted (Earle *et al*, 1996) so a summary of the methods most used is given below.

By far the most widely used technique for microscopic charcoal analysis in published works is quantification of charcoal on pollen slides. The pollen-slide method is used to detect regional trends in fire history (20-100km around study site (Tinner and Hu, 2003) due to the regional long-distance windborne origin of small particles that arise in pollen slides (Carcaillet *et al*, 2001). The preparation procedure for pollen slides can significantly reduce the total area and number of charcoal particles by physical or chemical removal, impeding the reconstruction of local fire events, however, the selective effects of pollen preparation do not affect the suitability of this method for regional scale fire reconstruction (Tinner *et al*, 1998). The bias towards this method is created due to the analysis being undertaken in conjunction with palynological studies, and thus is advantageous due to time saved by simultaneous sample preparation and counting. However, the sample preparation can have detrimental effects on charcoal particles, which are less robust than pollen grains, and the chemical and mechanically vigorous method can lead to particle fragmentation (Rhodes, 1998 and Kangur, 2002 among others). Pitkänen et al (1999) assessed the impact of chemical preparation on charcoal, comparing sieved samples, and un-sieved samples that were chemically treated. As very large particles were present in un-sieved pollen samples, and corresponded to local fire dates, they concluded there is very little or no fragmentation of large particles during chemical treatment. In an experiment investigating the water-logging and sedimentology of charcoal particles, Nichols *et al* (2000), showed that wood charcoal particles may undergo initial abrasion and attrition, but prolonged agitation has very little effect on size distributions and shape of particles. The principal effect of attrition or abrasion was slight rounding of originally angular edges, and the removal of bark. However, these particles were not subjected to chemical agents in addition to being mechanically stressed, and the results pertain only to wood charcoal fragments.

Tinner and Hu (2003) showed that charcoal size-class distributions are not statistically different between sites (vegetation-type variation) where the same preparation methods are used, but they do differ within the same core when different techniques are applied. They concluded that it is not necessary to determine size-class distinctions in order to reconstruct regional fire history, but just to measure the total area of charcoal particles >75 um^2 contained in pollen slides. The area of charcoal particles is thought to be a more accurate indicator of local fires than number of particles per sample and charcoal concentration values (Kangur, 2002). Several authors (Tolonen, 1985 and Tinner *et al*, 1998) found that charcoal area and number of pieces co-vary if plotted against depth, concluding that measurements of microscopic charcoal area from pollen slides are superfluous for the reconstruction of regional fire history.

Conclusive identification of charcoal based on structural features is often impossible when using the pollen-slide method as charcoal pieces are indiscernible at the usual magnification used for pollen identification, and therefore most analysts rely on angular, uniformity of blackness and opaqueness as the basis for characterisation, which are criteria that can be satisfied by other organic and inorganic particles (e.g. pyrite) (Rhodes, 1998). These characteristics are also used in image analysis, which has been found to consistently underestimate charcoal area compared to visual counts (MacDonald *et al*, 1991).

It is recognised that pollen slide methods for quantifying charcoal are far from perfect, and therefore a wide range of alternative methodologies have been developed. In order to quantify the elemental carbon content of sediments without having to perform counts, chemical digestion assay methods were developed (Winkler, 1984). This method is however, incapable of discriminating between carbon produced by burning of local biomass, and that of high temperature fossil fuel burning (Rhodes, 1998) for more recent sediments.

To reconstruct fire histories of annually laminated lake sediments, petrographic thin sectioning methods can be used, especially for 'local' fire histories with large charcoal particles (Clark, 1989). This method has not been widely adopted, possibly due to the complicated and time-consuming sample preparation technique and the fact that pollen would have to be counted separately (Rhodes, 1998).

The bleaching method developed by Rhodes (1998) was primarily concerned with minimizing the potential for particle fragmentation during sample preparation in order to determine size classes of charcoal, and is effective with lake sediments, peats and mor humus soils. The major advantage to this method is that instead of attempting to remove the organic component of the samples, it is bleached so that the remaining black charcoal particles are readily recognisable under the microscope, preserving the integrity of the charcoal assemblage. In addition, particle identification can be verified by altering the particle orientation or by observing the fracturing behaviour when pressed (charcoal fractures tangentially which is diagnostic) (Rhodes, 1998).

A convenient method for reconstructing local fire history is that of sieving sediments in order to select only the large (>125 μm) charcoal particles. At present the methodologies adopted remain poorly documented in the published literature (Carcaillet *et al*, 2001). The primary concern with the method is the variability concerning minimum mesh diameter (80-200 μm) to exclude particles from distant fires, in addition to the minimum sample size for replicable data (Carcaillet *et al*, 2001). This method has been used to quantify charcoal in sediments that are undisturbed due to anoxic conditions, therefore it should not be assumed to provide replicable series of charcoal concentrations with any other type of sediment.

When considering a time series of samples, sieving methods yield low background components of charcoal, which is due to the exclusion of the small sized long-distance transport component. The background charcoal that is apparent probably originates from the delayed sedimentation of secondary charcoal, which can occur up to 5 years after a fire event (Whitlock and Millspaugh, 1996). Distinct peaks can be identified, which when combined with the low background component lend weight to this method, which provides a less ambiguous local fire history to study long-term fire-vegetation relationships (Carcaillet, 2001). Many variations on a sieving method have been used, with the advantages of minimal particle fragmentation due to reduced mechanical and chemical preparation. Particles are deflocculated and disaggregated with a manual water spray but there is no current agreement on the optimal amount of sediment to sample to provide reliable data (Carcaillet *et al*, 2001).

From the already established protocols, the most advantageous for preservation of assemblage integrity appear to be a combination of the bleaching and the sieving methods. There are no studies published to date that compares the most commonly used methods, with some studies comparing pairs of methods (e.g. Carcaillet *et al*, 2001), but due to time constraints and the fact that methods are particularly suited to different research questions, this is hardly surprising. It would be interesting to run a comparison of methods at the beginning of a project in order to select the most appropriate, rather than relying solely on the 'best fit' option.

There is a possibility of integrating alternative techniques to lend support to the reconstructions determined from charcoal data. For example, living trees are poor producers of charcoal; the majority of wood charcoal is produced by dead trees (Begin and Marguerie, 2002), which has interpretive and dating implications, and it is suggested that the quantification of the proportion of charred versus un-charred needles could indicate the actual species dominance in forests at the time of fire, rather than the species composition of dead/downed wood.

Morphology may reflect the shape of the original parent material to some degree. Umbanhowar and McGrath (1998) investigated the potential of charcoal morphology as an indicator of vegetation type, and found that grass charcoal is longer and narrower than leaf and wood charcoal. They concluded that morphology, especially length:width ratios – may be an indicator of vegetation type, especially at transition sites, such as the forest-tundra ecotone.

Charcoal fragments of woody origin may be identifiable to the family, genus or species level (Carcaillet, 2001) by comparison with reference collections of extant carbonised woods and histological wood atlases (Figueiral and Mosbrugger, 2000). However, in reality, specific identification is frequently not possible due to similarities of wood anatomies between different species. Taxonomic analysis of the macroscopic component of a site may allow some degree of verification of local species composition when compared with the pollen record. The application of quantitative interpretation would require the determination of specific weight for each taxon and each level or profile (Thinon, 1992 cited by Fiqueiral and Mosbrugger, 2000). This method is used most commonly in archaeological work to determine the agricultural and domestic habits of past communities. However, it could be incorporated into any palaeoecological study concerned with fire and vegetation to lend support to the other analyses conducted.

2.9 Fire Ecology

Fire has long been recognised as a natural phenomenon that shapes the vegetation of the boreal forest (Chapman, 1952). However, discussion of the role of fire on the landscape has gained prominence in both the popular and scientific literature since the Yellowstone fires of 1988 (Brunelle-Daines, 2002). In the boreal biome, fire is the major disturbance agent affecting ecosystem change, and fire dynamics will likely change in response to climatic warming (Rupp et al, 2002). Furthermore, changes in vegetation associated with climatic warming could have significant regional effects on landscape processes, including seed dispersal, land-surface feedbacks to climate and fire spread (Chapin et al, 2000; Rupp et al, 2002). There is much evidence that climate is warming at high latitudes (Overpeck et al, 1997), the change being most pronounced in northwest North America and central Siberia (Figure 2.14).

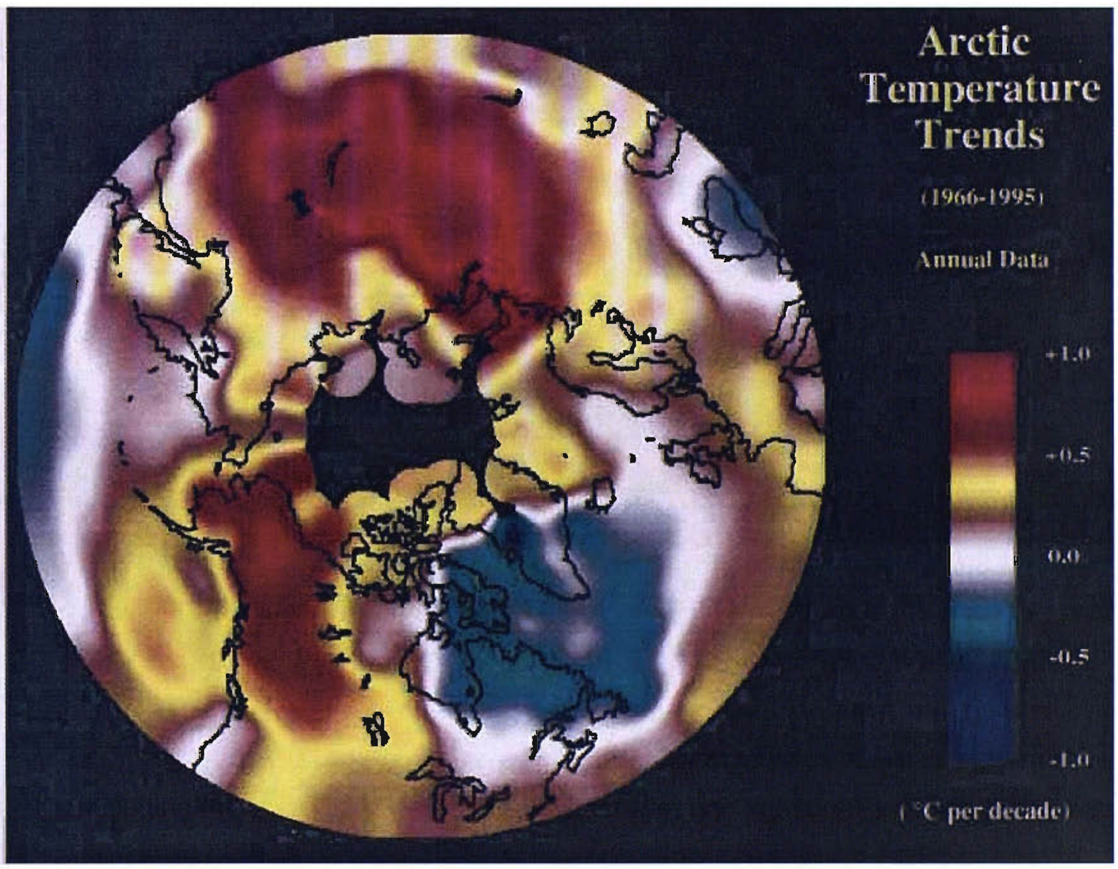


Figure 2.14: Observed changes of annual mean temperature in the Arctic, expressed in terms of a linear trend for the period 1966-1995 (Chapman and Walsh, 1993)

The fire regime is categorised as the characteristics of fire in a given ecosystem, including the frequency, predictability, intensity, and seasonality of fire. It has been predicted that wildfire occurrence/area burned will generally increase with climatic warming (Overpeck et al, 1990), particularly in the boreal forest where predicted future increases in area burned exceed 40% (Flannigan and Van Wagner, 1991, NIFC, 2000) (Figure 2.15). However, the effects of climatic warming on fire regimes may not be uniform. Studies in northwestern Quebec have shown that the post-"Little Ice Age" climate change has profoundly decreased the frequency of fires (Bergeron and Archambault, 1993). This has been linked to the reduced frequency of drought periods since the end of the Little Ice Age. However, this is not a consistent finding, and others have found increases in fire frequency (e.g. Carcaillet et al, 2001). Eastern Canada's climate is very different from Alaska and western Canada, with the former having much higher effective moisture and more maritime influences than the latter two, and it should not be expected that the impacts of climate change on the fire regime will be the same in both areas. As will be discussed in the interpretation of the data generated in this project, the patterns of fire evident in Alaska and western Canada show that fire regimes have increased in association with previous increases in effective moisture and temperature.

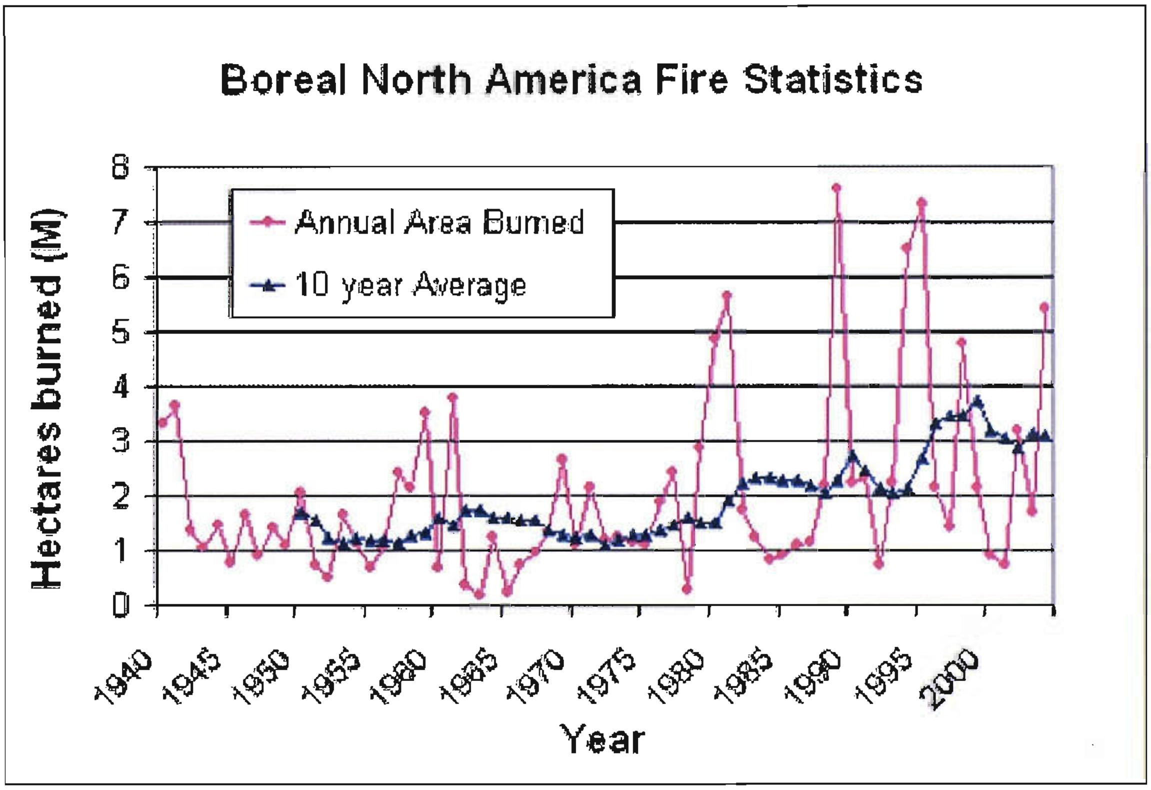


Figure 2.15: Annual area burned in the North American boreal forest illustrating a continuous rise in fire activity since the early 1970s (Kasischke 1999)

In fire-dependent ecosystems distribution of plant species and communities is partly controlled by fire, which in turn regulates fuel accumulation at both the stand and landscape scales (Bourgeau-Chavez *et al.*, 2000). Thus, the vegetation-type mosaic (stand age-class distribution and developmental stages) of a region is in part an expression of its fire regime (Bourgeau-Chavez *et al*, 2000). A fire regime refers to the kind of fire activity or pattern of fires that broadly characterises a given area. Fire regimes are defined by the climate, fuel type, fire cycle/interval, fire season, fire weather, and the number, type and intensity of fires. Climate (regional fire climate and lightning incidence), vegetation (and the balance between dry matter production and decay) and topography all interact with fire and vary over space in complex ways that are not fully understood (Morgan *et al.*, 2001). Additionally, vegetation and fire characteristics are often a legacy of both present and past events (i.e. temporally autocorrelated) (Crutzen and Goldammer, 1993; Morgan et al, 2001). In a model simulation increases in fire frequency and extent appear to cause a shift from conifer-dominated forest to deciduous-dominated forest (Rupp et al, 2002). If such a fire-induced shift were to take place it is probable that a positive vegetation-fire feedback system would further alter future fire regimes (Chapin et al, 2000).

Fire disturbance may also become increasingly important to the global carbon (C) budget. Climate is changing in a region where carbon reserves are large and where fire dominates the distribution of plant and soil carbon (Kasischke, 2000b). Changes in fire regime could alter the carbon balance, producing feedbacks to climate. Although boreal forests currently act as a slight net sink of C over decadal to century time periods, severe fire activity may result in net sources of C developing (Kasischke, 2000a). Alternatively, increases in fire frequency and area burned that increase the proportion of deciduous forests would likely act as a negative feedback to regional warming (Chapin et al, 2000). In addition to C release, ~90% of which is CO_2, forest fires are also one of the most important global sources of particulate matter (Novakov *et al.*, 1997), producing airborne particles that may influence global and regional climate (Einfeld *et al.*, 1991; Damoah *et al.*, 2004). Smoke from forest fires contains varying quantities of CO, NO_x, VOCs and CH_4, all of which will impact the global atmospheric composition, and the quantities released annually can influence climate over the longer-term (Ward & Hardy, 1991), and the smoke itself has been shown to promote a degree of surface cooling (Robock, 1991). The impact of climate warming on the fire regime may yield significant and rapid changes in vegetation such that the change in fire regime could possibly overshadow the direct impact of climatic change on species migration, substitution and extinction (Weber and Flannigan, 1997). The uncertainty of future disturbance rates in the boreal forest validate the study of past fire regimes and their interactions with climate and vegetation, in order to further inform models using palaeodata which can provide information for past environments which have no modern analogues for certain vegetation and/or climate scenarios and thereby aid in the calibration of predictions for future changes.

Understanding fire regimes across temporal scales of decades to centuries or further, and spatial scales from a single patch or stand to regions and continents is challenging for several reasons. Fire is a stochastic, spatially complex disturbance process (Morgan et al, 2001), yet the data for describing fire are typically collected at points or from small areas, and often from relatively short time series. Fire regimes in the North American boreal forest vary from short-interval crown fire/high-intensity surface fire regimes to very long-interval crown-fire regimes (Bourgeau-Chavez et al, 2000). In the Canadian boreal forest, large, high-intensity crown fires are common (Amiro et al, 2001), with on average between 1 and 3 million hectares of forest burning annually, and the average area burned in western Canada in fact doubling in the last two decades (Kasischke et al, 1999). In the Alaska on average 1 million acres burn annually, with over 6 million acres burning in 2004 (NIFC, 2005) and ~5 million acres in 2005 (Rodzell, 2006). It is clear that data from single points or short time series are not capable of describing the full patterns of fire variability. Therefore, it is essential to conduct large-scale studies that will allow regional interactions over long time periods to be analysed (e.g. Swain, 1973).

In this section a discussion of the main variables influencing fire and its interaction with vegetation and climate are presented, as in order to reconstruct fire histories in palaeo-records it is important to understand the theory of modern fire ecology in the first instance.

2.9.1 Fire Season/Weather

The occurrence of fires in Alaska is closely linked to large-scale atmospheric circulation patterns (Hess *et al*, 2001; Kasischke *et al*, 2002), which control summer thunderstorms and lightning patterns, and areas with frequent cloud-to-ground lightning strikes have the highest incidence of fires (Kasischke *et al*, 2002). Seasonal variation in the climate of the boreal forest is controlled by interactions between the Arctic, cool Pacific, mild Pacific and North Atlantic airstreams (Johnson, 1992), and are discussed further in the fire season/weather section. The Arctic airstream dominates the region during the winter, and is replaced in summer by the cool Pacific airstream (Johnson, 1992). The cool Pacific air from the Bering and Chukchi Seas and the proximity of Arctic air north of the Brooks Range give the summer a continuous series of weak cyclones which results in high summer precipitation in Alaska. The Pacific air also causes a slight increase in summer precipitation in the Yukon. The fire season is determined by the summer reorganisation of atmospheric circulation over the boreal forest, switching from a colder stable Arctic air-stream to warmer, unstable air caused by the deep intrusion of Pacific and Tropical air-streams (Johnson, 1992). Air-stream movements and boundaries limit the length of the fire season, as does the appearance and melt of the snowpack, and determines the seasonal geographic progression of fires (Johnson, 1992). Forest fires are possible once the Arctic air-stream has retreated north, as the warmer, unstable Pacific or North Atlantic airstreams are more effective at drying fuels, owing to their higher temperatures, and igniting them due to the generation of more lightning strikes (Johnson, 1992).

Generally the fire season in the North American boreal forest lasts from April to September, with the number of fire starts and area burned peaking in June through August (Bourgeau-Chavez et al, 2000). This is largely due to lightning ignition and a gradual seasonal drying trend for most years. Lightning is the most significant cause of fires in the boreal forest accounting for about 90% of the area burned (Stocks, 1974, Johnson and Rowe, 1975, Stocks and Street, 1983, Johnson, 1992). Lightning caused fires decline northward as the frequency of thunderstorm days declines (Johnson, 1992). Weather determines fire patterns and characteristics by generating the principle ignition source (lightning), controlling the most significant variable of fuel flammability (fuel moisture) and providing the high winds required for rapid spread (Johnson, 1992).

Further, a critical synoptic weather pattern of upper level ridge build-up and breakdown determined by a succession of tropospheric disturbances in the westerly flow usually along the Arctic front, is often responsible for the ignition and spread of large and widespread fires (Johnson, 1992). Strong westerly flows lead to alternations of warm-dry weather during upper level ridges, and cool-wet weather during upper level troughs (Johnson, 1992). However, this sequence can be interrupted by ridges that stall, bringing longer periods of warm-dry weather and lead to the rapid drying of fuel due to high temperatures, low humidity, and usually light winds (Nimchuk, 1983; Johnson, 1992). These blocking ridges tend to develop over Alaska, the Yukon and Northwest Territories during the fire season (Johnson, 1992). Harrington and Flannigan (1987) found that the area burned in the boreal forest of Canada increased with the duration of the dry spells, and indicates that a source of ignition is not a limiting factor in this region.

Generally the requirements for extreme fire behaviour in a forest environment (e.g. high intensity crown fires) are fairly well known; an ignition source, dry and plentiful fuels, strong winds and/or steep slopes, and an unstable atmosphere (Byram, 1954). However; there has been a common misconception that in order to develop a large high-intensity crowning forest fire, prolonged or severe drought is a necessary prerequisite. The propensity for crown fires to occur during the spring throughout much of North America's boreal forest does not concur with the drought theory (Stocks and Walker, 1973). The canopy moisture of the boreal forest is lower than a deciduous canopy, and has a spring and summer decline (Johnson, 1992). This difference in foliar moisture seems to play a significant role in the heat required to ignite the canopy. The conifer spring decline in moisture does not vary in timing from year to year. Hence, it is not a weather induced phenomenon, and it is believed to be responsible for the susceptibility of conifer stands to crown fires early in the fire season (Van Wagner, 1974).

2.9.2 Climate

The fire weather on a seasonal/annual basis influences the more long-term average, or fire regime, which is primarily dependent on climate (Johnson, 1992; Weber and Flannigan, 1997). Van Wagner and Methven (1980) proposed that for each vegetation type to be perpetuated by periodic fire there exists an optimum fire regime that best fulfils its ecological requirements. Although fire regimes are closely related to local site productivity and topography, climate variability entrains regimes at regional to national scales (Morgan *et al*, 2001). Unprecedented increases to regional/seasonal temperatures are expected in the future, with projected changes most pronounced at high latitudes, where they will be greatest in winter (Weber and Flannigan, 1997). Model simulations of future climates have predicted that a drier and warmer climate, as a result of global warming will lead to increased fire activity in boreal systems (Flannigan et al, 2001), and extension of the fire season in addition to altering fire weather, contributing to an increased area burned (Amiro *et al.*, 2001). The fire regime as an ecosystem process is highly sensitive to climate change as fire behaviour responds immediately to fuel moisture in the short term, which is affected by precipitation, relative humid, air temperature and wind speed (Weber and Flannigan, 1997).

Investigations of past fire regimes over the last ~7 ka in Alaska have found that contra-intuitively local fires can occur more frequently under wetter climatic conditions (Lynch *et al,* 2004). Recent studies in eastern Canada have also shown this trend, where higher fire frequencies have occurred under cooler and wetter conditions during the past 3000 years (Carcaillet and Richard, 2000, Carcaillet *et al*, 2001c). Some Alaskan studies also show this pattern for the mid-Holocene (Hu et al, 1996, Lynch *et al*, 2002). When looking at interactions affecting fire over such long time periods, the amount of detail available for climatic reconstructions declines. Climatic indicators such as "cooler" and "wetter" can be misleading. It is important to know which seasons are cooler and wetter, as fires are started only in the spring and summer in Alaska and Canada. Some studies from boreal and temperate regions of North America found higher fire frequencies were associated with warm and dry conditions during the mid-Holocene for example, in Alberta, Manitoba and western Oregon (Clark, 1990; Long *et al*, 1998; Hallett and Walker, 2000; Lynch, 2001). In central Saskatchewan, eastern Ontario and Minnesota fire frequency also increased with lower annual precipitation (Larsen and MacDonald, 1995; Weir *et al*, 2000; Bergeron *et al,* 2001). However, Carcaillet *et al.* (2001c) put forth the idea that precipitation increased during the winter and decreased in the summer during the past 3000 years in eastern Canada. This shift would result in dry surface fuels during the fire season, creating more favourable conditions for fire ignition and spread. Increases in summer precipitation, in comparison, would have led to a decreased potential for fire ignition and spread. Therefore, the seasonal timing of climate variability is critical to the potential impact on the fire regime. Wetter winters and drier summers may not explain the increase in fire frequency in south-east Alaska, an area with a generally wetter climate than in the above mentioned studies.

Hogg and Hurdle (1995) defined the boundaries of major forest types in western Canada based on a moisture index (P-E). Future climate predictions from GCM simulations suggest there could be large changes in forests, associated with changes in moisture, with an expansion of aspen parkland towards the north, and a reduction in "true" boreal forest (Hogg and Hurdle, 1995). Although species migrations may not follow these long-term climatic drivers directly, being influenced more directly by short term seasonal variability, the message is that different fuels could dominate in future climates, which will impact on the fire regime.

2.9.3 Fire Cycles

The fire cycle is generally the term adopted to signify the frequency at which a particular area or site will burn, and is principally dependent on two factors: the time required to build up an available fuel load since the last fire (fuel productivity) and the frequency of ignitions (Whelan, 2002). Natural fire cycles in the boreal forest have been estimated to vary from approximately 40 years in interior Alaska (Yarie, 1981), 126 years for central Canada (Amiro et al, 2001) to about 500 years in southeastern Labrador (Foster, 1983).

Different fire cycles have been linked to different vegetative adaptations to fire (Rowe, 1983). Shorter fire cycles typically found on drier sites favour endurers (for definitions see vegetation and regeneration section) e.g. trembling aspen. Intermediate fire cycles favour resisters and evaders e.g. lodgepole pine and black spruce. Very long fire cycles on moist sites favour avoiders e.g. balsam fir and white spruce. In general shade intolerant pioneer species such as aspen and lodgepole pine are replaced in the absence of fire by shade-tolerant species such as black and white spruce (Rowe, 1983). This is confirmed by studies such as that of Le Goff and Sirois (2004) in Quebec, who investigated how fire interval length can influence the abundance of regenerating black spruce, and modelled the future abundance for fire cycles of varying lengths. Model simulations suggested that fire intervals <60 years lead to the localised extinction of black spruce, with the length of interval between two consecutive fires providing a better prediction of abundance (including local extinction or expansion) of black spruce than the longer-term fire regime estimate.

Le Goff and Sirois (2004) found a predominance of shorter fire intervals in well-drained sites, and longer fire intervals in poorly drained sites in Canada. This agrees with other studies from American boreal forests (e.g. Rowe and Scotter, 1973, Viereck, 1983) which suggest that topography and drainage conditions exert a strong control on site-specific susceptibility to fire.

Alterations to the fire cycle are thought to exert some kind of forcing on plant populations, and to some extent vice-versa. For example, periods of high fire frequency during the Holocene are thought to have triggered the expansion of *Pinus banksiana* (Jack pine) in northern Quebec (Desponts and Payette, 1993) and, similarly, *Pinus pinaster* (Maritime pine) in southern Europe (Carcaillet et al, 1997). However, the large-scale replacement of *Picea mariana* (black spruce) by *P. banksiana* in Minnesota (Brubaker, 1975; Birks, 1976) and pine expansion in New York (Clark *et al,* 1996) during the Holocene were not related to changes in fire frequency. Lavoie and Sirois (1998) found a shift from spruce- to pine–dominated communities in frequently burned sites in Quebec during the 1980's, concluding that fire intensity, short fire interval and unfavourable climate during and after fires are three plausible mechanisms associated with post-fire vegetation changes.

Often, stochastic or mechanistic modelling is the only way of characterising and assessing the variability of fire effects under different fire regimes (Keane and Long, 1998). Model simulations show that vegetation shifts cause substantial changes to fire regimes. Rupp *et al*, (2002) showed that the number of fires and the total area burned increased in their model as black spruce forest became increasingly dominant. In addition they projected that early successional deciduous forest vegetation burned more frequently when black spruce was added to the model.

There is evidence that fires impose some degree of control over the vegetation composition. The vegetation also controls fire through determining the fuel loads and flammability status of a landscape. Therefore, the decomposition of the interaction between fire and vegetation, and the further interaction with climate is essential to understand the future dynamics of the boreal region.

2.9.4 Fire Type

Fires in the boreal forest are characterised by high intensities, which result in extensive mortality in canopy and understory plants (Johnson, 1992). Fire intensity describes the rate at which heat is given off by a flame, and it is this heat transfer which causes the adjacent fuels to be heated and subsequently burn, thereby releasing more heat and propagating the fire (Johnson, 1992; Whelan, 2002).

Fires are typically categorised by the fuel layers that are involved in the combustion process. Three broad types are most commonly recognised: ground fire (Plate 2.2a), surface fire (Plate 2.2b) and crown fire (Plate 2.2d). Generally a burned area will experience all three types at some point during a fire. Three types of crown fires are generally recognised (Van Wagner, 1977). Passive or intermittent crown fires (Plate 2.2c) do not burn continuously in the tree canopies, but burn into the crown then drop back to the surface fuels (Johnson, 1992). In this situation adjacent trees can suffer different fates. Active or dependent crown fires form a wall of flames from the surface into and above the tree canopy but depend on surface fire intensities and continuous moderate canopy bulk densities to sustain them (Johnson, 1992). Independent crown fires burn as the name suggests, independently of the surface fuel layer (Johnson, 1992) and is a rare type, existing for only short periods under very unusual heat transfer conditions. In the upland boreal forest, active crown fires are very common in closed canopied forest, while passive crown fires are common in the open canopied lichen woodlands (Wheelan, 2002). It is very rare for a fire to consume a whole area completely, and islands of unburned forest within the perimeter of a burn are common, constituting up to an estimated 10% of the total area (Bourgeau-Chavez et al, 2000).

Once crowning occurs, the fire has access to the ambient wind field, which greatly accelerates the fire's spread rate until an equilibrium rate of spread is achieved (Bourgeau-Chavez et al, 2000). Active crown fires are the most common type in Canada and Alaska (Bourgeau-Chavez et al, 2000, Amiro et al, 2001). Although the fast spreading active crown fires that dominate North American boreal forests require sufficient fuel availability (low fuel moisture content and significant biomass), they are primarily driven by strong winds and aided by both short- and long-range spotting of fire brands ahead of the flame front (Bourgeau-Chavez et al, 2000). Because active crown fires are common, complete tree mortality is usual; with trees only generally surviving passive crown fires and high intensity ground fires (Johnson, 1992).

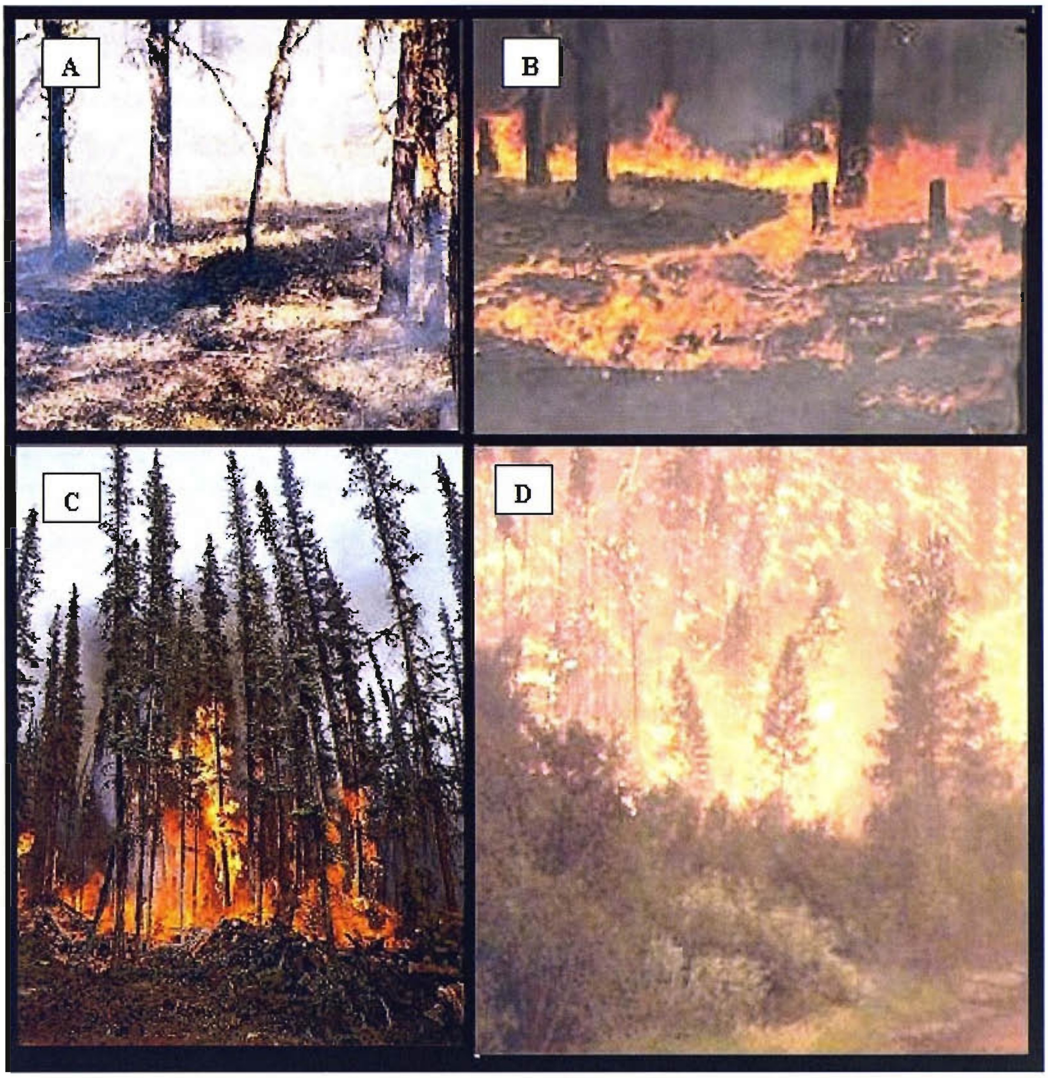


Plate 2.2: Illustration of variation in fire behaviour in Alaska (a) ground fire, (b) surface fire, (c) intermittent crown fire and (d) crown fire. Photographs: LMFS

2.9.5 Fire Spread/Behaviour

Fire spread in forest ecosystems encompasses very complicated phenomena, typically entailing turbulent flow and intricate fuel arrangement, with processes occurring over a wide range of scales (Fendell and Wolff, 2001). The most dominant processes are discussed briefly here, however, as research into fire spread and behaviour is still in its infancy, this is by no means a definitive guide.

The spread of individual fires directly affects plant communities on a short timescale, by re-setting the "successional clock". This is distinct from fire regimes which over many years help to define vegetation mosaics. Fire spread is determined by the rate of heat transfer from the burning matter to the unburnt material surrounding it (Weber, 2001). In the absence of other constraints, a fire will generate its own wind (Whelan, 2002), and the convection of heated gases upward draws in air from around the burning nucleus and the fire front spreads outward, into this self-generated wind, effectively as a "back-fire". Radiation from the front pre-heats the adjacent fuel and thereby dries it for ignition. One side of this initial circle of flames will be converted by the wind into a "head-fire", where the angle of the flame front, pushed over by the wind, causes rapid heating of fuel in front of it (Figure 2.16). With the right kind of fuel and appropriate winds, this head-fire can produce burning fragments that are transported away from the body of fire and set off "spot" fires in advance of the main front (Whelan, 2002).

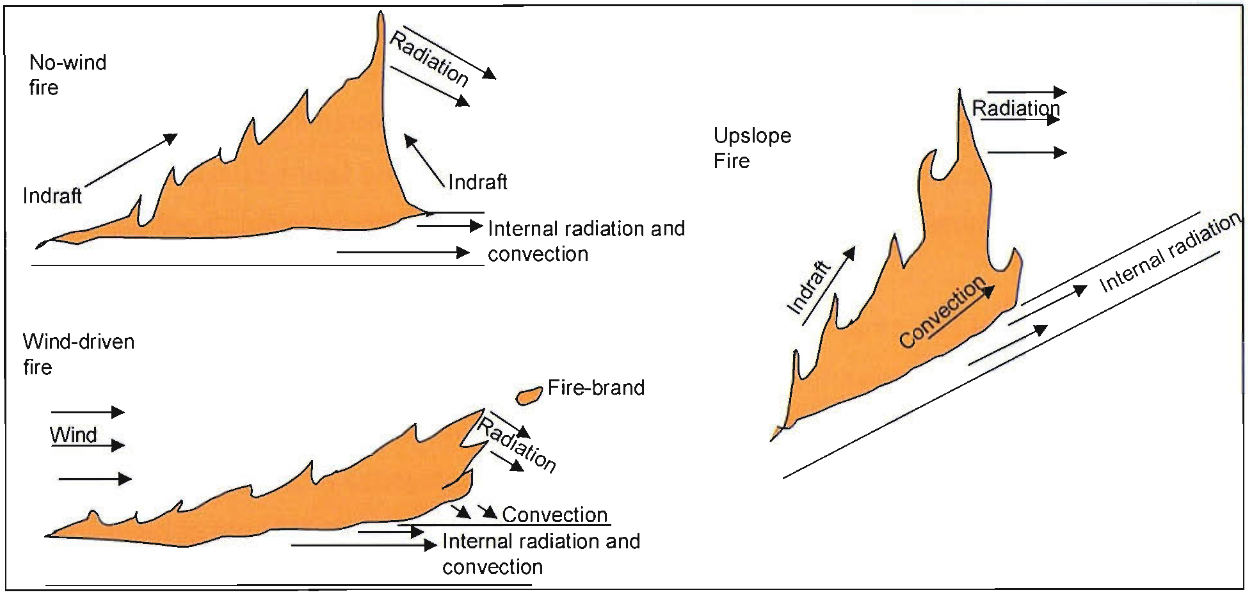


Figure 2.16: Illustration of the similar influences of wind and topography on fire behaviour. Both slope and wind bring the flame front nearer the adjacent, unburnt fuel, enhancing pre-heating and increasing the rate of spread (Adapted from Johnson, 1992).

Topography has a direct influence over the rate of spread of a fire. The effect of slope on a fire front is similar to that of wind (Figure 2.16), with the flames brought nearer to the ground up-slope, and thus fires burn more rapidly up-hill, and the flames taken further away on down-slopes thereby slowing or stopping down-hill. This is of course, a simplified statement, as topography is also a factor which impacts on local climate and vegetation.

As stated previously, a fire burning up-slope would be expected to burn rapidly and intensely all else being equal. In reality however, a large number of additional factors interact. For example, gully and valley vegetation is likely to be different from hilltop vegetation, perhaps more mesophytic and therefore less flammable (Whelan, 2002). In addition to slope, topography also provides natural fire breaks; lakes and rivers are very effective fire breaks!

Vegetation type affects the rate of spread due to variability in flammability (Johnson, 1992), density, fuel availability, and the propensity of a species to produce firebrands that are blown ahead of the fire front producing new ignitions. Chemicals in plants, such as oils and resins, will increase the heat yield of the burning reaction due to their greater energy content (Whelan, 2002). In contrast high concentrations of mineral elements in wood and leaf material can reduce flammability. Fuel continuity also plays an important role in determining whether a fire will spread, especially soon after ignition. Once burning is established, fuel continuity will determine the patchiness of the burned vegetation (Whelan, 2002).

As discussed above, the local climate at the time of the fire, and the preceding climatic conditions will have a strong influence on fire intensity and spread as the rate of combustion of cold, moist fuels is slower than for hot, dry fuels. Site-specific climate will determine the relative humidity of the air which will determine fuel moisture content to a certain extent.

The Canadian Forest Service has conducted measurements and field experiments over a 25-year period to compile the Canadian Forest Fire Behaviour Prediction System (Weber, 2001). This model includes vegetation, topography and weather variables appropriate to Canadian ecosystems to assess potential fire behaviour. The model most widely used in the United States is the Rothermel model, based on the principle of conservation of energy to determine rate of spread, but is over-simplified and does not for example distinguish between different modes of heat transfer. Since the Yellowstone fires of 1988, work has been on-going to update the model to contain sufficient variability to accommodate most wildland fire variables (Weber, 2001).

2.9.6 Fuel Type (vegetation) and Regeneration

Historically, wildfire seems to have always dominated the boreal zone, but until the mid-1960s vegetation was viewed as a passive subject of fire regimes (Mutch, 1970). Since then ecologists have concentrated more on the interactions between fire and vegetation. Vegetation brings properties to the ecosystem structure and function that impacts on the fire regime, and the fire regime determines, in part, the maintenance, regression, or succession of plant communities, although post-fire seral communities are both site- and fire-specific (Johnson and Strang, 1982).

It is generally accepted that the prevalence of wildfire has led to adaptations to fire for many boreal tree species (Clark and Richard, 1996). For example, it is atypical of forests in general to be dominated by wind dispersed seed bearers, e.g. usually less than 10% of tropical tree species are wind dispersed (Anderson, 1983). However, the tree species of the North American boreal forest (both gynosperms and angiosperms) all possess wind dispersed seeds (Johnson, 1992), which aids recolonization of burned sites from a distance. In the cooler northern climates of the boreal forest, adaptations to fire mostly take the form of protection of seed sources to gain competitive advantage for re-vegetating burned areas (Chapman, 1952), and hence the use of cone serotiny (the presence of waxes sealing cones which require heat to release the seed) in lodgepole pine and black spruce. Indeed it has been suggested that species may not only have developed survival mechanisms, but also inherent flammable properties that contribute to the perpetuation of fire-dependent plant communities (Mutch, 1970). By studying the functional adaptations of plants to fire, Rowe (1983) categorised boreal species based on strategies and means of reproduction. The five categories identified were invaders, evaders, avoiders, resisters and endurers. A species can belong to more than one category, if they exhibit more than one type of adaptation.

Invaders are pioneer species inhabiting recently disturbed landscapes. They have wind-disseminated seeds which have short-lived viability. Paper birch, white spruce and trembling aspen are examples.
Avoiders are shade-tolerant species that invade forests later in the successional chronosequence. As such, white spruce is an avoider, as is balsam fir.
Endurers vegetatively reproduce after fires. They have shallow to deeply buried vegetative parts (roots and tree trunks) that often survive a fire and are capable of sprouting. Trembling aspen and paper birch are also endurers.
Evaders are species whose seeds are both long-lived and are stored in places protected from fire (e.g. serotinous cones). Lodgepole pine and black spruce are evaders.
Resisters can survive low-intensity ground fires. Lodgepole pine is a resister once its canopy is high enough above the ground to avoid the potential for crown fires.

Post fire regeneration of a site is largely a function of pre-fire vegetation patterns, neighbouring vegetation, fire regimes and permafrost conditions (Bourgeau-Chavez et al, 2000). In addition, the depth of burn of the organic ground layers influences the type of regeneration on a given site as fire plays an important role in nutrient cycling. Rates of decomposition are generally low in the boreal forest, resulting in large organic mat build up over time, which once burnt results in direct nutrient mineralization. Increased soil temperature and microbial activity lead to heightened nutrient release through organic matter decomposition for several years post-burn (Bourgeau-Chavez *et al*, 2000, Johnson *et al*, 1992). This promotes favourable conditions for the establishment of seedlings and increased site productivity, especially in regions underlain by permafrost. Light burns result in high root sprouting and other forms of vegetative reproduction, whereas deep burns kill roots but create favourable seed beds.

While general regeneration sequences can be identified, site specific variables can lead to variations in the response to fire. For example, black spruce is both an early and late successional species (Bourgeau-Chavez et al, 2000). Fire in black spruce will generally lead to regeneration of black spruce; however, in some regions such as the Cordillera it will often be replaced by lodgepole pine (Johnstone and Chapin, 2003). In Alaska, instances have been found where severe fire that removes most of the organic layer allows aspen to invade black spruce forests (Chapin, 2004). Fire in aspen or lodgepole pine dominated sites will usually regenerate with the original species, in contrast, fire in white spruce or balsam fir will regenerate willow, aspen, alder, black spruce, lodgepole pine, or a combination depending on the seed source, climate and site conditions (Bourgeau-Chavez et al, 2000). Species specific responses are discussed in more detail below.

2.9.6.1 Aspen

Aspen is a short-lived, shade-intolerant, clonal species that frequently reproduces vegetatively post-fire (Johnstone, 2005). Despite being ecologically adapted to disturbance, aspen stands do not ignite easily and require specific site and climatic conditions for fires to occur (Wright and Bailey, 1982). For example, aspen forests are most flammable in the spring and late summer/early autumn during leaf drop (Bourgeau-Chavez et al, 2000). At these times, the surface vegetation is dormant and dry. In general fires in young aspen stands are of low intensity (Bourgeau-Chavez et al, 2000), with crowning fires not occurring unless there is sufficient dead fuel on the forest floor. Therefore, older stands with higher fuel accumulations tend to experience higher-intensity fires (Johnson, 1992; Bourgeau-Chavez, et al, 2000). In the absence of fire aspen in more southern regions, is replaced over time by shade-tolerant conifers or grasses and shrubs depending on the available seed source (Krebil, 1972).

Due to its relatively short lifespan, favourable adaptations to fire and shade intolerance, pure stands of aspen are perpetuated only by the higher fire frequencies. Early canopy development by asexually regenerating aspen after fire reduces conifer establishment, which favours the dominance of aspen across future disturbance cycles (Johnstone, 2005).

2.9.6.2 Lodgepole pine

Lodgepole pine has traditionally been regarded as fire-maintained (Lotan, 1976) as it occurs over a wide range of sites – from mesic to flooded to dry exposed sites – and the fire regimes are thus highly variable. Lodgepole pine is an aggressive pioneer species throughout much of its range, and produces serotinous cones (Lotan, 1976), enabling it to readily establish itself on newly burnt areas. Lodgepole pine trees produce cones with viable seeds at less than 10 years of age, with fires not only releasing the seeds, but also often providing the bare mineral soils that are the best substrate for seed germination (Bourgeau-Chavez et al, 2000. Critchfield (1957, cited in Lotan, 1976) proposed that atmospheric moisture and fire probably instigated the development of cone characteristics centred on serotiny. Lotan (1976) attributes the most direct influence over the adaptation as fire, due to the fact that moisture itself would affect fire frequency and intensity. The ecological significance of the serotinous cone was illustrated at the Sleeping Child Burn in the Bitterroot National Forest in Montana (Lotan, 1976), where 28,000 acres of lodgepole pine forest burned in 1961. Post-fire, lodgepole pine established itself in extremely high densities through most of the site. As lodgepole pine does not normally disperse sufficient seed for restocking beyond 60m, most of the burned area was restocked primarily from seeds in serotinous cones that burned in the fire.

The most flammable lodgepole pine stands are ones with an abundant lichen ground cover (Bourgeau-Chavez et al, 2000). The most common type of fire occurring in lodgepole pine stands is a low intensity surface fire, a fire type common when surface fuels are low and canopy bases are high. Crown fires are usually of low intensity in these forests, unless there is a large accumulation of woody debris on the ground which leads to stand-replacing high-intensity fires (Bourgeau-Chavez et al, 2000). Under severe drought conditions high-intensity crown fires occur in pine forests and burn very large areas, resulting in mortality of the whole canopy and understory trees, producing conditions highly favourable for regeneration and dominance of lodgepole pine.

Immediately post-fire there is typically an initial herb and shrub stage, which is short lived, and quickly progresses to overall dominance by lodgepole pine seedlings and saplings, which establish high abundances due to rapid germination and growth out-compete other species seedlings. If pine forests reach full maturity an understory of spruce and sub-alpine fir would develop. Usually however, before this stage is achieved a stand-thinning or stand-replacing fire occurs and dominance of lodgepole pine is maintained. In the absence of periodic fires lodgepole pine tends to be replaced by more shade tolerant species between ~50 and 200 years post burn (Lotan et al, 1984). In mixed stands, the proportion of lodgepole pine increases with each recurring fire (Lotan, 1976;

Johnstone and Chapin, 2003), indicating a capacity for rapid local population expansion. Light periodic surface fires may eliminate the shade-tolerant understory species and when the canopy becomes open enough lodgepole pine will regenerate in the understory to maintain dominance (Bourgeau-Chavez et al, 2000).

2.9.6.3 Black spruce

Black spruce retains low dead branches and has a low crown base; and therefore, crown fires are common with the shrub layer acting as a fuel ladder to the dead branches which act as a ladder into the canopy. Black spruce is especially well adapted to high fire frequencies. Most fires are high-intensity crown or surface fires that kill overstory trees (Viereck, 1983) due to a combination of low precipitation, ericaceous (and flammable) shrub layers, and persistent lower branches. Fire tends to be more frequent in black spruce stands of western Canada and Alaska, with frequency decreasing towards eastern Canada due to a more humid climate (Bourgeau-Chavez et al, 2000).

Black spruce produces cones as early as 10-15 years after fire with optimum seed production between ~50 and 150 years, allowing survival and regeneration under a range of fire frequencies (Bourgeau-Chavez et al, 2000), although with varying success (see above).

Black spruce stands are found on both upland and lowland sites. There are three main types of black spruce forests in terms of fuel types. Firstly, spruce-lichen woodlands consist of open stands that grow on well drained uplands with continuous mats of lichen ground cover with thin organic layers (Johnson 1992). Secondly, boreal-spruce types are moderately well-stocked closed canopy stands with a carpet of feather or sphagnum mosses and deep organic layers on uplands or lowlands (Johnson 1992). Finally, lowland bog sites exist which grade into muskeg. Spruce-lichen woodlands are much more flammable than the boreal spruce type as the canopy is very open, allowing direct insolation and drying of the lichen covered forest floor (Bourgeau-Chavez et al, 2000).

The successional chronosequence found in Alaskan black spruce forests was described by Van Cleeve and Viereck (1981). Vegetative regeneration begins almost immediately post-fire in burned black spruce boreal forest via sexual and asexual reproduction. After a fire (or the following spring depending on the timing of the fire) the charred landscape is invaded by mosses, liverworts, lichens and herbaceous plants during the initial stages regeneration that can last from 2 to 5 years, depending on the site and burn severity (Van Cleeve and Viereck, 1981). Black spruce seedlings become established during this stage, except in sites where severe fires have occurred and removed most of the organic soils, where aspen and birch seedlings can also establish (Kasischke et al, 2000).

For moist black spruce-feather moss forests, the next successional stage, lasting from 5 to 25 years, is dominated by re-sprouted shrubs. Black spruce is slow growing on these sites due to cold, wet, shallow soils and low nutrient availability, and is usually the only tree species to successfully grow in these sites. Black spruce trees begin to dominate the site between 25 and 50 years post fire. Feather mosses and sphagnum invade and the permafrost layer begins to rebuild to its pre-fire depth (Vireck 1983). As the canopy closes between ~50 to 100 years stands thin to a lower density. After 100 years, in the absence of fire, gaps begin to appear in the canopy, the density decreases further, and the system begins to decline (Bourgeau-Chavez et al, 2000).

For open-canopied black spruce-lichen forests, the initial stage of regeneration is dominated by mosses, lichens and sometimes grasses, in addition to black spruce seedlings (Duchesne *et al*, 2006). Between ~10-60 years the ground cover is dominated with fruticose lichens and low ericaceous shrubs (Black and Bliss 1978). Layering becomes an efficient regeneration mode ~50-75 years post fire (Lavoie and Sirois, 1998). By 100 years post fire, in the absence of fire, an open black spruce canopy has formed through self thinning mechanisms (Borgeau-Chavez et al, 2000).

Black spruce bogs are common throughout the boreal forest, where black spruce can occur in pure stands or in association with paper birch. Due to the wet conditions they are not as susceptible to fire as other lowland and upland forests. Very thick organic layers develop, with a very high moisture content which deters fire ignition. However, in severe drought conditions bogs will burn, which can lead to increased paludification, as the transpiring canopy is removed, the water table rises and light levels increase, promoting establishment and growth of sphagnum mosses (Bourgeau-Chavez et al, 2000). Conditions are not conducive to germination of tree seedlings, and thus a tundra-like community may develop (Duchesne *et al*, 2006).

Both regional and local factors influence post-fire regeneration success in black spruce. Regional factors are associated with latitudinal variations in climate (Sirois and Payette, 1991), given the high susceptibility of black spruce germinants to water stress (Black and Bliss, 1980), a gradient of increasing precipitation southwards could be conducive to higher post-fire survival of burned but not killed trees in the south. Local factors include the regenerative capacity of the tree population, fire intensity and severity, and competition for light and water (Sirois and Payette, 1991), and appear to be responsible for high intra-regional variability in the success of post-fire regeneration.

Drastically reduced post-fire regeneration has been reported in young almost pure black spruce stands in eastern Canada (Lavoie and Sirois, 1998) that have not had sufficient time to build up an aerial seed bank during the preceding fire interval. Indeed most studies in sub-arctic and boreal sites (e.g. Sirois and Payette, 1989) show that when no living seed-bearer remains, post-fire black spruce populations are typically even aged, with the bulk of seedlings established during the first ~20-25 years after a fire. The regenerated individuals gradually start seed production by the age of 25, but at the same time the seedbed suitability for black spruce decreases (Sirois and Payette, 1991). Therefore, frequent fires (<20 years) could eliminate black spruce recovery on a landscape.

2.9.6.4 White spruce

White spruce also retains lower, dead branches and low crown bases, leading to common crown fires (Bourgeau-Chavez et al, 2000). Upland mature white spruce forests accumulate deep organic layers making them susceptible to fire. White spruce will sometimes regenerate post-fire if fire occurs in late summer during seed dispersal (Bourgeau-Chavez et al, 2000). Seed regeneration is improved if the moss and lichen are burned exposing the mineral soil. However, white spruce is often out-competed by pioneer species that dominate the site for several years (Bourgeau-Chavez et al, 2000). The increase in abundance of white spruce towards tree line in western Canada may be attributed in part to the decreasing fire frequency trend (Johnson, 1979).

White spruce on floodplain sites are less susceptible to fire due to high moisture conditions in the ground layer, and shrubs (*Alnus* and *Salix* spp.) in the understory that are relatively non-flammable during the growing season. Typically these forests have fire cycles greater than 300 years (Rowe and Scotter 1973). In floodplain sites regeneration generally begins with mosses and herbs in the first 5 years, followed by willow dominance in years 6-30. Trembling aspen, white spruce and paper birch typically establish during the first 5 years, with the deciduous species taking over the canopy between years 31 and 45 dominating for up to 150 years. In contrast white spruce slowly develops in the understory and eventually dominates the canopy between 150 and 300 years (Bourgeau-Chavez et al, 2000).

White spruce and black spruce occupy largely overlapping ranges extending from eastern Canada to western Alaska (Nienstaedt and Zasada, 1990). In interior Alaska, mixed stands of white spruce and black spruce occur frequently (Yarie, 1983; Viereck and Johnston, 1990). Past studies have suggested that white spruce precedes black spruce in primary succession – "The Drury Hypothesis" (Drury, 1956; Viereck, 1970; Van Cleve et al., 1991; Viereck, Dyrness and Foote, 1993). Succession is thought to be driven by permafrost development and organic-matter accumulation together causing paludification which alters environmental conditions to favour black spruce over white spruce (Bonan and Korzuhin, 1989). The Drury Hypothesis has been a central paradigm in ecological studies of the Alaskan boreal forest for the last 35 years (Viereck, 1970; Van Cleve and Viereck, 1981; Van Cleve et al, 1986; Viereck et al, 1993; Mann et al, 1995).

Field data and models have tended to suggest that trembling aspen, paper birch and white spruce are the first tree species to re-establish by seed on all but the coldest of sites. White spruce grows more slowly than the hardwoods and therefore, becomes an important part of the canopy 100 years or more after a disturbance (Walker et al, 1986). Over time, dominance by white spruce results in thicker organic layers and reduced heat flux to the soil as its litter degrades more slowly than that of hardwoods (Anderson and Brubaker, 1994). In the absence of disturbance, litter and moss thickness continues to increase and as a consequence makes the site less suitable for white spruce (Van Cleve et al, 1991). In this late stage, permafrost reaches close to the soil surface severely impeding soil drainage, and black spruce, occasionally in mixtures with larch, becomes the most important tree species (Anderson and Brubaker, 1994). At this stage fires or other disturbances are seldom intense enough to remove the entire soil organic layer, consequently favouring the continued dominance of black spruce forest (Anderson and Brubaker, 1994).

The white spruce to black spruce transition although theoretically sound have not been well documented with observations on the modern landscape (Viereck et al, 1993) and it has been suggested that the Drury Hypothesis is an over simplification (Mann et al, 1995). Mann et al (1995) found that white spruce persistence on floodplain sites is consistent with a dynamic equilibrium existing between the spruce species abundances and the fire/flood disturbances that initiate secondary succession, and affect edaphic conditions of temperature, organic accumulation and moisture.

2.9.7 Establishment biology

The process that directly modifies species recruitment success through succession is typically viewed as competition for light, with the intolerant species gradually becoming unable to recruit, and being replaced by more shade-tolerant tree species. Light regulated processes could explain the lack of recruitment of intolerant deciduous species under their own canopy which results in the replacement of the deciduous species by more shade tolerant species such as conifers (Simard et al, 1998). However, recruitment success of conifers is not primarily explained by shade tolerance. For example, white spruce should be able to recruit as well as any other shade-tolerant conifer in the early successional hardwood stands, but its seedling density is always very low compared to other species (Kneeshaw and Bergeron, 1996). Regulators of seedling abundance such as predation, seed viability, and life histories may explain this trend. However, if one also considers that conifer seedlings are very rarely randomly distributed on the forest floor (Maguire and Forman, 1983), it seems probable that young seedlings are associated with particular micro-sites (Simard et al, 1998). For example dead logs and stumps (Takahashi, 1994), areas devoid of leaf litter (Bartlett et al, 1991) or areas lacking herbaceous vegetation (Maguire and Forman, 1983). Simard et al, (1998) found that most conifer seedlings less than 30cm tall were found on micro-sites with a higher than average light intensity and a lower than average cover of surrounding vegetation and which were devoid of litter, especially broad-leaf litter. White spruce seedlings, for instance, are frequently found on rotten logs, which occupy a significantly larger area in mid-successional stands than in early successional deciduous or late successional coniferous stands. In addition all seedlings were associated with substrates where competition is low and moisture was abundant.

Complete separation of the effects of fire, vegetation or climate in the boreal forest for the modern system is not possible as the interactions are extremely complex and variable. Although, generalised accounts of successional trends were provided above, the actual fire ecology and post-fire regeneration for a particular site will vary depending on climate, physiography, landform, season of burn, age of pre-burn trees, neighbouring ecosystems, post-fire weather patterns and other (possibly unknown) factors (Bourgeau-Chavez et al, 2000). It has been illustrated that although modern fire ecology has established certain patterns and/or rules for boreal forest vegetation and fire regimes, these patterns are not so clear over extended temporal or spatial scales. In order to unravel the interactions between fire, climate and vegetation in palaeo records it is essential to have high resolution records, with significant temporal length either side of a vegetation and/or climate change, to determine the order in which events took place. Only then will reconstruction of palaeo-fire and vegetation history be possible.

2.10 Palaeo-Fire history reconstructions

There are relatively limited data available on palaeo-fire occurrence in Alaska and the Yukon Territory, and those that are published are contradictory. Some authors suggest that higher fire frequencies are associated with the early Holocene (e.g. Earle *et al.*, 1996), when climate data suggest that it was warmer and drier than today (Kaufman *et al.*, 2004). Others document higher frequencies in the late Holocene (Hu *et al.*, 1993; Hu *et al.*, 1996; Lynch *et al.*, 2002; Lynch *et al.*, 2004) when the climate became wetter and cooler.

These discrepancies may be due to climatic factors. The apparent importance of climate in determining fire was demonstrated by Carcaillet *et al.* (2001a; 2001b), where significant changes in the fire occurrence were not accompanied by alteration in vegetation composition to more flammable species. It may be that differences are moderated by the timing of species immigration, or by associated soil developmental or edaphic conditions. There have been studies which illustrate alterations in fire history corresponding with a change from conifer (*Picea-Pinus*) to hardwood-dominated forests in an early Holocene record from northern New York (Clark *et al.*, 1996), where fire occurrence decreased across the transition. This study suggests that the forest type has a direct influence over the determination of fire event frequencies in addition to climatic conditions.

This study aims to present a regional-based reconstruction to identify trends within the fire regime's response to climate and/or vegetation shifts in order to clarify some of the discrepancies associated with the single-site approach of most published studies.

2.11 Stomata

The migration and range of lodgepole pine (Figure 2.17) has been extensively studied using fossil pollen analyses to reconstruct the last 14ka following the onset of deglaciation of the North American Ice Sheets (MacDonald & Cwynar, 1985; Cwynar & MacDonald, 1987; Delcourt & Delcourt, 1987; Davis *et al.*, 1991; MacDonald & Cwynar, 1991). Invasion of an area by a species consists of an initial arrival, followed by establishment on the landscape and subsequent population increase to the ecologically and environmentally determined carrying capacity (McLeod and MacDonald, 1997). However, although migrational patterns are generally well known, there is relatively little research and information available on postglacial tree population growth and expansion (Bennett, 1983; Tsukada, 1983; MacDonald & Cwynar, 1991), with attention turning more in this direction recently (Giesecke, 2005). In order to predict responses of modern vegetation to forecasted future climate change it is essential to be able to clearly understand the relationship between climate and vegetation (Hansen et al, 1996).

Johnstone and Chapin (2003) propose that lodgepole pine is still expanding northwards due to the lack of climatic limitation; as shown by its ability to grow healthily further north and west of its current range. This further undermines the ability to successfully predict future vegetation changes if the current vegetation is not in equilibrium with the climate. The regional and local composition of vegetation surrounding the sites under investigation in this study will impact on the interpretation of the fire regime, and therefore it is essential that the vegetation histories are able to provide information on the species abundances within the individual catchments.

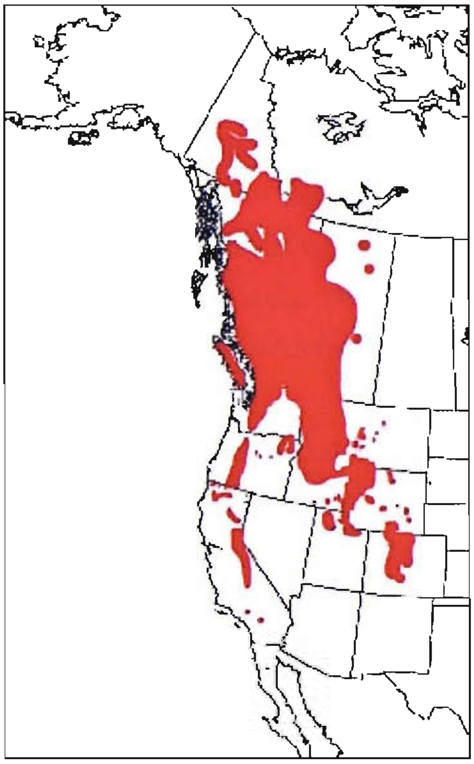


Figure 2.17: Modern geographic range of Lodgepole pine (*Pinus contorta*). Source: http://www.nearctica.com/trees/conifer/conlogo.jpg

Palynology is a well established and standard technique used to reconstruct palaeoenvironments (Birks & Birks, 1980; Faegri & Iversen, 1989). But whilst inference from fossil pollen can distinguish large scale vegetation zones and assemblages, precise vegetation boundaries and local presence of species are often difficult to confirm from the pollen record alone (Clayden et al, 1997, Pisaric et al, 2001). Pollen percentages have been found to not correspond numerically with vegetation percentages, due to the biases in production and dispersal which vary between species (Prentice, 1985). MacDonald and Ritchie (1986) found that lodgepole pine pollen contributes a minimum of 15-20% to the pollen sum of areas when it is only present in low numbers on the landscape. This has fuelled the debate over whether low concentrations of pollen from prolific producers should be regarded as small local populations or distant colonies (Parshall, 1999). The presence of pollen alone, therefore, cannot be regarded as an indisputable indicator of species presence, even when percentages are fairly high (Hansen, 1995).

Conventionally thresholds have been assigned to indicate local presence of different species (e.g. 15% for lodgepole pine, MacDonald and Cwynar, 1985; MacDonald and Ritchie, 1986; Cwynar and MacDonald, 1987). Of course, such thresholds are not independent of the relative abundance of other species (Fagerlind, 1952), and Smith and Pilcher (1973) proposed an “empirical-limit” whereby taxa could be considered locally present when their pollen is first consistently recovered from consecutive samples, although this is generally restricted to taxa that do not produce abundant pollen. However, even in this case, species ranges have been found to be poorly represented by the application of thresholds to pollen values (McLachlan and Clark, 2004). Pollen influx values are more reliable than pollen percentages (Hyvarinen, 1975, 1976; cited in Lamb & Edwards, 1988; Gervais & MacDonald, 2001), assuming that there is a steady relationship between species populations and pollen influx (MacDonald and Cwynar, 1991), but in lacustrine systems this is at the mercy of sedimentary dynamics (eg sediment focussing) (Lamb, 1985; Gervais *et al.*, 2002).

As it is generally accepted that fossil pollen and plant macrofossil data provide limited information on distribution and abundance of small or diffuse tree populations (McLachlan & Clark, 2004), it seems odd that most reconstructions use estimates of potential migration rates reliant on assumptions from fossil pollen and not macrofossil records. Species existing in low densities over extensive areas in the past are not able to be accurately mapped using fossil data (McLachlan & Clark, 2004). Therefore low initial pollen occurrence and detection of species arrival requires the application of additional techniques (Sweeney, 2004).

Stomatal analysis has been undergoing rapid development as a palaeo-technique to more accurately reconstruct plant migrations and expansions (e.g. Froyd, 2005). The stoma is part of the epidermis of the needle, in the form of a pore surrounded by guard cells. They are particularly resistant to decay and are indicators of conifer needle presence. Stomata are generally thought not to be dispersed far from their source (Clayden *et al.*, 1996), and therefore, their presence in sediments is an indicator of local presence (Carlson & Finney, 2004; Leitner & Gajewski, 2004).

It has been shown that species presence represented by stomata are limited to a source area of approximately 0.1ha (Parshall, 1999), which is a significant improvement for local reconstructions when compared to the 1-3ha representation of pollen (Sugita, 1994; Parshall, 1999). Therefore the presence of stomata in lake sediments should give a more reliable indication of local presence of a particular species (Hansen, 1995; Sweeney, 2004). However, it is necessary to note that absence of stomata does not necessarily mean populations were absent (Jackson et al., 1997; McLachlan and Clark, 2004), and it has been suggested that its possible for them to be transported longer distances in fire ashes (Leitner & Gajewski, 2004).

Coniferous stomata can be identified to genus level (Parshall, 1999), by the thickness of the upper woody lamellae and their relationship to the lower lamellae (Trautmann, 1953). Advantageously, samples can be prepared and counted simultaneously with pollen, using conventional pollen preparation methods (Parshall, 1999; Pisaric *et al.*, 2000; Pisaric *et al.*, 2001; Sweeney, 2004).

The use of stomata data can support evidence about species range and migrations (Birks, 2003; McLachlan & Clark, 2004), as when used in conjunction with high pollen percentages in spatially contiguous sites, they are generally more accurate (Jackson et al, 1997), and at the very least supplement the pollen record in the absence of macrofossils (Hansen et al, 1996). Lake parameters are not thought to affect stomata representation, as is the case for macrofossils (Hansen, 1995; Hansen *et al.*, 1996), but due to dilution effects they are less abundant in lake sediments in comparison to peat (Hansen, 1995). Stomata are usually present in higher concentrations than intact needles and/or seeds, and are therefore capable of being used quantitatively in sub-arctic sediments where macrofossils are rare/absent (Clayden *et al.*, 1996; Hansen *et al.*, 1996).

This technique has the potential to become a standard palaeo-proxy, and it is increasingly used to supplement palynology to improve resolution and determination of vegetation boundaries (Gervais *et al.*, 2002) and pinpoint the local arrival of species (Hansen, 1995; Clayden *et al.*, 1996; Hansen *et al.*, 1996; Clayden *et al.*, 1997; Carlson & Finney, 2004; Froyd, 2005). It has been recommended as an adjunct to pollen studies in areas where conifers dominate or are an important component of the vegetation and vegetation history (Hansen, 1995). The identification of conifer stomata has been used in paleoecological studies of tree-line fluctuations in Europe (Trautmann, 1953; Ammann and Wick, 1993; Wick, 1994), Canada (Hansen *et al.*, 1996) and America (e.g. Hansen, 1994; Hansen, 1995).

As the study area is a conifer-dominated system, population expansion dynamics have a potential to alter the flammability status and fuel load of the landscape, stomatal analysis will be conducted in this study to further inform the vegetation reconstruction, and ultimately the fire history of the sites used.

Stomatal implications

Through the discussion of the potential impact of the arrival pine at the sites on the fire regime, it has been concluded that it is necessary to conduct stomatal analyses to further inform the interpretation of the charcoal record. In addition, a secondary question arises concerning the biogeography of lodgepole pine in the region. Stomatal analyses will be used to refine the vegetation histories of each of the sites, and potentially lead to a re-evaluation of the conventional view of lodgepole pine migration history

Chapter 3. Methods

3.1 Introduction

The methods used during this study for determining fire regimes in association with past climate and vegetation included the analysis of the physical properties of the sediment, in addition to pollen analysis, stable-isotope analysis, and charcoal analysis. Radiocarbon dating is used in order to correlate the records between the sites, in addition to known and dated tephra, in the Yukon cores, to create robust chronologies with radiocarbon dates that are on average no more than 1000 years apart. Tephra aid the radiocarbon based chronology by acting as constraining points for the age-depth models. Due to the vast amounts of analyses required to complete a regional fire history reconstruction linked to transitions in climate and vegetation not all the analyses were performed in full by the author, although the majority were either fully, or partly completed by her. For a complete breakdown of which analyst performed each individual analysis please refer to the "Table of Analyses" on page xii.

3.2 Site Selection and Coring Strategy

Alaska and the Yukon Territory are areas rich in lakes, relatively few of which have been studied for vegetation reconstruction in the past. For the purpose of this project, and the objectives outlined in Section 1.2, site selection was based upon a number of criteria outlined here. The main factor determining site choice was that the basin was closed, or virtually so, to reduce the possibility of charcoal sources which are outside the lake catchment being transported into the lake (Faegri & Iversen, 1989). It was necessary to core small lakes (see Table 3.1), in order to ensure that the macro-charcoal record is reflective of the vegetation immediately around the lakes (Clark & Royall, 1996; Whitlock & Millspaugh, 1996). Seven lakes were selected, three of which have been used in previous studies addressing other palaeoecological issues, with this project constructing charcoal records as a further proxy.

Sites were chosen to cover the ~10 ka deciduous-spruce transition, and/or the spruce-pine transition, which was thought to be a late Holocene time-transgressive event among sites. Pairs of lakes were selected that were situated either within current pine-dominated vegetation or within spruce-dominated vegetation within the region of the current pine invasion in the Yukon Territory. Pine is regionally absent in Alaska, and therefore both lakes were selected due to known vegetation histories containing the deciduous- to spruce-transition.

Due to the limited number of published charcoal records in the region, a distribution of sites was sought to better enable an assessment of regional variability. The two sites in Alaska were selected in order to analyse a long pre-spruce record, as this part of the record is generally very short in the Yukon due to late onset of local deglaciation. The Yukon lakes were selected to have either spruce- or pine-dominated catchments, within the current geographical range of pine and therefore form a transect from north of Whitehorse, to the border with British Columbia. Site choice was constrained to some extent by the accessibility of lakes, although sites that were in direct proximity to potential human influence (i.e. situated close to, and likely influenced by, major highways, or areas undergoing construction) were avoided.

In consideration of the factors described above, the following lakes were selected for the study (Table 3.1). Their distribution is shown in Figures 4.1 (Alaska) and 4.8 (Yukon Territory) (Chapter 4).

SITE	REGION	LAT N - LONG W	AREA	SPRUCE INCREASE PRESENT?	PRESENT VEGETATION AT SITE/REGIONAL PINE RISE PRESENT?	SITE CODE USED
Salmo	Yukon	60.444580 133.562590	0.3km^2	Pre-spruce record very short	Spruce in pine region / yes	SLA
Dragonfly	Yukon	60.767 135.417	0.5 km^2	Pre–spruce record very short	Spruce in pine region / yes	DFLA
Jelly Bean	Yukon	60.35145 134.80563	0.4 km^2	Unknown, unlikely	Spruce in pine region / yes	JBL
Haircut	Yukon	60.516190 133.211950	1.1km^2	No	Pine in pine region / yes	HCLA
Marcella	Yukon	60.073690 133.807920	0.4 km^2	Yes – record very short	Pine and aspen in pine region / yes	MLA-Z
Little Harding	Alaska	64 25 20 146 51 10	1.0km^2	Yes	Spruce, not in pine region / no	LHLZ
Birch	Alaska	64 19 146 50	3.0km^2	Yes	Spruce, not in pine region / no	BL97B

Table 3.1: Lake sites used in the study. Sites are listed by name, with latitude and longitude, approximate area and current catchment vegetation. Site name abbreviations are included here for ease, and are also included in the list of abbreviations on page xi.

Coring strategy in the three previously studied lakes was constrained by the primary objectives of those specific projects. However, all cores used were taken from the deepest part of the lake basin as determined by bathymetry, and/or in a gently sloping area, and as far as possible away from steep slopes or shelves, to avoid the possibility of sediment slumping etc. compromising stratigraphic integrity.

Each lake was sampled using a modified Livingstone piston corer with a 6cm barrel (Wright, 1968, 1980), with a 20cm overlap between cores. All lakes were cored to refusal, or basal material was obtained. Cores were extruded into carbon-stable plastic tubing, wrapped in cellophane, secured with duct-tape and labelled. On return to the laboratory cores were split into work and archive halves, and stored horizontally in cold stores (<4°C).

Surface samples were taken using a piston tube sampler (~10cm diameter) to preserve the sediment-water interface (Rowley & Dahl, 1956). The cores were extruded in the field at either 0.5 or 1cm intervals, into carbon-stable plastic bags until sediment stability was reached. Cores were then sealed with bungs and duct-tape, and transported horizontally. In the laboratory cores were split into work and archive halves and again stored horizontally in cold stores (<4°C).

The vegetation surrounding each site was noted, and the exact coring location was obtained using a GPS.

3.3 Lithostratigraphy

A study of the matrix from which pollen and charcoal are to be recovered and interpreted is integral to the evaluation and interpretation of the pollen and CHAR diagrams (Faegri & Iversen, 1989). The main objective of stratigraphy is to subdivide stratigraphical sequences for description, comparison, correlation and ecological interpretation. Correlations within and between cores are concerned with the relationships of events in time, and the temporal equivalence of the events (Birks and Birks, 1980). Once stratigraphic integrity is confirmed or unconformities are identified, influx calculations can be adjusted and interpreted accordingly.

Major sediment boundaries observed in the field during coring were used to record gross stratigraphy. More detailed observations were made in the laboratory during core splitting. Munsell geological colour charts were used (GSA, 2000), and locations of tephra and silt bands were noted. Systematic descriptions of the sediments were made with reference to physical status (colour, elasticity, dryness and stratification).

3.4 Loss on Ignition

3.4.1 Introduction

A commonly used method to estimate the organic and carbonate content of sediments is sequential loss on ignition (LOI) (e.g., Dean, 1974; Bengtsson and Enell, 1986). This method was first evaluated by Dean (1974), who concluded that LOI provides a means of determining carbonate and organic content of clay-poor calcareous sediments and rocks with precision and accuracy comparable to other, more sophisticated geochemical methods. With LOI it is possible to highlight inconsistencies in the sediment record that are not necessarily visible to the eye, such as erosional events, which will impact influx calculations and could explain anomalous dates, and/or shifts in autochthonous production.

The temperature recommended for determining the loss on ignition varies depending on the information required (Sheldrick, 1984). For the purpose of this study, a first reaction oxidises organic matter at 550°C to carbon dioxide and ash (Heiri *et al.*, 2001). In a subsequent reaction, carbon dioxide is evolved from carbonate at 850°C, leaving oxide (Heiri *et al.*, 2001). Assuming a weight of 44g mol^{-1} for carbon dioxide and 60 g mol^{-1} for carbonate (CO_3^{2-}), the weight loss by LOI at 850°C multiplied by 1.36 should theoretically equal the weight of the carbonate in the original sample (Bengtsson and Enell, 1986).

3.4.2 Laboratory Procedure

Samples were weighed and then frozen for 2 hours, and then freeze-dried for 20-24 hours and weighed to 0.01 g accuracy to obtain bulk density values. Samples were then burnt in a muffle furnace at 550°C for 2 hours. The weight loss is approximately equal to the amount of organic matter in the sample (Heiri *et al.*, 2001), although the result obtained is usually somewhat higher than that determined from the amount of carbon present (Sheldrick, 1984). Samples were subsequently ignited at 850°C for 2 hours, and the resulting weight loss includes structural water and carbonates in addition to organic matter (Sheldrick, 1984).

3.5 Magnetic Susceptibility

3.5.1 Introduction

The magnetic properties of minerals can provide significant clues for sediment source identification (Yu and Oldfield, 1989). Magnetizability informs us mainly about iron-bearing minerals found in soils, rocks, dusts and sediments (Dearing, 1999). Magnetic susceptibility is a measure of the response of a sample to the presence of a magnetic field and is the ease with which a sample can be magnetised. It can provide a guide to the concentration of magnetic minerals in a sample and can help identify the processes of their formation or transport (Dearing, 1999). Environmental information such as erosional events around the site can therefore be identified creating an 'environmental fingerprint' for matching stratigraphies, and hence the overlap between the different drives of a core (Thompson and Oldfield, 1986). The measurements have also been found to be diagnostic of specific processes, and have been used in archaeology to identify burning (Dearing, 1991, 1999).

3.5.2 Laboratory Procedure

Cores were initially passed through a Bartington MS2C core scanning susceptibility sensor to determine the overall magnetic susceptibility of each drive, to allow core overlap-matching, which influenced the sub-sampling regime. Subsequently, cores were sub-sampled at 1cm contiguous intervals, and freeze-dried. Each dried sample was then placed in a Bartington MS2B dual frequency susceptibility sensor connected to a Bartington MS2 magnetic susceptibility meter to determine the magnetic susceptibility at high resolution for each level used for other proxy analyses. The results are quoted in S.I. units against volume.

3.6 Pollen Analysis

3.6.1 Introduction

This is the principal technique used to reconstruct Quaternary terrestrial environments (Birks & Birks, 1980). The analysis of pollen preserved in sediments allows the reconstruction of past vegetation environments. Stratified sequences of sediment can provide evidence of variation in pollen deposition over time, which can aid the interpretation of vegetation history and past climate.

Arctic lakes provide specific challenges to pollen preparations largely due to such low concentrations of pollen preserved in the sediment. Non-routine preparation techniques are sometimes necessary, for example the use of flotation liquids (see Faegri & Iversen, 1989).

3.6.2 Laboratory Procedure

The core surface was cleaned in the laboratory prior to sampling to remove contamination and any prevent sampling of modern material moved down core as an artefact of coring. The cores were then sampled at either 5 or 10cm intervals (depending on overall core length) and subsequently 1cm contiguous samples were obtained for high-resolution analysis over key vegetation transitions.

Pollen samples were prepared using the standard methods as outlined in Faegri and Iversen (1989). All samples were spiked with *Lycopodium clavatum* spores to enable determination of pollen influx (Davis, 1966; Stockmarr, 1972; Tipping, 1987). The samples were subjected to a potassium hydroxide treatment for deflocculation, dilute acid and base washes to remove carbonates and humates, hydrofluoric acid to remove silts, and acetolysis to remove cellulose. Samples were exposed to the acetolysis mixture for three minutes (Charman, 1990) to prevent grain deterioration. Pollen slides were mounted in silicone fluid rather than glycerol jelly as pollen grains remain more spheroidal in a fluid media of low viscosity than in a highly viscous or solid media (Praglowski, 1970). Silicone fluid also allows grains to be rotated under the cover slip, permitting a complete assessment of grain morphology (Andersen, 1965), and it does not alter the size of the pollen grain with storage, thus increasing the longevity of the slide (Reitsma, 1969; Faegri *et al.*, 1989, 1975).

3.6.3 Microscope Analysis

All samples were counted on Nikon Optiphot microscopes, to a total pollen sum of at least 300 terrestrial grains (Wright and Patten, 1963). Experimental studies that have assessed the variation in pollen frequency with respect to the total pollen count have shown that frequencies do not vary significantly once a sum of 250 – 300 grains has been reached (Crabtree, 1968).

Pollen grains were identified at x400 magnification. To aid interpretation of certain grains x100 apochromatic oil immersion lenses were used (i.e.x1000). Counting was systematic, traversing the prepared slides at 1mm intervals to avoid duplication, recording all pollen grains and spores encountered (Moore *et al.*, 1991). In most cases the whole slide was counted, if not multiple slides of the same level in order to a) obtain a total count of >300 grains, and b) to eliminate errors caused by the non-random distribution of pollen grains on the slide (Brookes and Thomas, 1963). Pollen identification was made with reference to type slides held in the palynology laboratories at the University of Alaska, Fairbanks, and the PLUS laboratory at the University of Southampton, in addition to the works of Moore *et al.* (1991), Faegri and Iversen (1989), and McAndrews *et al.* (1973). In general the preservation of grains was good, and very few unidentified or degraded grains were recorded.

Betula grains are classified as undifferentiated as the grains were not measured to separate *Betula papyrifera* (tree birch) from *Betula glandulosa/nana* (shrub/dwarf birch), as these species hybridise on the modern landscape. Analyses of modern *Betula* grains demonstrates that grain measurements do not accurately separate these two groups in Alaska (Edwards *et al.*, 1991; Clegg *et al.*, 2005)

3.6.4 Analytical and Presentation methods

The pollen sum was based on the proportions of all tree, shrub and herb data. Pollen and spores of obligate aquatic plants are excluded from the pollen sum as they are locally produced from a different environment that which is under investigation (Birks and Birks, 1980).

Pollen diagrams were calculated and drawn using the TILIA and TILIA GRAPH packages (Grimm, 1991). Pollen concentration values were calculated for each site using the indirect method of proportions of native and exotic pollen using the following formula:

$$\text{Pollen Concentration} = \text{Number of added exotic (grains/ml)} \times \frac{\text{Fossil Pollen Counted (/ml)}}{\text{Exotic Pollen Counted (/ml)}}$$

Determination of pollen influx was achieved with the use of radiocarbon dating (see section 3.9) to calculate the sediment matrix accumulation rate or sediment deposition time (Chapter 5). Pollen concentration was divided by the deposition time to calculate the pollen influx. Pollen assemblage zones were based on changes in dominant tree species and frequencies of occurrence (Gordon and Birks, 1972 cited by Birks and Birks, 1980). Zonation was undertaken using computer generated clustering algorithms (Constrained Incremental Sum of Squares (CONISS) cluster analysis) (Grimm, 1987), which were then visually interpreted to identify the main vegetation zones central to this study.

3.6.5 Black and White Spruce determination

Despite the similar geographic distribution of black (*Picea mariana*) and white (*P. glauca*) spruce, the two species occupy different ecological niches, which means that they potentially have different effects on, or responses to fire regime, so that accurate identification of their pollen is of great importance. The pollen grains of the two species are morphologically similar and are often difficult to distinguish by eye. Several approaches have been used to separate the pollen including total grain size (Wilson and Webster, 1942), morphological criteria (Richard, 1970; Hansen & Engstrom, 1985), linear discriminant analysis using six morphometric variables (Birks & Peglar, 1980) and reciprocal averaging techniques. All of these methods have their respective merits and limitations (see Brubaker *et al.*, 1987), however in general all require extensive time in identification, measurement and calculations before determination is achieved.

In order to resolve the overall pattern of spruce population expansion in this study a quick measurement approach was developed to assign a classification of either probable black or white spruce dominance to each level analysed (Gallagher, 2006).

Measurement of grain characteristics were determined (corpus breadth and saccus heights) for modern grains sampled from Alaska, to determine a suitable single measurement for discrimination (Gallagher, 2006). It was found that corpus breadth could statistically differentiate between the two species. This mean was then applied to determine species composition of fossil samples, although some separation of the size classes may be reduced in the fossil samples due to treatment with HF acid, which can shrink grain size (Andersen, 1960). A minimum of 25 grains per slide which presented in equatorial view without rips or tears were counted. This resulted in many of the grains being outside the preparation counted for the pollen sum.

3.7 Stomatal Analysis

3.7.1 Introduction

A small number of palaeoecological studies have highlighted a discrepancy between Quaternary pollen and macrofossil records from the same, or near-by, sediment cores (e.g. Cwynar and Spear, 1991 and Froyd, 2005). Pine and spruce produce abundant pollen and disperse it widely (Davis et al., 1991) making it particularly difficult to accurately identify their first appearance in the catchment. Conventionally, a 15% frequency threshold has been accepted for the presence of pine in the landscape (MacDonald & Cwynar, 1985). It has recently been shown that even when the pollen percentages are low (potentially <1%), the occurrence of conifer stomata in sediments may indicate the local presence (i.e. presence within the lake catchment area; Bennett, 1983) of a tree in the landscape (Hansen, 1995; Hansen *et al.*, 1996; Froyd, 2005).

As the timing of the transition from spruce to pine dominated vegetation was pivotal to this study due to its potential impact on the fire regime, it was essential to establish the first appearance of pine in the catchment area of the lakes used in this study.

3.7.2 Laboratory Procedure

Pollen samples prepared and mounted as described in section 3.6 above were used for the determination of stomatal presence or absence at each site.

3.7.3 Microscope Analysis

Slides were screened for the presence of stomata by counting grains of the exotic marker (*L. clavatum*) at x20 magnification, and distinguishing pine stomata from other conifer species likely to be present (i.e. spruce). A more detailed analysis equivalent to a 2000-grain pollen count was conducted at all levels through the length of the "pine-tail" and into the preceding spruce zone to accurately identify the first appearance of stomata in the record. Counts ceased at the positive identification of a pine stoma within each level, and the exotic count was recorded, or, if no pine stomata were found, when a total of 2000 exotic grains had been recorded.

3.8 Charcoal Analysis

3.8.1 Introduction

As discussed in the literature review (Chapter 2), a modification of the preparation method put forwards by Rhodes (1998) was designed. Charcoal is a relatively fragile substance, and any method that involved extensive centrifugation or harsh chemical digestion was avoided to reduce the amount of charcoal lost or degraded per sample.

3.8.2 Laboratory Procedure

One ml of sediment was placed in 10% Tetra-sodium pyrophosphate solution for 24 hours to deflocculate the sediment without the need for mechanical separation. Samples were bleached in 6% hydrogen peroxide to discolour any organic material to allow quick identification of charcoal or pyrite particles, which remain unbleached. Samples were passed through 250µ and 125µ nested sieves using distilled water to gently disaggregate any remaining clumps. Material smaller than 125µ was discarded.

3.8.3 Microscope Analysis

The two fractions were then analysed separately under a Nikon SMZ-1000 stereomicroscope to positively identify and tally charcoal particles. Samples were scanned at x20 magnification for locating fragments, with charcoal identification confirmed at x40 magnification. The area of fragments >250µ was determined at x400 magnification. Charcoal fragments contained within the 125-250µ fraction were assumed to be 1 eye piece graticule unit in area at x400 magnification.

3.8.4 Statistical Interpretation and Models

In order to describe the fire regime (fire regime is taken to mean the mean peak frequency or fire return frequency of a record), charcoal accumulation rate (CHAR)/influx was calculated to eliminate the noise produced by uneven sedimentation rates over time. Influx values were then plotted against time. The overall fire regime was determined by distinguishing the background "noise" from "peaks" or fire episodes (a peak is not equivalent to a single fire, or even fire year, as discussed in Chapters 2 and 7, therefore fire episode is used to describe these features).

In order to determine a peak from a background value, a charcoal modelling program (CHAPS (Long *et al.*, 1998) and the Environmental Change Research Group at the Department of Geography, University of Oregon) was utilised to calculate a moving average (window width user defined – Chapter 7). CHAR data were normalised (log-transformed) in order to reduce the effects of extremely high peaks. The threshold value was determined by identifying the area of insensitivity in the number of peaks identified by increasing the threshold further. Each record reached a plateau, where increasing the threshold value no longer influences the total number of peaks recorded (see Chapter 7). The background was then subtracted from the influx data to leave a residual charcoal record and an estimation of peak frequency. A statistically significant change in peak frequency, determined by ANOVA and/or Kolmogorov-Smirnov two-sample tests were used to identify potential alterations in the fire regime.

3.9 Dating Methods

3.9.1 Introduction

In order to establish the sequence of events and the synchronicity of changes in fire regime, vegetation type and climate, and to make realistic comparisons with palaeo-data from other sites, it is essential to establish a reliable geochronology. Radiocarbon dating is the most frequently used method to establish the geochronology of lake sediments dating back as far as the last glaciation, with the most recent or surface samples most commonly being dated using ^{210}Pb, ^{137}Cs, or tephrochronology. In this study it was not possible to utilise ^{210}Pb and ^{137}Cs analysis, due to a lack of sufficient material in the surface sediments. However, the presence of a known and well dated tephra (the WRA) near the top of the Yukon cores was used to confirm chronological integrity near the core surfaces. In the Alaskan sites the period around 10,000 cal yrs BP was of most interest, and therefore dating of the surface sediments was not essential.

3.9.2 *Radiocarbon*

The carbon-14 (^{14}C) isotope (radiocarbon) is formed continuously in the upper atmosphere by the action of cosmic rays (Pilcher, 1991). These atoms combine with oxygen to form carbon dioxide, which enters the Earth's carbon cycle. As ^{14}C is radioactive, there is a continual decay (half life of 5730 years (Stuiver & Polach, 1977)) which is balanced in dynamic equilibrium by its production. The assumption is made that the production rate is constant, thus the amount of ^{14}C in the atmosphere should remain constant (Pilcher, 1991). Also in equilibrium is the carbon dioxide in the biosphere as this is continually exchanged with the atmosphere. Therefore, all living organisms will contain the same proportion of ^{14}C while they remain alive. The equilibrium is lost as soon as an organism dies, as the ^{14}C will continue to decay while exchange with the atmosphere ceases. The rate of decay follows a negative exponential curve and the point at which the amount of remaining ^{14}C falls below the limits of detection provides the upper limit of approximately 50 000 years BP for this dating method (Pilcher, 1991; Libby, 1965).

Radioactivity of a sample can be measured, usually by gas or scintillation counting or by mass spectrometry for accelerated mass spectrometry (AMS) dates. Once the activity is established, the proportion of ^{14}C per sample can be calculated, and the age of the material determined. The age reported is that of radiocarbon years, and is calculated from AD 1950 which equates to 0 BP. The proportion of ^{14}C in the atmosphere since 1950 has been altered significantly by nuclear bomb testing and therefore 1950 is the baseline used for radiocarbon years.

3.9.2.1 *Placement and material used*

Arctic and sub-arctic lakes are notoriously difficult to date due to the lack of macrofossils preserved in the sediment at high latitudes, and the sample size available for dating (Oswald *et al.*, 2005). In addition, carbonate lakes (such as those used in this study) have very limited detrital material due in part to a dilution effect caused by the production of carbonates, which also precluded the use of bulk dating, due to reservoir or hard water effects. Where present, terrestrial macrofossils were identified and picked from the cores to date. Terrestrial macrofossils are the most preferable substrate to date, as they will have been in equilibrium with atmospheric CO_2, and thus are not potentially contaminated with older sources of carbon that could produce error in the age estimation. In the absence of macrofossils at the appropriate levels, charcoal was used, where it was present in sufficient quantities to yield enough carbon for dating. Pollen (spruce or pine) was used in areas where charcoal was insufficient to produce a date, within the spruce and pine zones of each record. Spruce and Pine pollen grains are large enough to be purified from all other fractions in the sediment, to allow a "pure" sample to accurately dated.

Placement of dates was determined by the need for at least one date per thousand years as recommended by PAGES (1994) protocol, to allow calculation of influx at high resolution. Dates were preferentially placed on vegetation transitions as determined by the pollen diagrams in the first instance. In addition a basal (or near-basal) date was obtained for all cores. Subsequent dates were placed in conjunction with shifts in sediment type, influx (determined from previously dated cores) and/or physical sediment properties. Any remaining dates required to produce millennial chronologies were evenly spaced down core.

3.9.2.2 Methodology

Macrofossils were treated with acid-base-acid washes to eliminate contamination absorbed through storage in carbonate water over time, then stored in distilled water prior to AMS dating.

Charcoal samples were sieved out of the sediment, and fragments greater than 250µ were collected to a weight of up to 1g. These samples were also treated with an acid-base-acid wash to remove contamination, and stored in distilled water.

Pollen samples were prepared for dating using a normal pollen separation technique up to and including HF silica removal. Samples were then suspended in heavy liquids to separate out the larger pollen grains (see Brown *et al.*, 1989). In samples with high proportions of the aquatic plant, *Pediastrum*, samples were subjected to multiple short spins (1000 RPM) to separate the pollen from the *Pediastrum*. A minimum of 2500 spruce or pine grains were needed in each sample to produce enough carbon for dating (Brown *et al.*, 1989). Pollen samples were stored in distilled water prior to analysis.

All samples were kept refrigerated prior to analysis at the NERC Radiocarbon Laboratory, UK, or Lawrence Livermore National Laboratory, USA, the distribution between laboratories was a function of sample size (Livermore laboratory is able to process far smaller sample sizes than the NERC laboratory), and the number of dates awarded to the project through applications to the NERC Radiocarbon Steering Committee.

3.9.2.3 Calibration

Reconstructions from past levels of ^{14}C concentrations require radiocarbon ages to be calibrated against, for example, dendrochronological records (Stuvier and Pearson, 1993), as the production of exchange parameters that govern atmospheric ^{14}C levels have changed over time (de Vries, 1958; 1959). These changes constitute a need to calibrate the radiocarbon date to calibrated years BP.

Many programs/methods are available to calibrate dates. This project utilised the on-line radiocarbon calibration software Calib 5.1 (Stuiver & Pearson, 1986-2006). This program incorporates the IntCal04 Northern Hemisphere radiocarbon data set (Reimer *et al.*, 2004). Once calibrated dates were obtained, the weighted average of the 2σ range was determined and used in this project with associated error bars.

3.9.2.4 Age models

Age-depth models were constructed using the WRA, where present, as a constraining point. The WRA has been thoroughly dated in the published literature (Lerbekmo & Campbell, 1969; Lebekmo *et al.*, 1975; Robinson, 2001), and therefore the age-model has to pass through the age range of the tephra. Linear interpolation was used between dates where plausible, and spline interpolation was used where there were significant alterations in sedimentation caused by the ages of radiocarbon determinations, but where no other proxy supported such a change. For full discussion of site-specific age-depth models see Chapter 5).

Chapter 4. Site Descriptions

4.1 Introduction

As detailed in the site selection section of Chapter 3, lakes were chosen that were relatively small and which had closed, or virtually so, basins in order to reflect local burning conditions. In this chapter site-specific characteristics are discussed, and location maps, bathymetry and aerial photographs are presented.

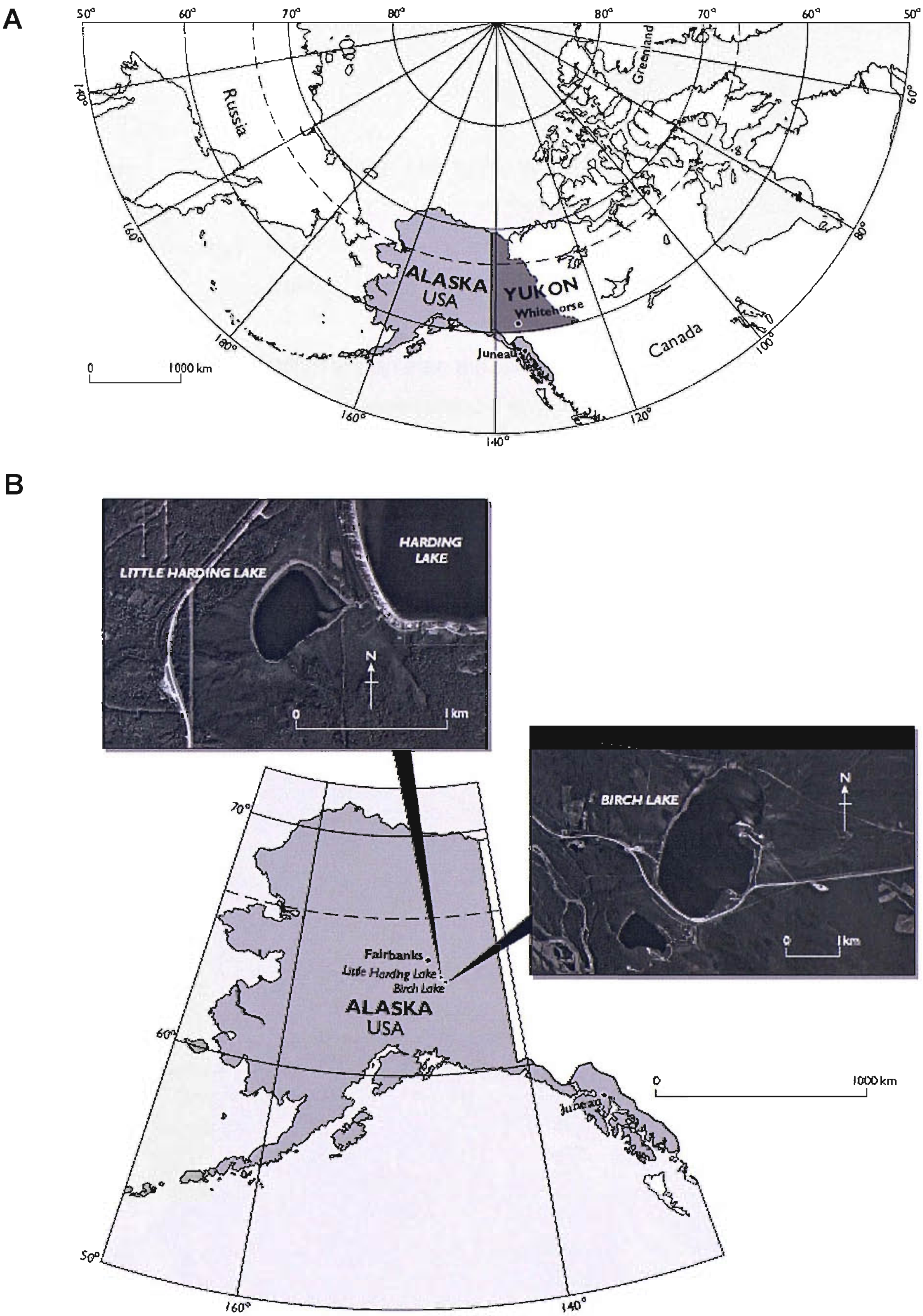


Figure 4.1: Location map of A) Alaska and B) Little Harding and Birch Lakes, Alaska

4.2 *Alaskan Lakes*

4.2.1 Little Harding Lake

Location: 64° 25' 20 N, 146° 51' 10 W

Elevation: 220m a.s.l.

Underlying Geology: late Precambrian to early Paleozoic metamorphic rocks of the Yukon-Tanana Terrain

Approximately 350m in diameter, the lake lies at the mouth of a large east-west trending valley of the Yukon Tanana upland (Figures 4.1 and 4.2). The bathymetry is simple, consisting of an almost circular basin with a broad flat base away from the shore reaching a depth of 8.5m. The region was unglaciated, and basal dates and pollen counts show the record extending back to the late-glacial herb zone (>14,000 cal yr BP; see also Ager and Brubaker, 1985; Anderson et al. 2004). Little Harding Lake most likely formed by aggradation of the Tanana River at the onset of the previous interglacial damming a large low-lying basin as sediment load and discharge increased (Ager, 1975).

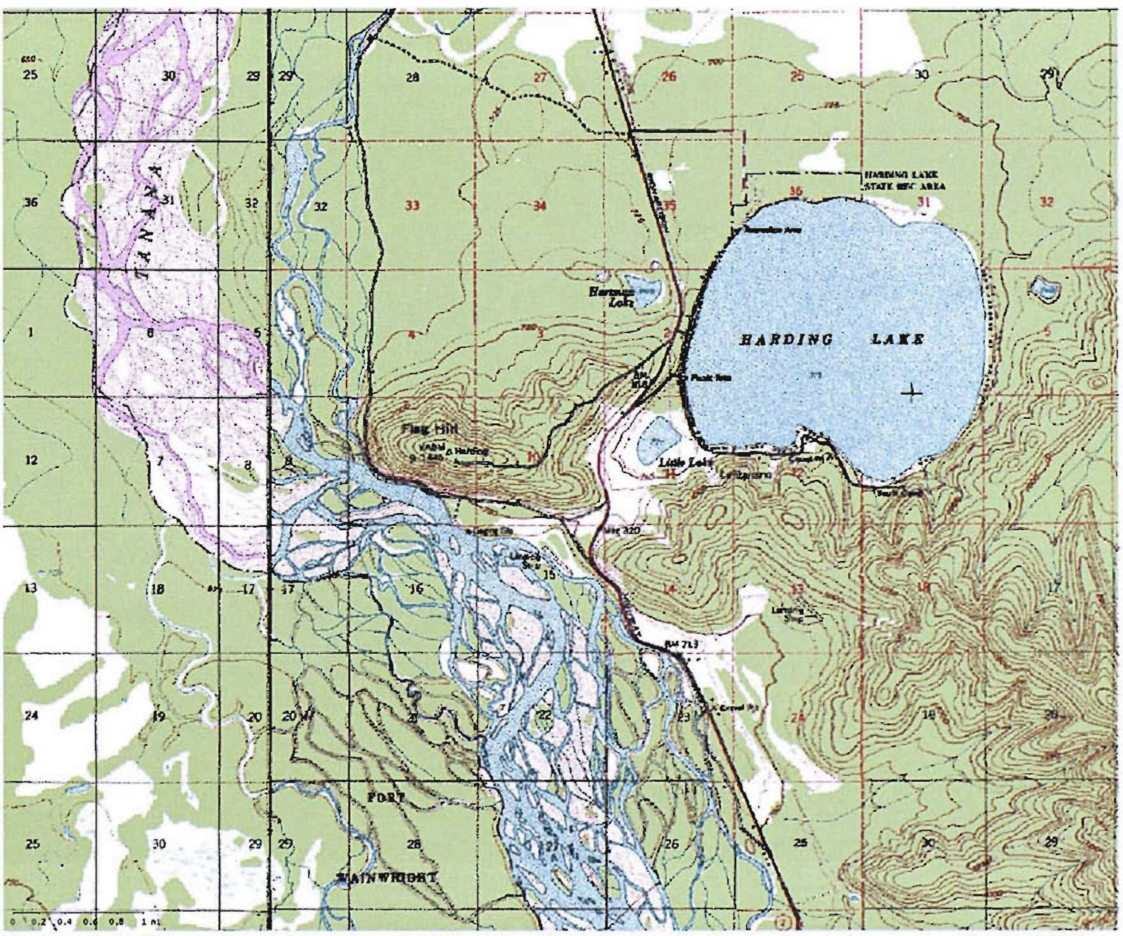


Figure 4.2: Topographic map of Little Harding Lake and surrounding area (Labelled as Little Lake, USGS Sheet: Big Delta B-6 NW, AK)

Plate 4.1: Photograph of Little Harding Lake, showing surrounding terrain and vegetative cover. Photograph: LMFS.

Plate 4.2: Photograph of Little Harding Lake, showing the shoreline sedge mats. Photograph: LMFS

The lake is surrounded by black spruce forest (Plate 4.1), with discontinuous permafrost within the basin. Large areas of sedge mat are present at the shoreline (Plate 4.2). The immediate surrounding terrain is broadly level (Figure 4.3) with increasingly steeper hills and ridges defining the catchment. There are no surface outlets although it is almost certainly hydrologically linked to its neighbour Harding Lake (known as Salchaket Lake by the USGS until 1909). The area surrounding Harding Lake is populated, and the Richardson Highway passes around Little Harding Lake to the west.

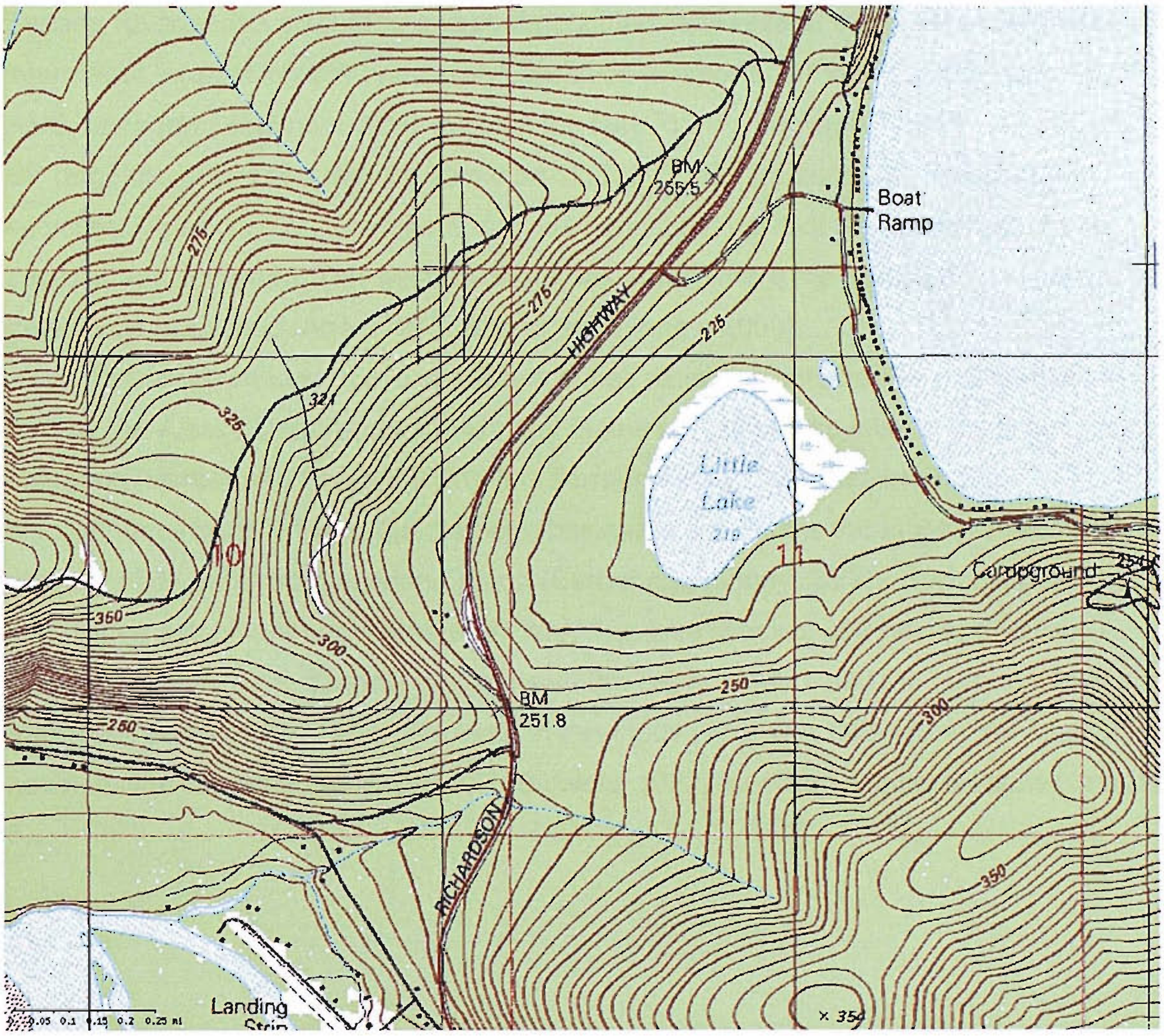


Figure 4.3: Topographic map of Little Harding Lake and surrounding area. (Labelled Little Lake. USGS Sheet Big Delta B-6 NW, AK)

4.2.2 Birch Lake

Location: 64° 18' N, 146° 40' W

Elevation: 850m a.s.l.

Underlying Geology: late Precambrian to early Paleozoic metamorphic rocks of the Yukon-Tanana Terrain

The lake lies within a small east-west trending valley of the Yukon Tanana Upland (Figure 4.5). Similar to Little Harding Lake, it was probably formed by aggradation of the Tanana River as the discharge and sediment load increased at the onset of the previous interglacial, or possibly the Wisconsin Glaciation (Barber and Finney, 2000; Ager, 1975). AMS dates from sediment cores indicate the lake formed between 11,000 – 12,000 ^{14}C BP (Bigelow, 1997) and may be as old as 12,500 ^{14}C BP (radiocarbon dates on bulk sediment (Ager, 1975)). The lake is enclosed on all but parts of the western shore by bedrock and colluvium. The western shore has areas where the dam of outwash sand and gravel was deposited (Ager, 1975; Barber and Finney, 2000).

Today Birch Lake has one surface outlet, which is contained by a concrete weir built by the Alaska Department of Fish and Game to control lake level and prevent fish migration (Barber and Finney, 2000). It is surrounded by boreal forest and has discontinuous permafrost in the drainage basin (Plate 4.3). It is thought to receive the majority of its water from overland runoff (Barber and Finney, 2000). Surface area is ~3.01 km^2, with a catchment basin area of 37 km^2. The lake consists of 2 basins (Figure 4.4). The south basin is the deepest with a maximum depth of 14m, and the north basin has a maximum depth of 12m. The core used in this study (97b) was taken from the south basin, and is correlated with core G/H (Bigelow, 1997) for chronological reasons (Chapter 5).

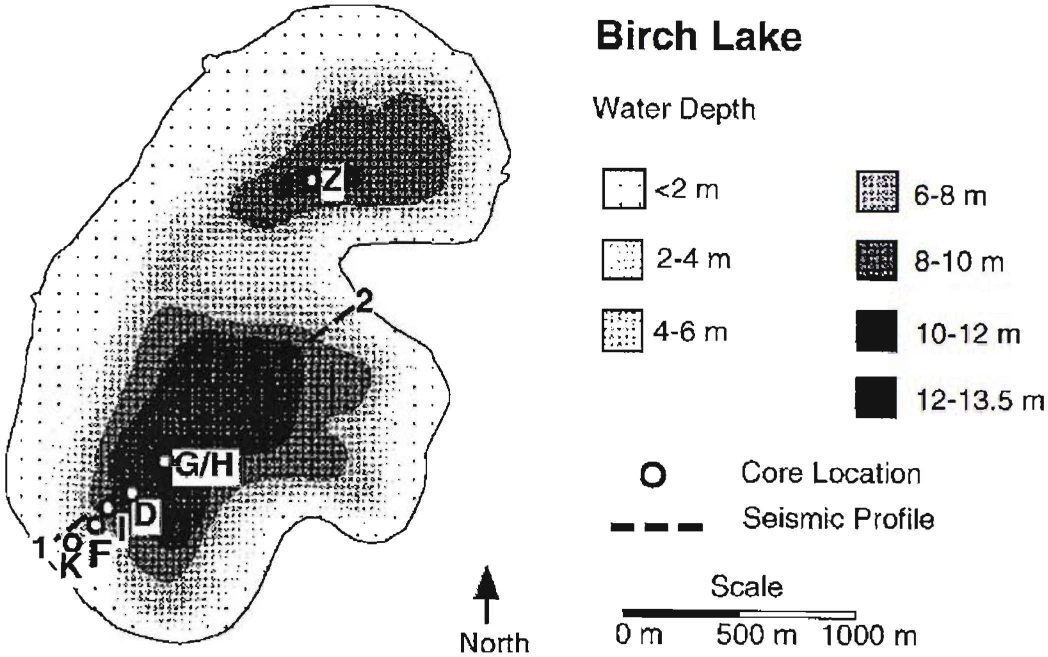


Figure 4.4: Bathymetric map of Birch Lake, Alaska. Source: Bigelow (1997)

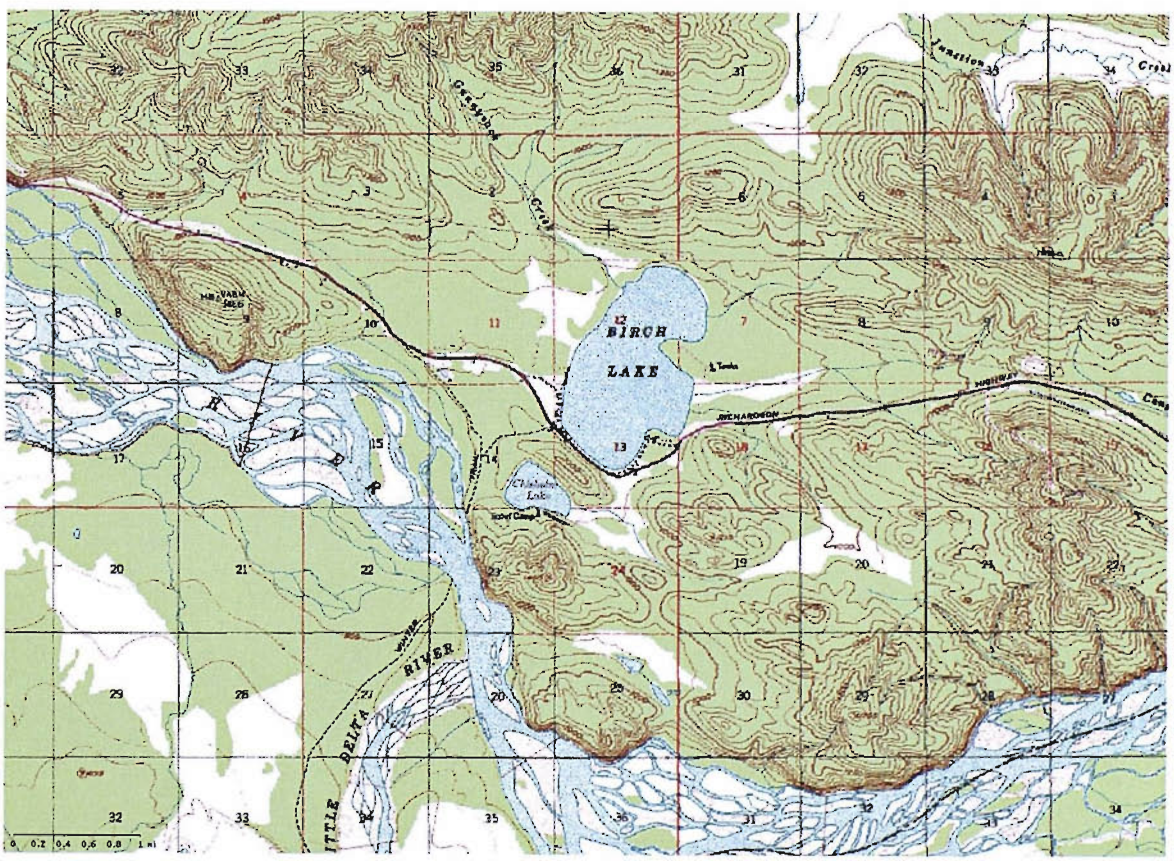


Figure 4.5: Topographic map of Birch Lake, Alaska (USGS Sheet: Big Delta B-6, AK)

Plate 4.3: Photograph of Birch Lake Alaska, south basin showing the surrounding boreal forest vegetation. Photograph: LMFS

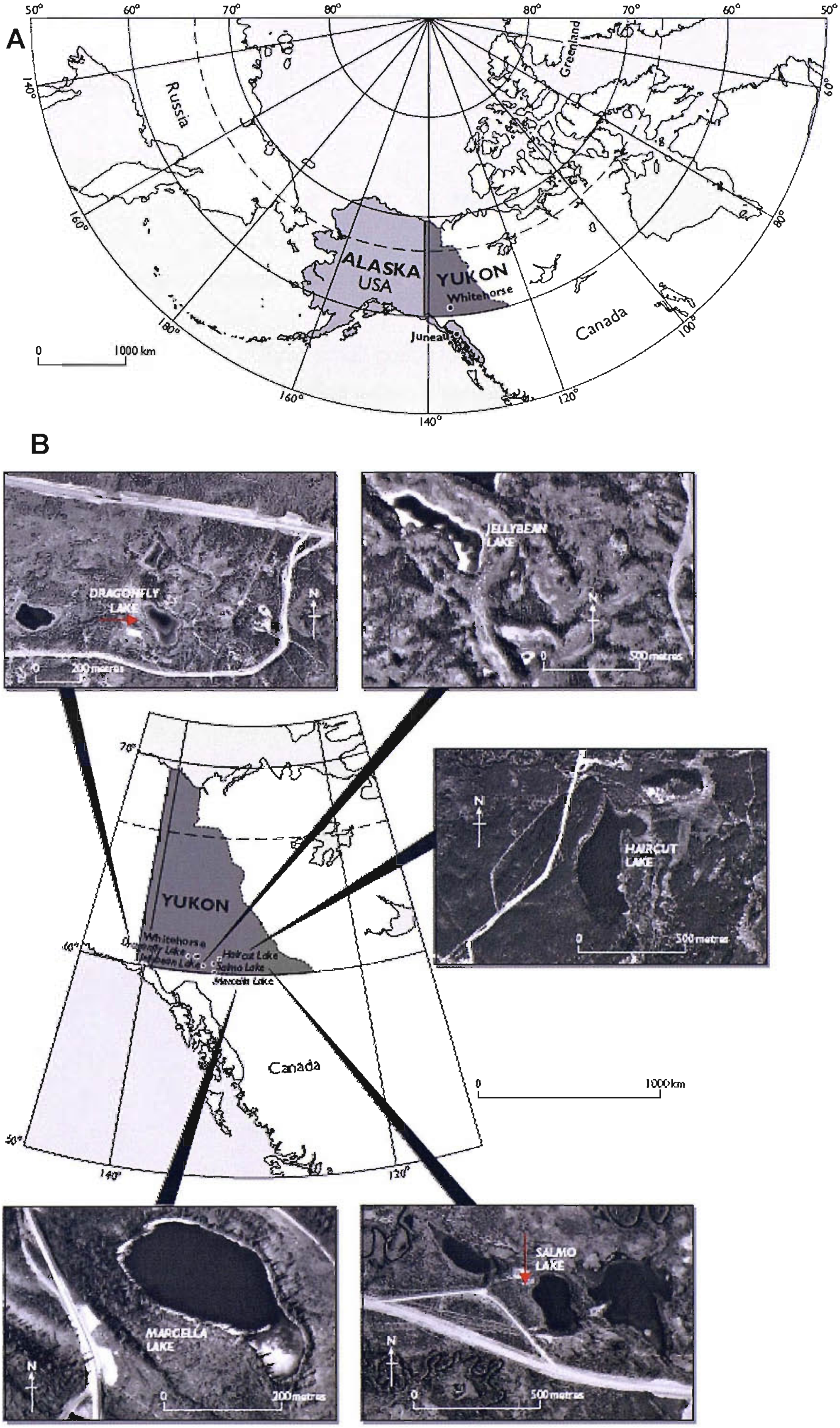


Figure 4.6: Site location map for A) Yukon Territory and B) Yukon Lakes

4.3 Yukon Lakes

4.3.1 Salmo Lake

Location: 60° 26' 44 N, 133° 33' 41 W

Elevation: 866m a.s.l.

Underlying Geology: Laberge group rocks

Salmo lake is one of three small ponds located within dead ice topography (Figure 4.6 and 4.7), in an east-west trending valley. It consists of an almost oval basin, with one steep bank (Plate 4.4), and with a relatively simple bathymetry culminating in a relatively flat bottom at ca. 8m (Figure 4.8). The lithology (Chapter 6) shows an abrupt contact between coarse sand and overlying organic sediments, dated to ~11.000 cal yr BP (Chapter 5), probably marking local deglaciation. This core contains an almost complete Holocene record.

To the north the terrain is sandy and dominated by pine, to the south and west, and immediately around the lake spruce dominates (Plates 4.4 and 4.5). An extensive area of low-lying muskeg (black spruce) is present surrounding the river in the valley to the west.

Plate 4.4: Photograph of Salmo Lake, July 2004 showing surrounding terrain and vegetation. Photograph: LMFS

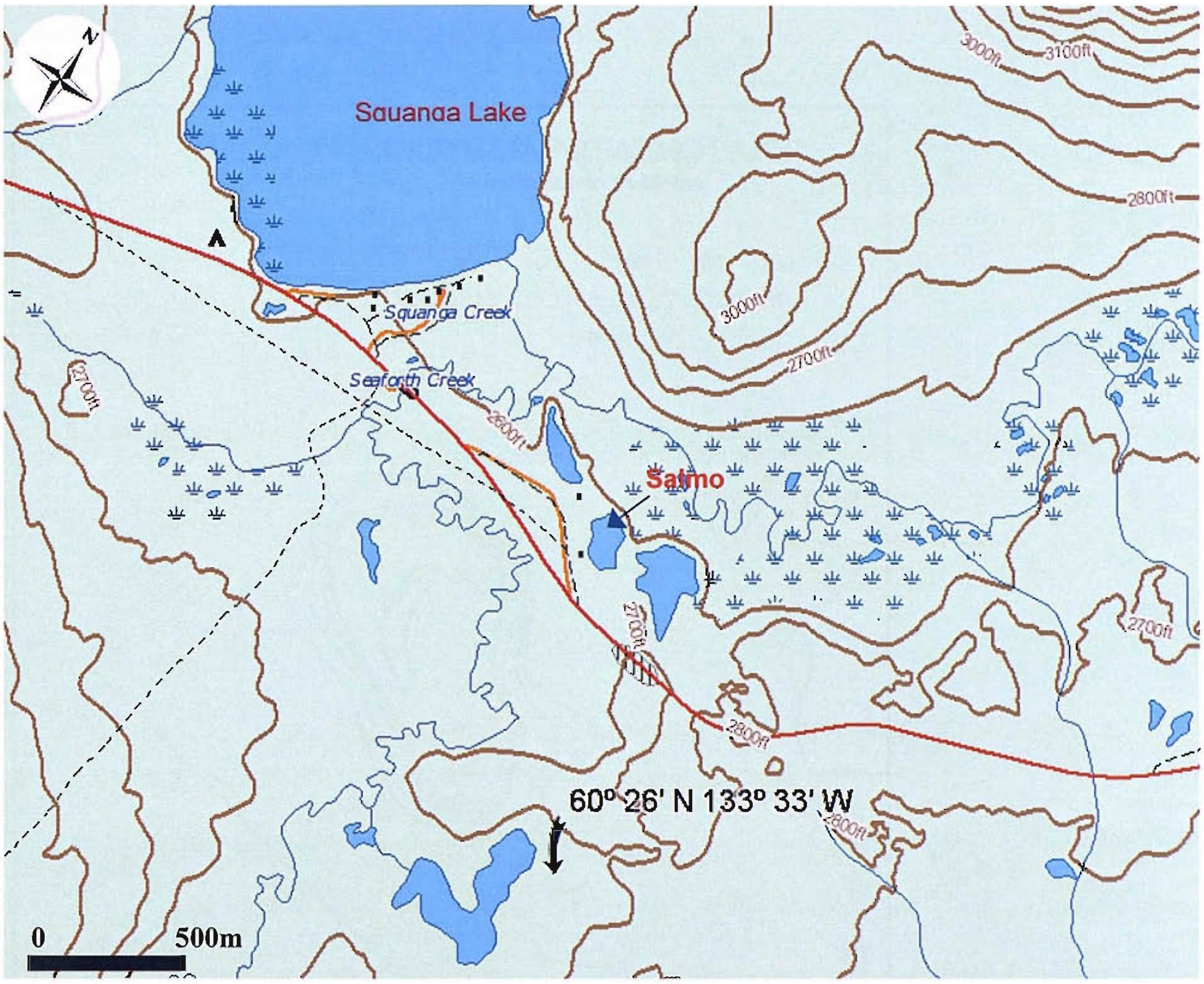


Figure 4.7: Topographic map of Salmo Lake and surrounding area. (Source: Squanga Lake sheet 105 C/5, Energy, Mines and Resources Canada).

Plate 4.5: Photograph of Salmo Lake, and surrounding vegetation. Photograph: LMFS

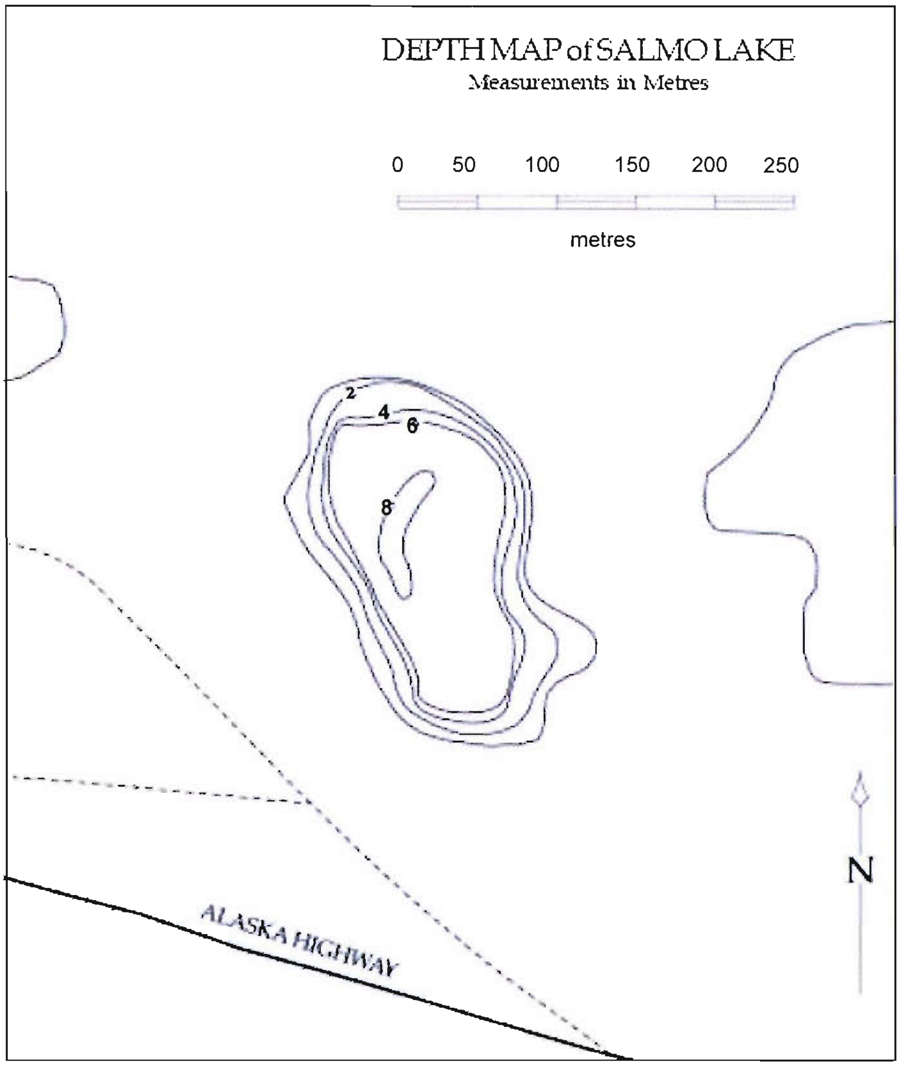


Figure 4.8: Bathymetric map of Salmo Lake, Yukon Territory (Source: Yukon Environment, 2003)

4.3.2 Dragonfly Lake

Location:	60° 48' 43 N, 135° 20' 24 W
Elevation:	760m a.s.l.
Underlying Geology:	rocks of the Whitehorse Trough and some granite

Situated approximately 20km north of Whitehorse, Dragonfly Lake (Figure 4.9) is the most northerly of the Yukon sites. It is a small, shallow pond *ca.* 300x200m in area, containing marl rich sediments. The basin has one steep bank to the south-west (Plate 4.6). The simple bathymetry consists of a small deeper (~2m) area roughly central to the lake, surrounded by a ~1m deep ring. An abrupt onset of organic sedimentation marks local deglaciation at ~10,700 cal yr BP, and the sediments provide an almost complete Holocene record. The surrounding vegetation is mostly dominated by spruce forest. The catchment is known to have partially burned in 1958, as part of the Takhini burn. The pale yellow cloudy water colour suggests high internal productivity.

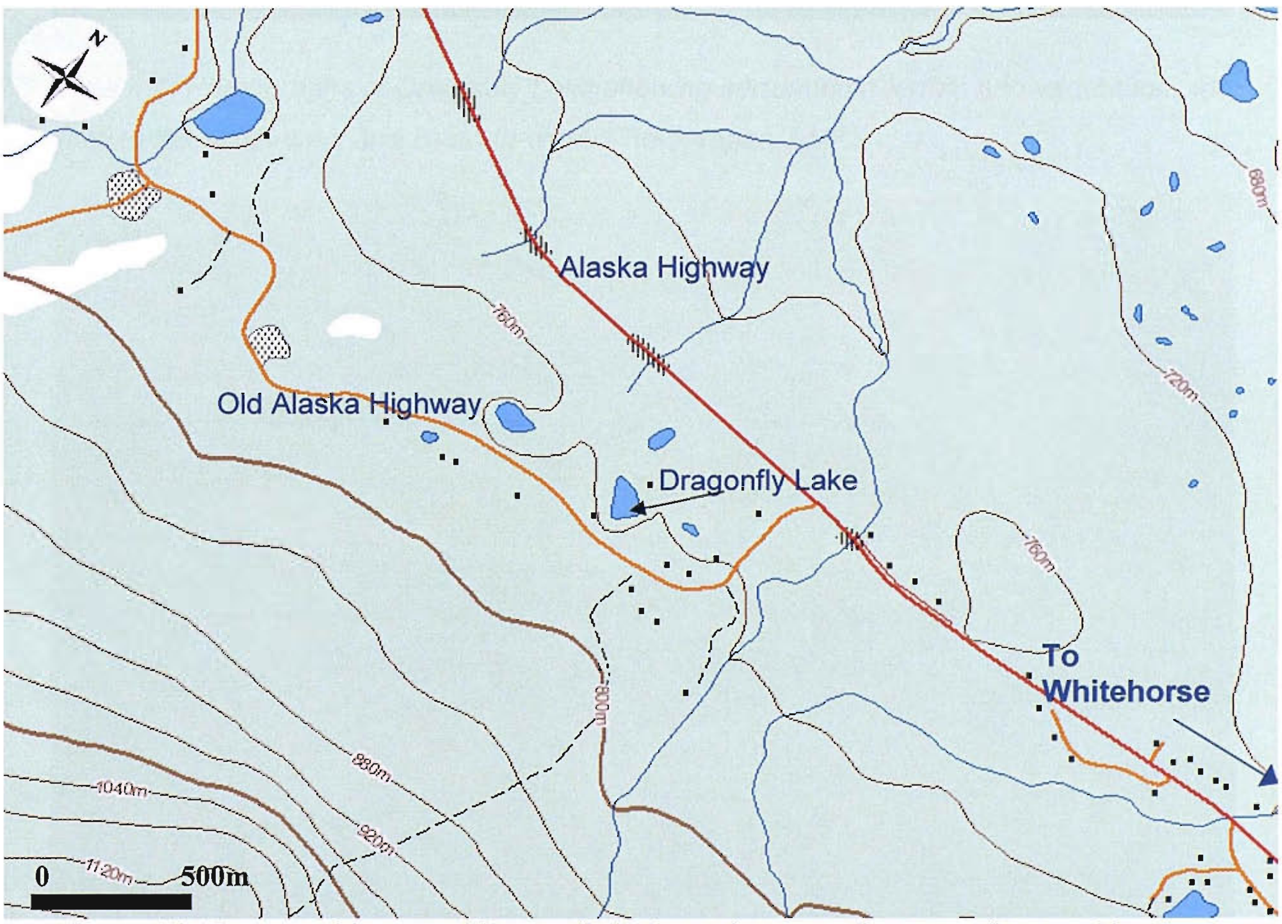


Figure 4.9: Topographic map of Dragonfly Lake and surrounding area. Source: Upper Laberge 105 D/14, Energy, Mines and Resources Canada

Plate 4.6: Photographs of Dragonfly Lake showing surrounding terrain and vegetation. A) view to the south-west and B) north-west. Photographs: LMFS

4.3.3 Haircut Lake

Location: 60° 30' 56 N, 133° 12' 20 W

Elevation: 1000m a.s.l.

Underlying Geology: Laberge group rocks

Haircut Lake is a relatively small lake (~500x200m) set in a basin lying approximately 30m below surrounding terrain (Figure 4.10 and Plate 4.7). The bottom is largely flat at ~8.0m (Figure 4.11), but basin sides shelve steeply. A small inlet enlarges the catchment to the south of the lake, but remains within the pine stand. A small ponded-up outlet is present; however, there was no visible current in July 2004. Core basal date ages the lake to ~7,500 cal yrs BP, indicating the lake may have formed from blocking of the outlet during the Holocene. A full Holocene record is not present at this site; however, the pine rise transition is present.

Steep surrounds are completely dominated by dense pine forest (Plate 4.7), with very little understory species observed (Plate 4.8). Presence of deciduous tree species and shrubs is limited to the shoreline. This is a First Nations (Teslin Tlingit) sacred site.

Plate 4.7: Photograph showing Haircut Lake and the surrounding terrain and vegetation. Photograph: LMFS

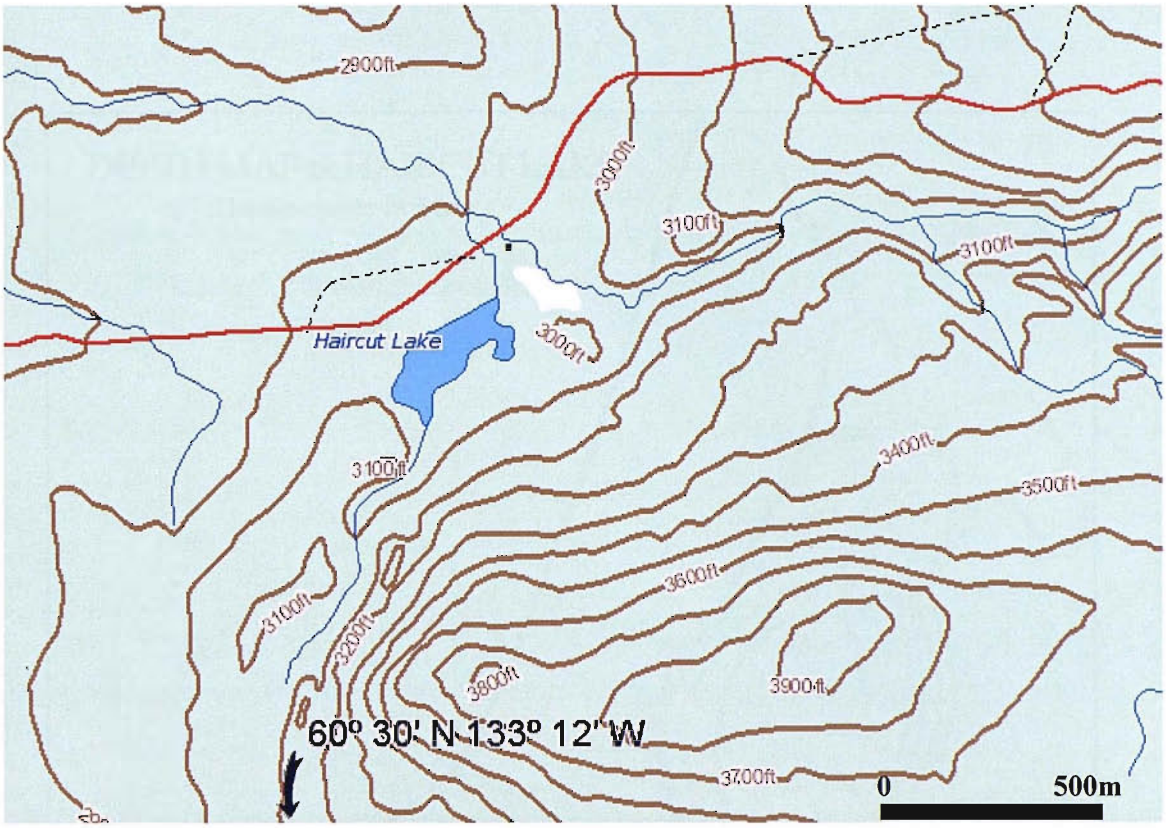


Figure 4.10: Topographic map of Haircut Lake and surrounding area. Source: Mount Grant sheet 105 C/11, Energy Mines and Resources Canada.

Plate 4.8: Pine forest surrounding Haircut Lake. Photograph: LMFS

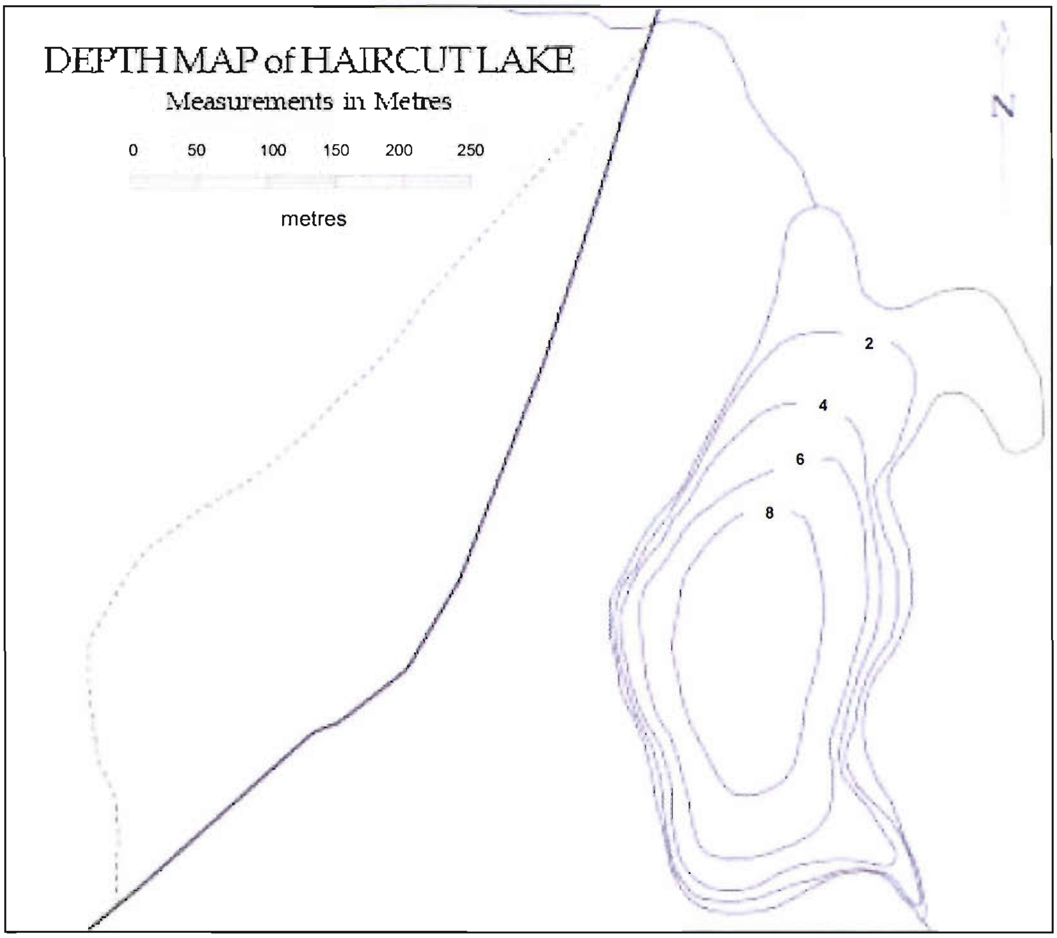


Figure 4.11: Bathymetric map of Haircut Lake, Yukon Territory. Source: Yukon Environment (2003)

4.3.4 Marcella Lake

Location: 60° 4' 25 N, 133° 48' 28 W

Elevation: 697m a.s.l.

Underlying Geology: Upper Paleozoic to middle Mesozoic marine volcanic and sedimentary rocks containing granitic intrusions, on the Yukon Plateau

Marcella Lake is a kettlehole pond located in a northwest to southeast trending depression on a terrace of unconsolidated till and outwash (Figure 4.13). The depression is a former melt-water channel formed during the recession of the northern edge of the Cordilleran Ice Sheet. The lake is approximately ~0.4km^2 in area (Anderson, 2004), with a simple bathymetry (Figure 4.12), the deepest point of 9.7m located in the middle of the basin (Anderson *et al.*, 2005b). It lies ~20m below the surrounding terrain with a well defined 0.8km^2 catchment (Anderson, 2004 and Plate 4.9).

Plate 4.9: Photograph of Marcella Lake and surrounding terrain and topography. The marl bench can be seen in the far left of the photograph. Photograph: LMFS

The catchment is dominated by lodgepole pine and trembling aspen, with sage (*Artemesia*) and grasses growing on well-drained south facing slopes. Spruce trees can be found at the shoreline. This site has been studied previously in an on-going US-Canadian collaborative study of climate change using sediment-based proxies and chironomids. A pollen record and ^{14}C chronology are available, but the surface section was unsuitable for charcoal work, and was re-taken in 2004.

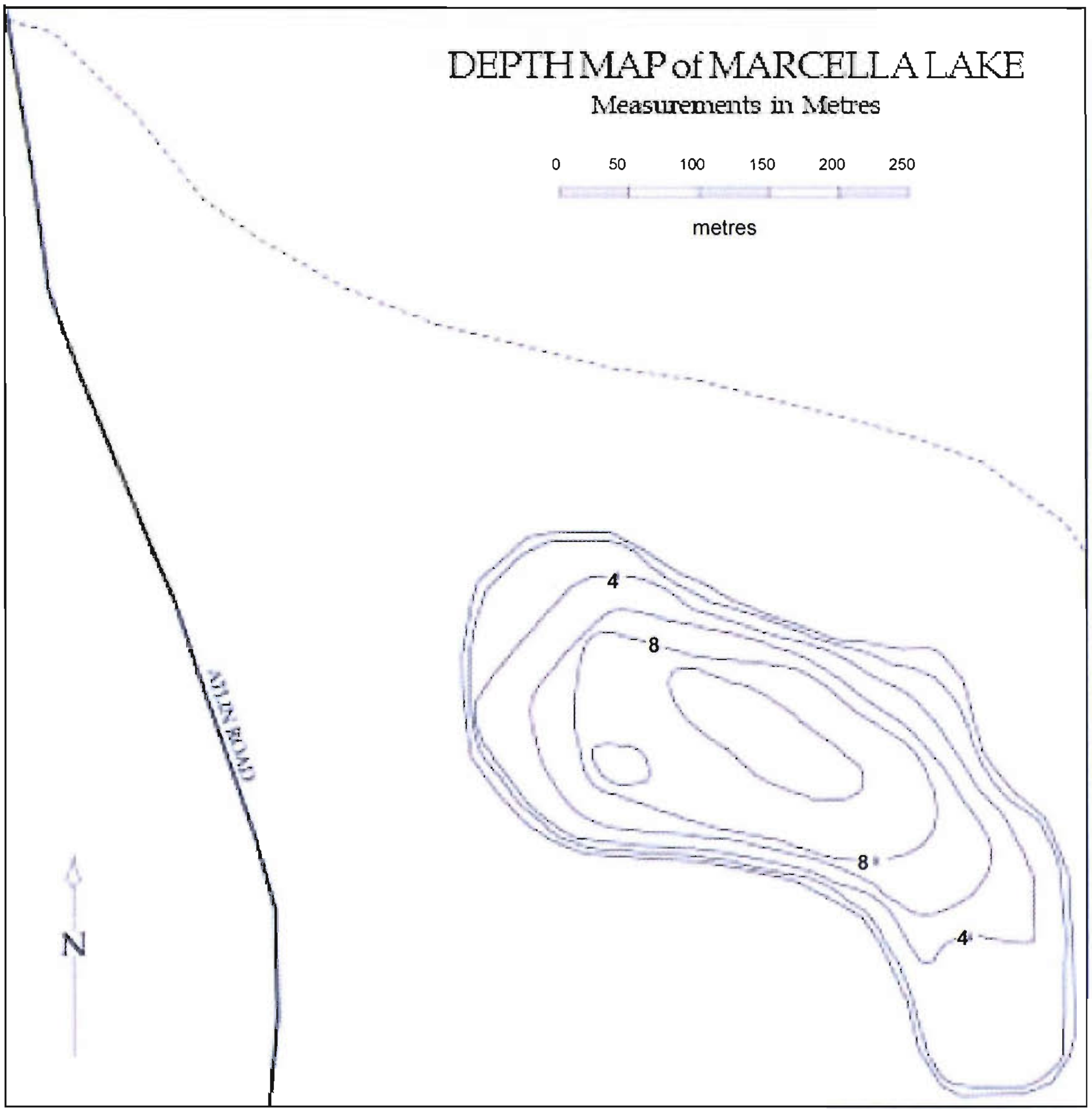


Figure 4.12: Bathymetry map of Marcella Lake, Yukon Territory (Source: Yukon Environment, 2003)

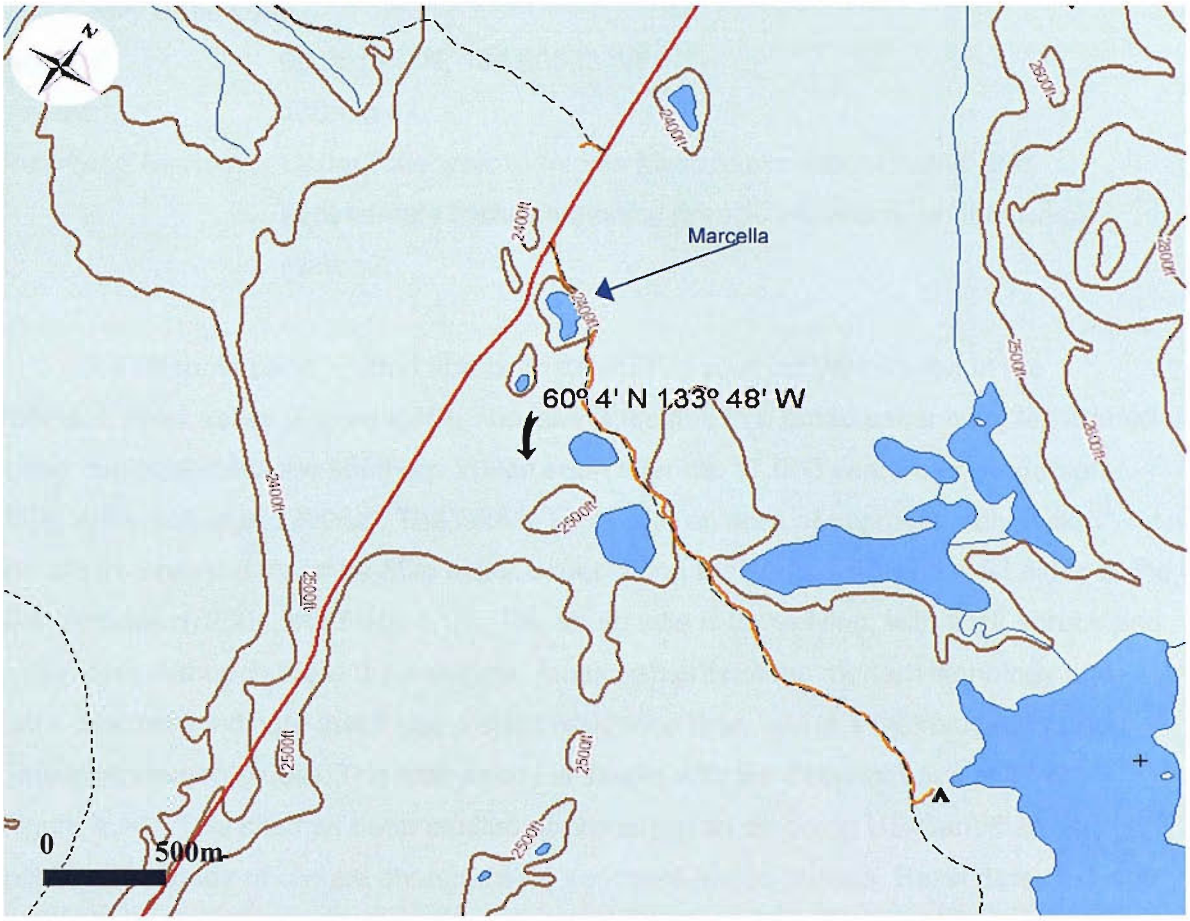


Figure 4.13: Topographic map for Marcella Lake showing surrounding area. Source: Lubbock River, sheet 105 C/4, Energy, Mines and Resources Canada.

4.3.5 Jelly Bean Lake

Location: 60.35145 °N, 134.80563 °W

Elevation: 730m a.s.l.

Underlying Geology: Upper Paleozoic to middle Mesozoic marine volcanic and sedimentary rocks containing granitic intrusions, on the Yukon Plateau

A kettlehole pond located approximately 50km south of Whitehorse in the Robinson River valley (Figure 4.15). The lake is located in a broad esker complex formed during deglaciation of the southern Yukon soon after ca. 17,000 years ago (Anderson, 2004; Anderson *et al.*, 2005a). The lake is small with an area of approximately 0.4km^2, and sits in a basin defined by 50m esker ridges along the north, east and west sides of the lake (Anderson, 2004 and Plate 4.10). The south side is lower lying, with thick spruce and birch forest. Although there is no outflow, measurements of the modern limnology and water chemistry indicate that it has a short residence time, and is a hydrologically open system (Anderson, 2004). The bathymetry is simple with the deepest point at 11.6m (Figure 4.14). This site has been studied previously in an on-going US-Canadian collaborative study of climate change using sediment-based proxies. Basal dates (~7,400 cal yr BP) indicate that a full Holocene record is not present.

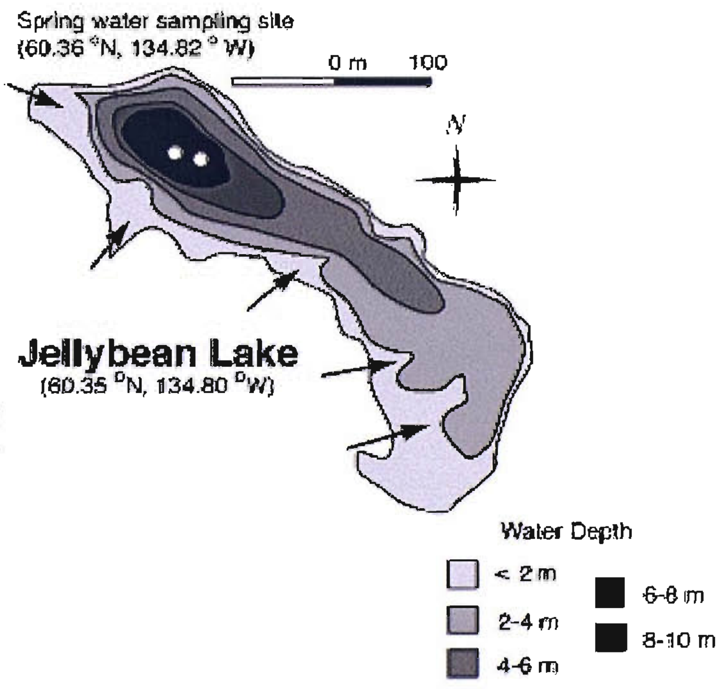


Figure 4.14: Bathymetry map of Jelly Bean Lake, Yukon Territory (source: Anderson et al, 2005

Plate 4.10: Photograph of Jellybean Lake and surrounding terrain and vegetation. Photograph: LA

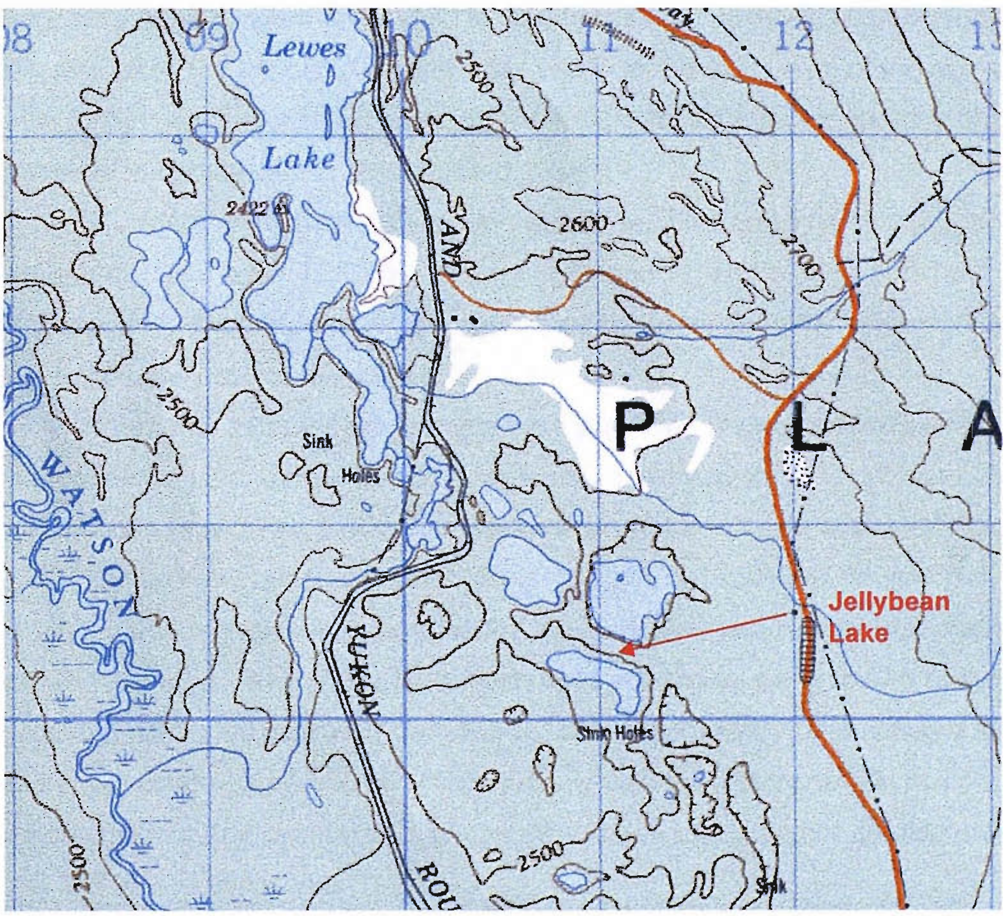


Figure 4.15: Topographic map of Jellybean Lake and surrounding area. Source: Robinson sheet 105 D/7, Energy Mines and Resources Canada.

Chapter 5. Chronology

5.1 Radiocarbon Dating

Samples were dated at the NERC Radiocarbon Laboratory (NERC) (East Kilbride), the Lawrence Livermore Radiocarbon Laboratory (LLNL) (Washington, USA) and Poznan Radiocarbon Laboratory (Poz) (Poland). Samples for radiocarbon dating consisted of macrofossils (picked from the cores) where possible, and concentrated *Picea/Pinus* pollen or charcoal samples (core samples of 1cm vertical thickness processed for either pollen or charcoal) where necessary. The total carbon was recovered as CO_2 by heating the sample with CuO in a sealed quartz tube, and then converted to graphite by Fe/Zn reduction. The graphite was then mounted on a pellet and the ^{14}C determined at the respective laboratories using 5MV National Electrostatic Corporation AMS systems.

The samples sent to the NERC Radiocarbon Laboratory were awarded to the project through applications to the bi-annual Radiocarbon Steering Committee's competition. Due to internal NERC laboratory policies and procedures, not all of the samples submitted were processed in time for use in this thesis. Samples that were received after 1st October 2006 were omitted from the evaluation of the age-depth models described below, unless they were significantly different from the existing trend, in order to complete the analysis and discussion of results.

Samples were submitted independently to LLNL and Poz laboratories, funded through the NSF and LTER palaeohydrology projects, and other grants held by Dr. M.E. Edwards, to further inform the age-depth models, allow definition of sampling resolution and to complete the chronologies.

Individual samples are identified in the text via the publication codes assigned by the respective laboratories (SUERC for NERC determinations, CAMS for LLNL determinations and Poz for Poznan determinations) and depth where relevant.

Dates have been calibrated using the Calib 5.0 software (Stuiver & Reimer, 1986) available on-line, which incorporates the IntCal04 Northern Hemisphere radiocarbon data set (Reimer *et al.*, 2004). The calibrated dates are shown using the 2σ range, and are quoted in calibrated years BP denoted as "cal yrs BP". Both the measured radiocarbon and the calibrated ages are reported, but only calibrated radiocarbon ages are used for the discussion. Age estimates of uncalibrated ages used in other studies are denoted as "^{14}C BP" where applicable.

The age-depth models utilise the weighted average of each calibrated radiocarbon date, with the 2σ range error estimate shown. Telford *et al., (*2004*)* concluded this to be the most accurate method. At sites where chronologies have been determined and published by colleagues before the completion of this project, established age-depth models are reported, and are not under the control of the author.

Linear interpolation has been used to form the age-depth models for most sites. Although this type of model assumes that the sedimentation rate alters abruptly at the depth of the dates, the majority of the dates have been placed on lithological and/or vegetational shifts, and interpolation was determined as the most appropriate method.

At some sites it has been necessary to apply a spline interpolation due to the effects of the WRA tephra on the age-depth relationship. A spline is a polynomial fitted between pairs of points, but whose coefficients are determined slightly non-locally, i.e.: information is used from other points than the pair under immediate consideration. This non-locality is intended to make the fitted curve smooth overall, and not change gradient abruptly at each data point. This method takes no account of the errors on the radiocarbon ages, and therefore has been limited to short sections of the relevant cores covering the surface samples only.

Other types of models were tested for comparison. Polynomial models did not improve the chronologies, producing models which were too stiff and unresponsive to the data when low orders were applied, and creating reversals at the extremes of the chronology due to the shape of the functions when high orders were used. Linear regression produces a global fit to the data, with changes at one end of the core affecting the relationship fitted to the whole stratigraphy, and was therefore rejected as a method to produce age-depth relationships.

5.2 Tephra

The White River Ash (WRA) is a relatively recent tephra with a bi-lobate distribution, originating from the St Elias Mountain Range, which covered the southern half of the Yukon Territory, easternmost Alaska and part of the Northwest Territories with pyroclastic deposits (Lerbekmo & Campbell, 1969; Lerbekmo *et al.*, 1975). The deposit consists of a northern and an eastern lobe which have axis lengths of noticeable deposits in excess of 500km and 1000km respectively (Lerbekmo *et al.*, 1975). The sites located in the Yukon Territory contain a tephra layer identified as the eastern lobe of the WRA, which has the reported age range of between 1390 yrs BP (+/- 70) (Stuiver *et al.*, 1964) to 1200 yrs BP (+/- 140) (Lowdon & Blake, 1968) with a generally accepted date of 1250 BP (+/- 110) (Lerbekmo *et al.*, 1975) calibrated and used in this study as 1150 cal yrs BP.

5.3 Age Depth Models

5.3.1 Introduction

In this section site-specific chronologies will be presented, and the age-depth relationship discussed. Each site was considered separately, with a range of age-depth relationships explored before a final age-depth model was constructed, in order to assign the most appropriate model, taking site-specific characteristics and sedimentation rates into account. As outlined above, for sites which have been previously studied the existing/published age-depth models are reported.

5.3.2 Little Harding Lake, Alaska

Twelve samples were analysed for radiocarbon determinations on Little Harding Lake core A (Table 5.1). The age-depth relationship is presented in Figure 5.1 (a, b) for the entire core and a section of the core containing the analytical window (the section of the core fully analysed for the project, see section 6.2 for reasoning behind this approach). The basal sample dates from *c.*14 cal yrs BP (CAMS 111153). The penultimate radiocarbon determination (CAMS 120921) was located within an area of oxidised gyttja of questionable integrity (see Section 6, Figure 6.1), and returned a date which is within the errors of CAMS 111153, ergo the same date. It is probable that this sample consisted of reworked material, and as such only the basal date is used to inform the age-depth relationship.

At the time of writing, four radiocarbon determinations are outstanding at the NERC facility. In addition to informing the age-depth model, the four dates form two pairs of charcoal and pollen concentrates as a separate experiment to constrain dating errors. Dating of pollen concentrates is now a well established method (Brown *et al.*, 1989; Long *et al.*, 1992; Mensing & Southon, 1999; Regnell & Everitt, 1996; Vandergoes & Prior, 2003). However, in the absence of large spruce or pine grains, pollen samples become unreliable as a dating medium as it is not possible to separate smaller grains from aquatic material. It is possible to date macro-charcoal, and as the project's goals rest on the assumption that charcoal pieces are contemporaneous with sediments, we proposed a calibration study on paired charcoal-spruce pollen samples to NERC. It is expected that the samples should yield very similar (within error) results, with the probability of the charcoal sample being slightly older, due to the combustion of established trees in the landscape.

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age 2σ range (years BP)
CAMS-111151	876-876	Pollen (*P. glauca* and *P. mariana* concentrates)	1220 +/- 45	1072-1180 (86.5%) 1209-1229 (13.5%)
CAMS-111152	1090-1091	Pollen (*P. glauca* and *P. mariana* concentrates)	4180 +/- 50	4628-4637 (5.1%) 4642-4680 (23.5%) 4688-4762 (50%) 4790-4792 (1%) 4797-4830 (20.3%)
SUERC-7835	1152-1153.5	Pollen (*P. glauca* and *P. mariana* concentrates)	6203 +/- 78	7001-7177 (87.9%) 7212-7242 (12.1%)
SUERC-	1190-1191	Charcoal Fragments		
SUERC-	1190-1191	Pollen (*P. glauca* and *P. mariana* concentrates)		
CAMS-120919	1235-1237	Pollen (*P. glauca* and *P. mariana* concentrates)	8790 +/-60	9680-9917 (95.8%) 10096-10112 (4.2%)
SUERC-	1259-1260	Charcoal Fragments		
SUERC-	1259-1260	Pollen (*P. glauca* and *P. mariana* concentrates)		
CAMS-129020	1273-1274	Leaf and wood fragments	8900 +/- 130	9787-9848 (13.1%) 9865-9876 (2.1%) 9886-10197 (84.8%)
SUERC-7024	1313-1314	Wood fragment	10222 +/- 70	11777-11780 (1%) 11812-12075 (99%)
CAMS-120921	1350-1351	Charred vegetation and seeds	12300 +/- 90	13949-14746 (100%)
CAMS-111153	1386-1387		12260 +/- 100	13859-14682 (100%)

Table 5.1: Radiocarbon determinations for Little Harding Lake, Alaska, Core Z. Samples highlighted in red are still in the NERC processing system at the time of writing. The sample shaded grey is not used to inform the age-depth model due to suspected reworking of material within the sediment.

The sediment accumulation rate appears to be relatively constant down core (Figure 5.1a) with an apparent alteration below 1235cm (mid-way through the analytical window, Figure 5.1b). The main sediment type of organic brown gyttja is interrupted by a temporary appearance of red banding (see Figures 6.1 and 6.2), and it is probable that the sedimentation rate altered as a consequence through this period. The sediment returns to gyttja below 1310cm, and the accumulation rate again appears to be steady to the core base. The outstanding date at 1259cm should inform the model more thoroughly as it is positioned at the transition in sediment types, and will constrain the inflection in the model.

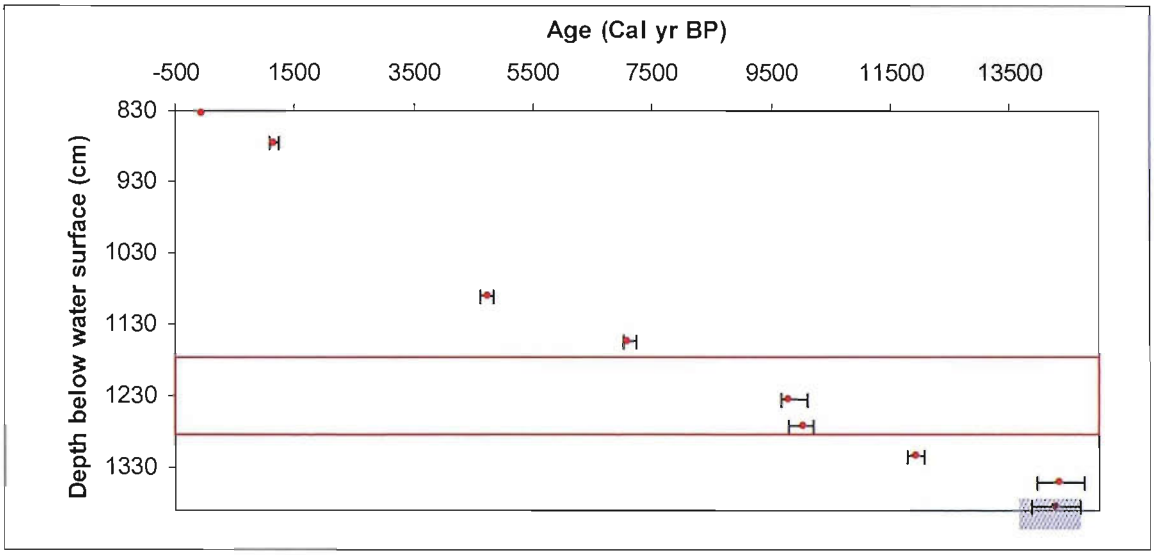


Figure 5.1a: Age-depth relationship for Little Harding Lake, core Z, using nine radiocarbon dates, with the surface marked as modern (red dot with no error bars). The grey shaded box denotes the anomalous date, and is not used to inform the age-depth model. The area demarcated by the red box represents the analytical window.

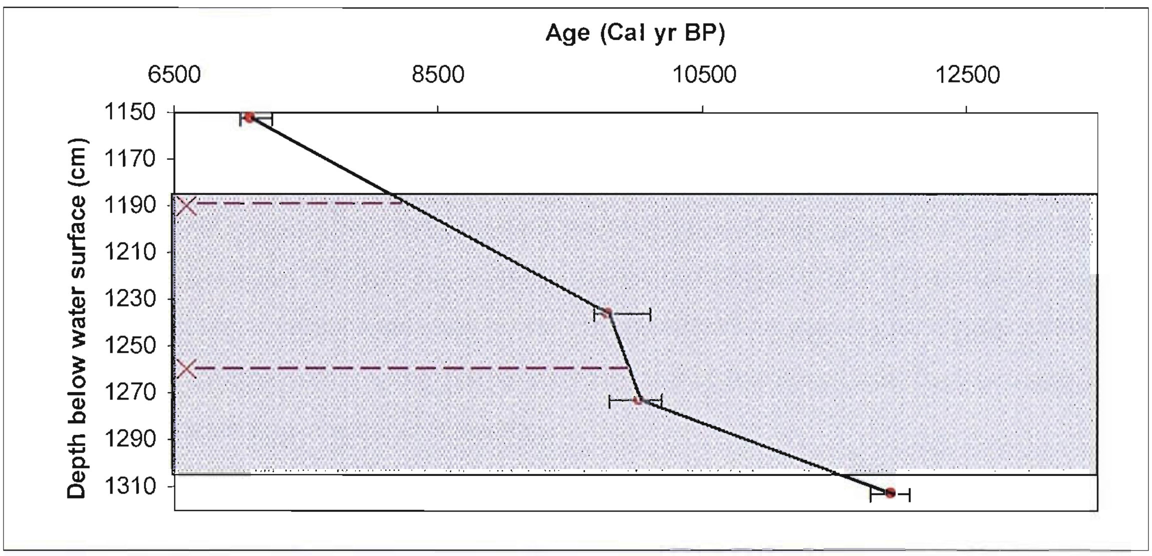


Figure 5.1b: Age-depth relationship for Little Harding Lake, core Z, covering the analytical window (grey shaded area). The trend line denotes linear interpolation between stratigraphically adjacent dates. Dates still outstanding at the NERC radiocarbon lab are denoted by an X and the dashed line.

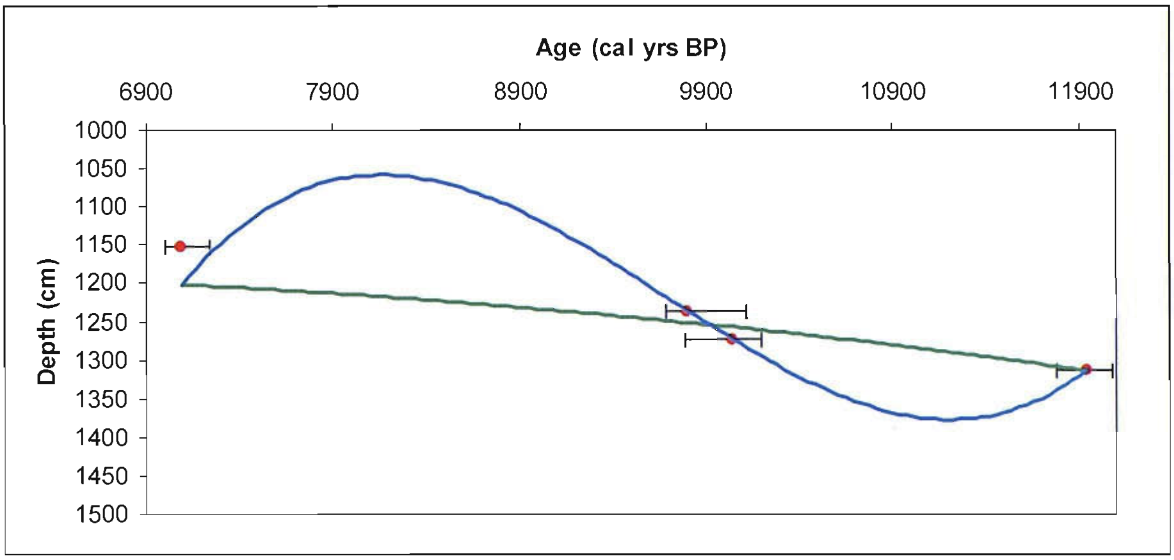


Figure 5.2 the analytical window of Little Harding Lake, core Z, polynomials used to describe the age-depth relationship. The green line represents a second order polynomial, and the blue line a third order polynomial. Higher order polynomials did not alter the trend line further from that of a third order polynomial.

It is unfortunate that the uncertainty in the model occurs close to the transition to spruce dominated vegetation (see Figure 6.3), however, in the absence of the date at 1259cm, the most parsimonious model to fit through the data is that of linear interpolation. This will create two sudden changes in accumulation rate, for which there is some evidence in the core stratigraphy (see Section 6.2.1). It would be preferable to use a polynomial with the current chronology in order to smooth out the implied sedimentation changes due to their potential impact on the CHAR calculation. However, as can be seen in Figure 5.2, the use of a second order polynomial provides an age-depth model that runs through very few of the 2σ error bars on the calibrated dates, and a third order or higher polynomial creates artificial reversals in the chronology. The age-depth model will be reassessed when the outstanding dates are returned in the event that they provide an altered or adjusted age-depth relationship.

Using linear interpolation, the analytical window shows an initial accumulation rate of 32.5 yrs/cm (0.03 cm yr^{-1}) to 1236cm.The accumulation rate increases to 6.57 yrs/cm (or 0.15cm yr^{-1}) in the mid section (1236-1273cm), slowing to 47.5 yrs/cm (0.021 cm yr^{-1}) by the end of the analytical window.

5.3.3 *Birch Lake, Alaska*

Four radiocarbon determinations are represented in Table 5.2a for core 97B. As for Little Harding Lake, core Z, a window of analyses was produced to cover the spruce rise, and dates were obtained to bracket this event. The lowest determination (CAMS-124460) is anomalous and is thought to be a result of contamination of the sample either during sub-sampling of the core, or during the radiocarbon determination process. Therefore the sample at 1687-1688cm was submitted as a near-basal date to inform the age-depth model (Figure 5.3), however, the laboratory had not processed the date in time for inclusion in this thesis.

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age 2σ range (years BP)
CAMS-124458	1584-1586	5 *Betula* seeds and a bract	6000 +/- 60	6677-6707 (3%) 6713-6987 (96.9%)
CAMS-124459	1646-1648	Leaf fragments	8320 +/- 190	8662-8666 (0.1%) 8704-8709 (0.1%) 8716-9683 (99.8%)
Poz	1687-1688	Charcoal		
CAMS-124460	1701-1703	Charcoal and seed fragments	6580 +- 160	7164-7755 (100%)

Table 5.2a: Radiocarbon determinations for Birch Lake, Alaska, Core 97B. The date shaded in grey is not used to inform the age-depth model, and is thought to be anomalous due to contamination. The date highlighted in red is awaiting processing at the laboratory at time of writing.

In the absence of a sufficiently robust chronology for this section of the core, correlation with core G/H (see Chapter 4) was conducted in order to estimate the basal age of the analytical window. The cores were correlated using the magnetic susceptibility and LOI curves (see Section 6.2.2 for data and Appendix 2 for correlation), and adjusting the depth of radiocarbon determinations secured on core GH (Table 5.2b) by +18cm to correspond with the appropriate depth in 97b, to provide an amalgamated chronology (97b-GH).

Publication Code	Depth in core G/H (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age 2σ range (years BP)
CAMS-56307	1415	Plant macrofossil	7740 +/- 140	8318-8994 (100%)
CAMS-25427	1633	*Picea* pollen	8480 +/- 60	9327-9343 (1.5%) 9402-9548 (98.5%)
CAMS-25423	1691	Plant macrofossil	9210 +/- 50	9524-11268 (100%)

Table 5.2b: Radiocarbon determinations for Birch Lake, Alaska, Core GH (data source: Bigelow, 1997, with permission). The depths were adjusted by +18cm to correlate with core 97b.

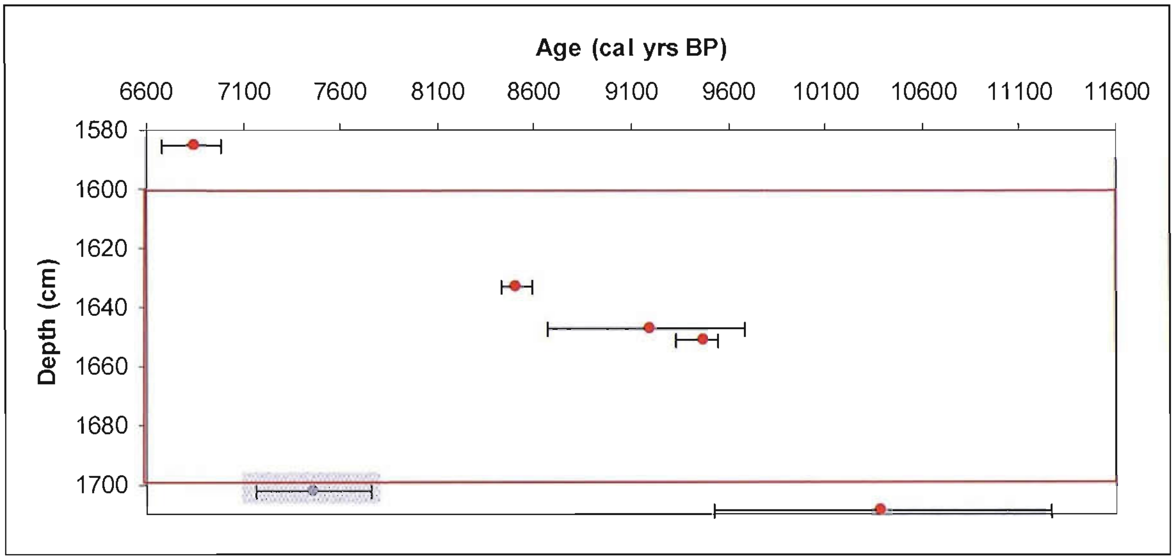


Figure 5.3: Age-depth relationship using two radiocarbon determinations for Birch Lake, core 97B, and three determinations from Birch Lake, core GH. The anomalous date is shaded in grey and was not used to inform the age-depth model. The red box denotes the analytical window.

Using this amalgamated chronology the analytical window is dated to *ca.* 10300 cal yrs BP. A second order polynomial was considered appropriate for the age-depth model due to the visible trend displayed by the determinations, and for consistency with Bigelow's (1997) interpretation. The sediment accumulation rate can be seen to gradually decrease up-core from ~14 yrs/cm (or 0.07 cm/yr) at the base of the analytical window to ~42 yrs/cm (or 0.24cm/yr) at the top.

Once the sample at 1687-1688cm core 97b has a radiocarbon determination, a 97b chronology will be constructed extrapolating from 1688cm to the end of the analytical window at 1700cm. It is unfortunate that extrapolation is necessary; however, due to the age of the core, and previous analyses carried out on it, the amount of material available to be dated was extremely small, resulting in a sub-optimal sampling strategy for the chronology.

5.3.4 *Salmo Lake, Yukon Territory*

Seven radiocarbon determinations were made on Salmo Lake, core A, represented in Table 5.3, with the age-depth relationship presented in Figure 5.4. Unfortunately there was a handling error at the NERC laboratory at the time one sample was processed (846-846.5cm), and a determination was not possible (highlighted in red in Table 5.3).

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age 2σ range (years BP)
	846-846.5	Charcoal	Lab did not	secure a date
CAMS 120913	862-863	Pollen (*Picea glauca, P. mariana, and Pinus contorta* concentrates)	2750 +/- 35	2789 – 2869 (100%)
SUERC-11736	905-905.5	Charcoal	3579 +/- 39	3725-3752 (4.1%) 3761-3795 (5.8%) 3820-3982 (90.1%)
SUERC-7206	941-943	Pollen (*P. glauca* and *P. mariana* concentrates)	4810 +/- 31	5473-5551 (72.8%) 5573-5602 (27.1%)
SUERC-11735	960.5-961	Charcoal and charred needle tip	4691 +/- 40	5317-5482 (84.0%) 5529-5579 (16.0%)
Poz-16782	1000	Pollen (*P. glauca* and *P. mariana* concentrates)	7030 +/- 50	7745-7959 (100%)
SUERC-7023	1024-1024.5	Wood macrofossil	8090 +/- 49	8778-8834 (7.5%) 8858-8924 (6.9%) 8928-9135 (82.8%) 9177-9202 (1.4%) 9222-9241 (1.2%)
CAMS 120914	1048-1048.5	Leaves and wood fragments	10090 +/- 100	11403-11584 (37.5%) 11586-11825 (56.9%) 11924-11952 (5.6%)

Table 5.3: Radiocarbon determinations for Salmo Lake, Yukon Territory, core A. The sample that did not secure a determination is highlighted in red. Dates considered anomalous and excluded from the age-depth model are shaded in grey.

The age-depth relationship (Figure 5.4) shows a relatively stable sediment accumulation rate down core, from the surface to 1024cm. Sample SUERC-11735 is not used to inform the model due to the younger than expected determination. This sample contained high levels of sulphur compared to the other samples processed, indicating a probable difference in the origin of the sample.

The chronology includes the WRA tephra, which is present in the core as a discrete 2cm thick layer, and therefore any age-depth model must pass within the errors of the tephra date to reasonably represent the true age-depth relationship. Due to the apparently constant accumulation rate displayed by the radiocarbon determinations from the surface to the penultimate date, linear interpolation was used to produce the age-depth model. The basal date indicates an increase in sediment accumulation towards the bottom of the core. This is thought to be indicative of the processes involved in lake formation (abrupt onset of organic and carbonate production, and transition from gravels to gyttja at 1045cm) and is supported by the LOI and magnetic susceptibility results (see section 6.3.1). A determination on this sediment transition was not possible due to lack of datable material. Therefore it is probable that the inflection in the age-depth model is offset by ~3cm due to CAMS 120914 at 1048cm. This determination causes an abrupt alteration in sedimentation rate at the base of the core compared to the rest of the chronology. It is probable that with the exception of the off-set, this is a fair estimation of the transitional lake environment, with the main core chronology extending to 1045cm (transition from organic gyttja to gravels), where there is a slowing of accumulation rate within the gravel section to the core base.

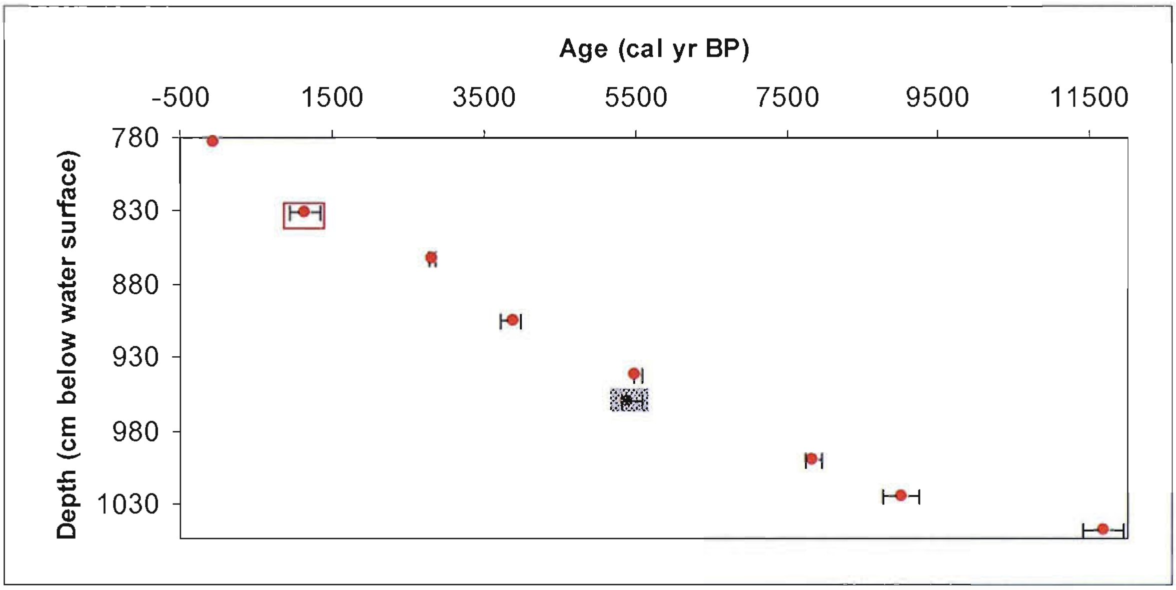


Figure 5.4: Age-depth relationship for Salmo Lake, core A, incorporating seven radiocarbon determinations, the WRA tephra (bordered with a red box) and the surface (red dot without error bars). An anomalous determination is shaded grey.

The base of the core is dated from *c.* 11,000 cal yrs BP, and shows an initial sedimentation rate of 111.3 yrs/cm (or 0.009 cm yr^{-1}), which increases above 1024cm to 49.3 yrs/cm (or 0.02 cm yr^{-1}). The sediment accumulation rate gradually decreases up core, reaching 27.2 yrs/cm (or 0.037 cm yr^{-1}) at the surface.

5.3.5 *Dragonfly Lake, Yukon Territory*

Nine radiocarbon determinations were conducted on Dragonfly Lake, core A (Table 5.4), and the age-depth relationship is represented in Figure 5.5.

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age (years BP +/-2σ)
SUERC-11406	240-241	Pollen (*Picea glauca*, *P. mariana*, and *Pinus contorta* concentrates)	606 +/- 35	543-655 (100%)
CAMS-120915	280-282	Pollen (*Picea glauca*, *P. mariana*, and *Pinus contorta* concentrates)	1155 +/- 35	987-1032 (38.2%) 1051-1091 (38.8%) 1107-1135 (20.1%) 1162-1167 (3%)
Poz-16785	365-366	Pollen (*P. glauca* and *P. mariana* concentrates)	2870 +/- 30	2879-2912 (6.3%) 2918-3078 (92.6%) 3097-3102 (0.6%) 3132-3136 (0.5%)
SUERC-11404	426-427	Pollen (*P. glauca* and *P. mariana* concentrates)	4050 +/- 35	4422-4625 (93.7%) 4763-4788 (6.3%)
SUERC-11403	456-457	Pollen (*P. glauca* and *P. mariana* concentrates)	4529 +/- 35	5049-5194 (64.6%) 5212-5311 (35.4%)
SUERC-7828	523-525	Charcoal and wood fragments	5816 +/- 88	6502-6695 (92.1%) 6701-6719 (7.9%)
Poz-16783	566-567	Pollen (*P. glauca* and *P. mariana* concentrates)	6950 +/- 40	7685-7860 (96.7%) 7903-7920 (3.3%)
SUERC-7829	627-628	Charcoal, wood fragments and *Carex* seeds	8294 +/- 147	9124-9463 (100%)
CAMS-120916	674-675	Wood and leaf fragments	9480 +/- 90	10586-10798 (68.1%) 10850-10864 (3.2%) 10956-11009 (14.4%) 11017-11067 (14.3%)

Table 5.4: Radiocarbon determinations for Dragonfly Lake, core A. All determinations are used to inform the age-depth model

The radiocarbon determinations produce an age-depth relationship that is near linear in nature (Figure 5.5). Due to the seemingly constant accumulation rate, linear interpolation was used to produce an age-depth model.

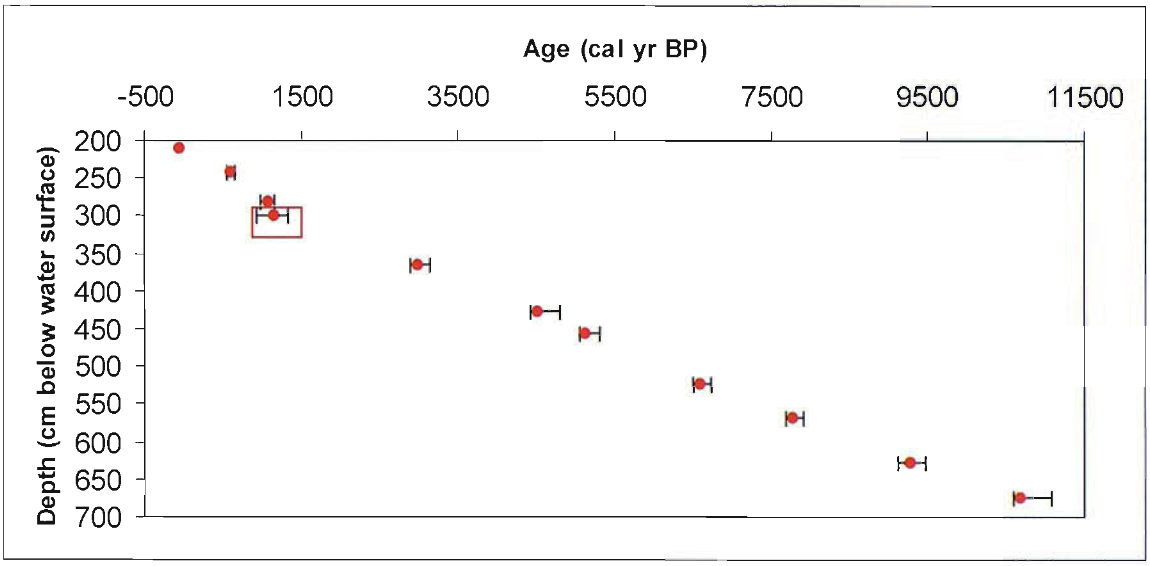


Figure 5.5a: Age-depth relationship for Dragonfly Lake, core A, based on nine radiocarbon determinations, the WRA tephra (demarcated by the red box), and the surface (red dot with no error bars).

The basal sample dates the core from *c.* 10900 cal yrs BP, with an initial accumulation rate of 29.76 yrs/cm (or 0.034 cm yr^{-1}), staying relatively constant up core to the WRA where accumulation rate is calculated at 28.44 yrs/cm (or 0.035 cm yr^{-1}). Sediment accumulation appears to increase after the WRA starting at 4.11 yrs/cm (or 0.24 cm yr^{-1}), which is probably a function of less compacted surface sediments, as the tephra is relatively close to the top of the core, aswell as an abrupt onset of carbonate production around this time (see section 6.3.2). From the WRA the sedimentation rate gradually decreases to 22.14 yrs/cm (or 0.05 cm yr^{-1}) at the core surface.

Concern over the effect of the abrupt alteration in sedimentation rate near the WRA on the CHAR calculations prompted the use of a spline interpolation to produce an alternate age-depth model (Figure 5.5b). Ultimately it was discovered that there was little difference reflected in the charcoal record between the two age-depth models (see Chapter 7). It is probable that the radiocarbon determination CAMS120915 should not be considered as a separate date to the WRA due to the nearly identical age estimate, and the errors associated with it.

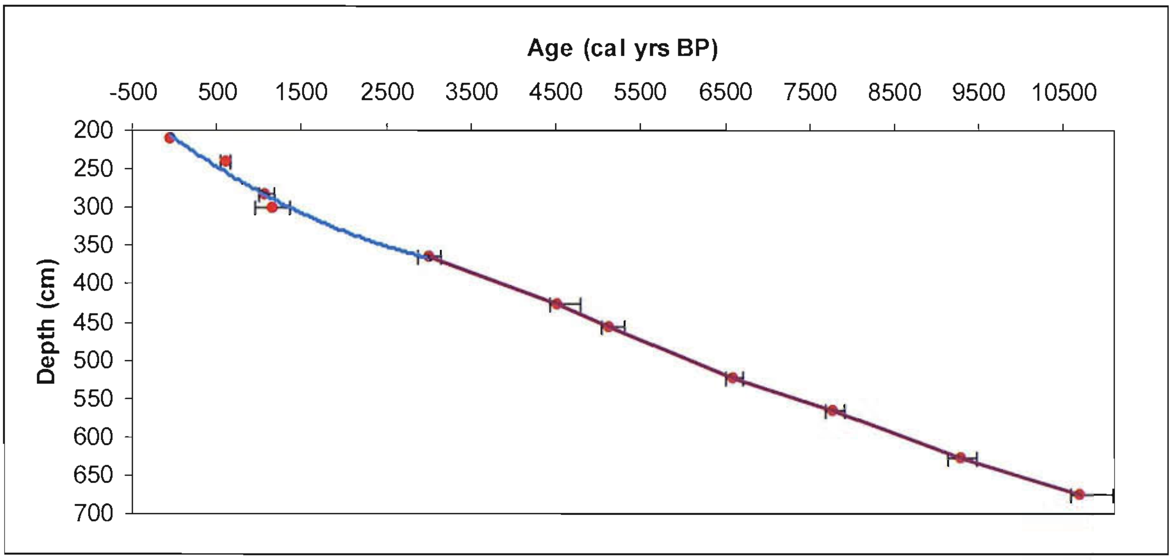


Figure 5.5b: Age-depth relationship for Dragonfly Lake, core A, based on nine radiocarbon determinations, the WRA tephra (demarcated by the red box), and the surface (red dot with no error bars). A spline interpolation was applied to the top 5 ages (blue line) and linear interpolation was applied to the remaining 6 ages (purple line).

5.3.6 *Haircut Lake, Yukon Territory*

Eight radiocarbon determinations were conducted on Haircut Lake, core A (Table 5.5), and the age-depth relationship is presented in Figure 5.6.

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age (years BP +/-2σ)
CAMS-120917	879-881	Pollen (*Picea glauca*, *P. mariana*, *and Pinus contorta* concentrates)	610 +/- 40	553-569 (20%) 582-612 (39.4%) 620-650 (40.7%)
SUERC-11412	904-905	Pollen (*Picea glauca*, *P. mariana*, *and Pinus contorta* concentrates)	1952 +/- 35	1825-1952 (91.9%) 1958-1987 (8.1%)
CAMS-120918	921-923	Pollen (*Picea glauca*, *P. mariana*, *and Pinus contorta* concentrates)	3940 +/- 70	4258-4217 (4.5%) 4287-4445 (79.6%) 4474-4477 (0.7%) 4481-4514 (15.1%)
SUERC-11409	942-944	Pollen (*Picea glauca*, *P. mariana*, *and Pinus contorta* concentrates)	3122 +/- 35	3257-3408 (97.3%) 3426-3441 (2.7%)
SUERC-7831	956.5-957.5	Charcoal	3659 +/- 52	3846-4095 (95.9%) 4118-4145 (4.1%)
SUERC-	980-981	Charcoal		
SUERC-11407	990-991	Pollen (*Picea glauca* and *P. mariana*, concentrates)	5581 +/- 35	6297-6413 (97.8%) 6425-6436 (2.2%)
SUERC-7834	1012-1013	Charcoal	6935 +/- 113	7585-7916 (100%)

Table 5.5: Radiocarbon determinations undertaken on Haircut Lake, Core A. Sample highlighted in grey is suspected to be contaminated by older sources of carbon and is excluded from consideration of the age-depth relationship. The sample highlighted in red is still in the NERC processing system at the time of writing.

The radiocarbon date CAMS-120918 (921cm) is much older than the radiocarbon dates either side, and is thought to be anomalous. There are no stratigraphical or physical sediment properties that would suggest a reversal or mixing of the sediment (see section 6.3.3), and the anomaly most likely results from contamination of the sample with old carbon via the presence of *Pediastrum* nets, for example. For these reasons the date has not been used to inform the age-depth model (Figure 5.6). The determination for the sample 980-981cm (highlighted in red) is still being processed by the NERC facility at the time of writing.

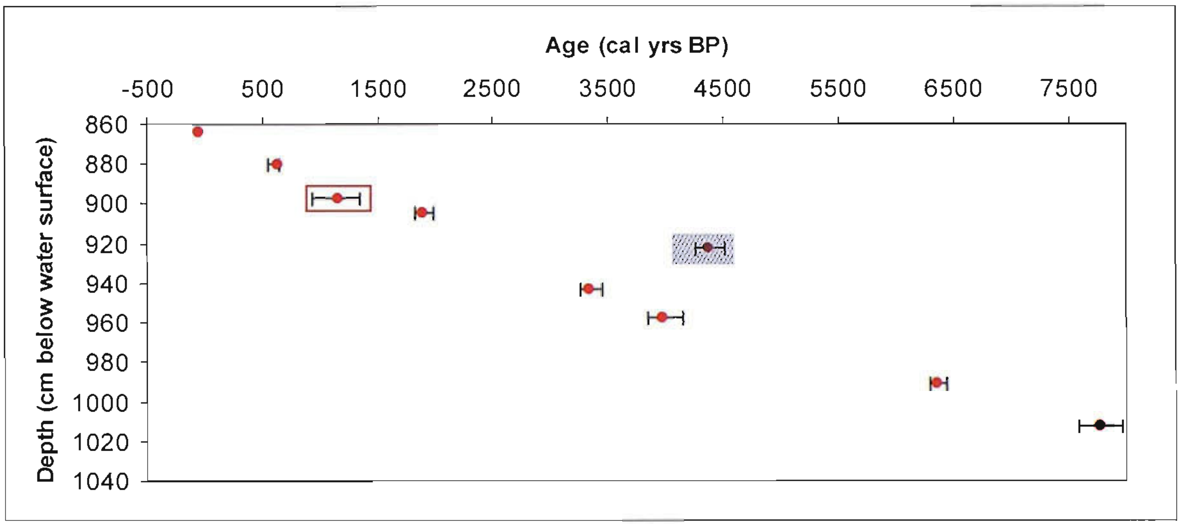


Figure 5.6: Age-depth relationship for Haircut Lake, core A, using six radiocarbon determinations, the WRA tephra (demarcated by the red box), and the surface (red dot without error bars). The anomalous date is demarcated with the grey box, and does not inform the age-depth model.

In general there is a near linear age-depth relationship shown by the radiocarbon determinations on this core, and linear interpolation was used to produce an age-depth model for Haircut Lake. The basal determination dates the core to *c.* 8000 cal yrs BP, with an initial accumulation rate of 64.46 yrs/cm (or 0.016 cm yr^{-1}). The sedimentation rate increases to 37.51 yrs/cm (or 0.027 cm yr^{-1}) from 957cm to 904cm. There is a slight inflection in the curve around the WRA tephra as seen in other Yukon lakes in this study, where the sedimentation rate decreases temporarily. The general trend in the core is for a gradual decrease in sediment accumulation rate, reaching 43.063 yrs/cm (or 0.023 cm yr^{-1}) by the surface of the core.

5.3.7 *Marcella Lake, Yukon Territory*

Marcella Lake, core Z (a short surface core) was taken due to the unsuitability of the uppermost section of core A (Anderson, 2004) for the determination of the charcoal record. In order to accurately reconstruct CHAR, two radiocarbon determinations were undertaken on samples from core Z. These together with the WRA tephra and correlation using the LOI and magnetic susceptibility results (see section 6.3.4), were used to combine core A and Z.

The radiocarbon determinations for core Z are shown in Table 5.6a. The date CAMS-115581 (highlighted in grey) is significantly older than CAMS-115580 (a paired date), and the WRA tephra (which occurs 38cm below the dated level), suggesting that the macrofossil was old material probably deposited during an erosional event, or that there was an impurity in the sample that was submitted due to the carbonate nature of the lake. It is conceivable that reworking of the macrofossil up-core could occur (as an artefact of coring for example); however, material is not collected from the sides or near the surface of the cores, to reduce the chance of such a possibility. This date is considered to be anomalous, and is not used to inform the age-depth model.

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age Range (years BP)
CAMS-115580	50-51	Charcoal	1080 +/- 35	932-1057 (100%)
CAMS-115581	50	Macrofossil	3570 +/- 35	3726-3751 (4.9%) 3762-3794 (7%) 3821-3975 (88%)

Table 5.6a: Radiocarbon determinations for Marcella Lake, core Z

Core A has been used in a previous study to reconstruct paleohydrology in the southern Yukon, and a chronology exists (Anderson, 2004; Anderson *et al.*, In Press; Anderson *et al.*, 2002, 2005b). Five radiocarbon determinations from core A are presented in Table 5.6b (Anderson *et al.*, 2005b), and are used in Figure 5.7 to construct an amalgamated age-depth model for Marcella Lake core A-Z.

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age (years BP)
CAMS-96832	101	Wood	2365 +/-40	2351
CAMS-73144	198	Wood	5330 +/- 40	6070
CAMS-73145	241	Wood	7370 +/- 110	8180
CAMS-96834	466	Wood	8605 +/- 40	9546
OS-12131	520	Wood	9090 +/- 55	10220

Table 5.6b: Radiocarbon determinations for Marcella Lake, core A (see Anderson, 2004). Sample depths are un-adjusted depths (see section 6.3.4 for details).

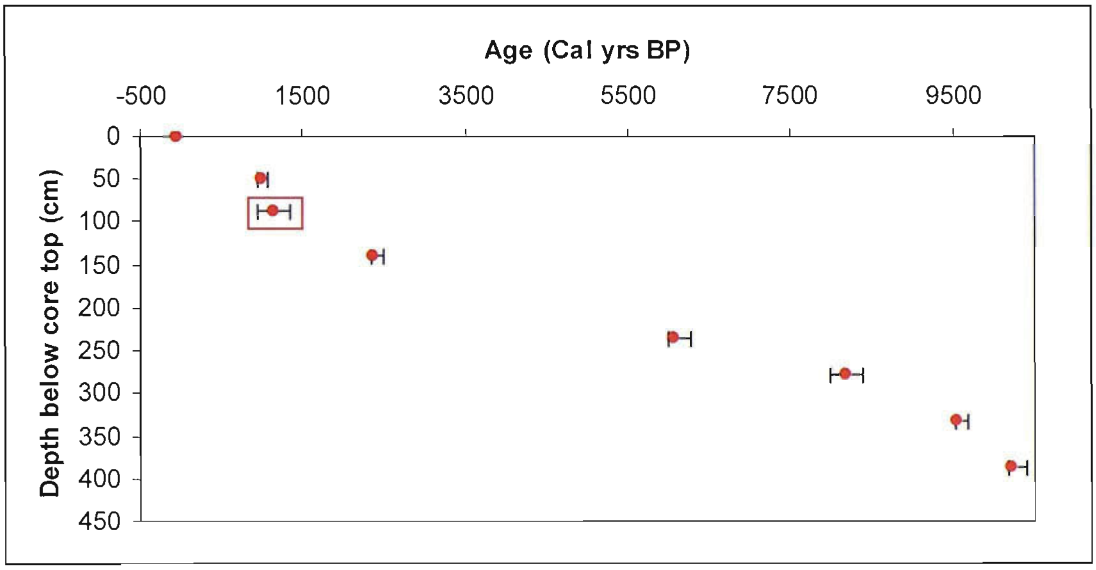


Figure 5.7: Age-depth relationship for Marcella Lake, core A-Z, using six radiocarbon determinations, the WRA tephra (demarcated by the red box), and the surface (red dot without error bars).

Sedimentation rate appears to be gradually increasing over time, with a period of highly increased sediment accumulation after the WRA deposition. The age-depth relationship for core A has already been interpreted using linear interpolation (Anderson *et al.*, In Press; Anderson *et al.*, 2005b) The surface of the amalgamated core A-Z will be interpreted using a spline interpolation due to the inflection in the curve created by the WRA, which impacts the CHAR calculations. Linear interpolation is used to interpret the age-depth relationship from 138cm to the base (consistent with the Anderson *et al.*, 2005b interpretation). The basal determination dates the core to *c.* 10000 cal yrs BP, with an initial accumulation rate of 12.48 yrs/cm (or 0.08 cm yr^{-1}). The sediment rate decreases up core to 49.02 yrs/cm (or 0.02 cm yr^{-1}) until ~240cm. After this point the sedimentation rate increases slightly to 38.31 yrs/cm (or 0.03 cm yr^{-1}). The sedimentation rate continues to increase up-core to the surface at 15.85 yrs/cm (or 0.063 cm yr^{-1}).

5.3.8 *Jelly Bean Lake, Yukon Territory*

Jelly Bean Lake has been used in a previous study (Anderson, 2004; Anderson *et al.*, 2005a), and a complete chronology exists. The radiocarbon determinations are shown in Table 5.7 below. A reservoir correction of 155- ^{14}C years was applied to each radiocarbon age before calibration (^{14}C-year equivalent of the difference between the modern carbon fraction of 240 +/- 30 radiocarbon age and 90 BP determined by ^{210}Pb dating of the sample at 15cm – see Anderson, 2004 for a further details).

Publication Code	Depth (cm)	Sample Material	Conventional Radiocarbon Age	Calibrated Age (years BP)
OS-38712	15	wood	240 +/- 30	
CAMS-91761	41.5	Wood	920 +/- 35	670
CAMS-91762	59.5	Wood	1175 +/- 35	930
CAMS-96825	87	Charcoal	1750 +/- 40	1520
CAMS-92155	170.5	pollen	4000 +/- 35	4220
CAMS-91763	244.5	Wood	4800 +/- 35	5420
CAMS-92157	336.5	pollen	6675 +/- 50	7430

Table 5.7: Radiocarbon determinations for Jelly Bean Lake, core C (see Anderson, 2004)

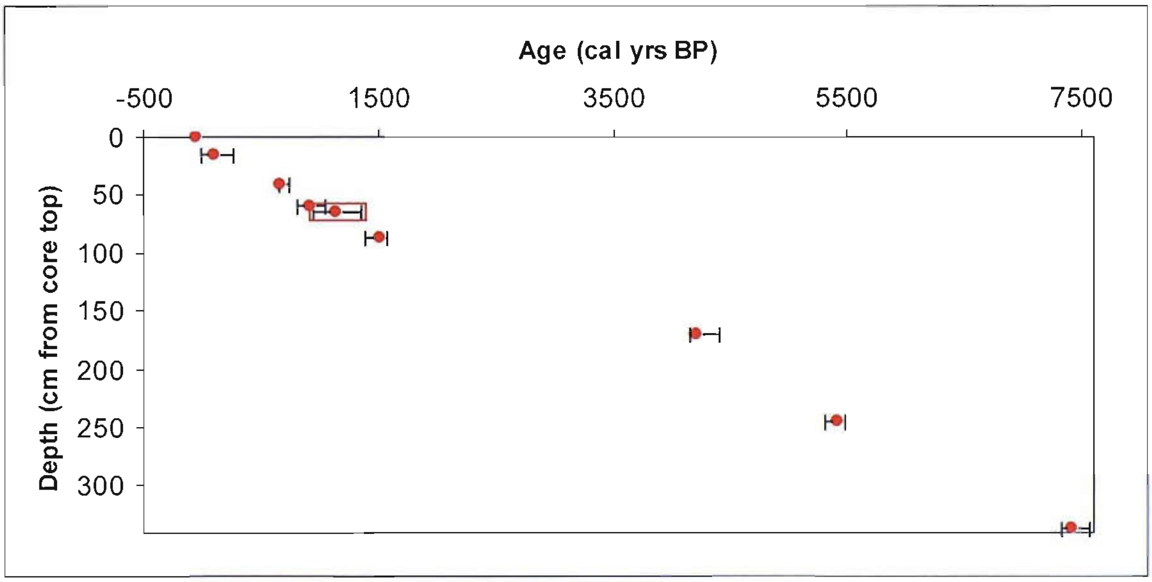


Figure 5.8: Age-depth relationship for Jelly Bean Lake, core C, using seven radiocarbon determinations, the WRA tephra (demarcated with the red box), and the surface of the core (red dot with no error bars).

The total sediment core chronology is based on ^{210}Pb, ^{137}Cs, ^{14}C AMS measurements and the WRA tephra. The age-depth relationship presented in Figure 5.8 indicates a relatively steady sedimentation rate.

Anderson applied linear interpolation between dated depths to determine the age-depth relationship (Anderson *et al.*, 2005a), which is preserved here for correlation of results from this project. The slower sediment accumulation between 64 and 59.5cm is consistent with an abrupt decrease in calcium carbonate production and/or preservation following the deposition of the WRA, presumed to be an abrupt bicarbonate equilibrium shift in response to the ash deposit (Anderson, 2004). This is the most parsimonious explanation; however, the author would like to note that due to the relatively large error bars associated with the WRA date, a linear relationship could be plotted through the dates immediately above and below which would fall within the errors on the WRA, and not indicate a shift in sedimentation.

Considering that the chronology is not fully controlled by the author, it should also be acknowledged that an extra date between 120 and 150cm would ideally have been obtained to constrain the model within the ~ 2500 yr gap. There is no evidence in the core stratigraphy to indicate an alteration in sediment regime, and this is probably why a sample was not submitted previously, however, for the interpretation of CHAR it is better to have a near-millennial scale chronology. Jellybean lake is used here as a supplementary record, and therefore these issues are not pivotal to the project as a whole, but further dating will be addressed in conjunction with future research to produce a vegetation history, and to further interpret the charcoal record (see chapter 8).

Chapter 6. Vegetation History

6.1 Introduction

The following section has been sub-divided into individual reports on each site investigated for this project. Each section consists of an overview of the stratigraphy and the physical sediment analyses plus the pollen (and stomata where relevant) analyses, and each is based on the chronology established in chapter 5. Site-specific interpretations are included in each section where appropriate. Results of the charcoal analyses will be presented in Chapter 7.

An overview is presented for sites that have been used in previous studies and for which site details are published elsewhere. Proxies that were analysed specifically for this project are presented and interpreted separately. For a complete list of all analyses, and the relevant analyst please refer to the "Table of Analyses" on page xii. All depths are reported from the water surface, unless otherwise stated.

6.2 Alaskan Lakes

The Alaskan sites are essential to the project as the Alaskan interior remained unglaciated, and they therefore provide long pre-spruce records. However, as pine is absent from the region it was not considered essential to conduct charcoal analyses over the entire length of the core. Therefore, both Little Harding and Birch Lake data consist of windows of analyses (charcoal and pollen) over the transition from deciduous- to coniferous-dominated forest. Stratigraphy and physical sediment characteristics have been determined for the entire core lengths in order to establish core integrity.

6.2.1 Little Harding Lake, Core Z

6.2.1.1 Stratigraphy

The total core length retrieved was 559cm (Figure 6.1), from a water depth of 833cm. The analytical window (1185-1305cm) is shown in Figure 6.2. The core consists of silty gyttja of varying shades of brown (see Munsell colour references in Figures 6.1 and 6.2). From core surface to basal material the sediments increase in dryness and elasticity. Fine banding is observed throughout the length of the core, with silt bands regularly present. Between 1335 and 1355cm the core is oxidised. From 1260-1315cm there is a sediment change to very dry red banded sediment, with fine plant remains observable. The transition from gyttja to gravel at 1402cm is abrupt, with the core terminating at 1415cm.

The analytical window (1185-1305 cm, Figure 6.2) begins with a major silt band. This is probably a temporary erosional event in the catchment which deposited a large block of silt into the record. From 1190-1260 cm the core consists of brown gyttja, with numerous thin (between 1 and 3mm wide) silt bands (laminations). In this section the core is relatively moist and has a degree of elasticity.

Between 1260cm and 1275cm the sediment changes to a reddish colour, but retains the laminations. There is a short-lived reversion to grey-brown gyttja with silt banding between 1275 and 1285cm. From 1285cm until the end of the window (at 1305cm) the sediment type reverts to the laminated red gyttja. In this section the sediments appear visually to be increasingly desiccated, although this is not reflected in the water content curve, and the structure consists of large sections of plant material.

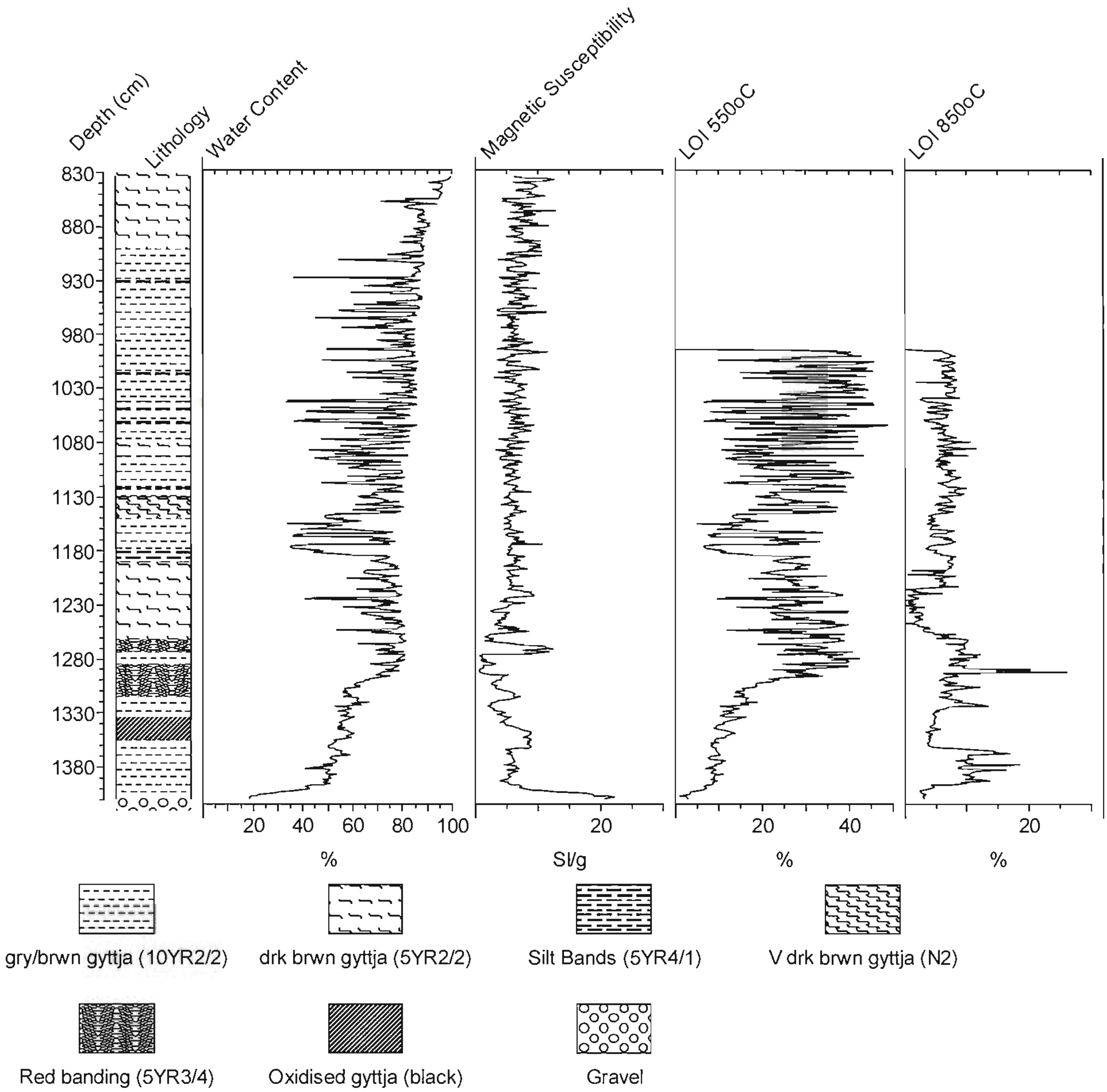


Figure 6.1: Little Harding Lake, core Z. Whole core representation of stratigraphy, magnetic susceptibility, organic and carbonate content. LOI results for the surface sediments were not conducted due to the distance from the analytical window.

6.2.1.2 Loss on Ignition

The basal part of the core (below 1350cm, Figure 6.1) contains higher carbonate levels (~10-15%) than at any other point in the core, with the exception of a peak at 1279cm (Figure 6.2). From this point onwards, carbonate levels decrease to minimum values at 1228cm (<1%). Up-core from this carbonate levels are relatively stable averaging ~8%.

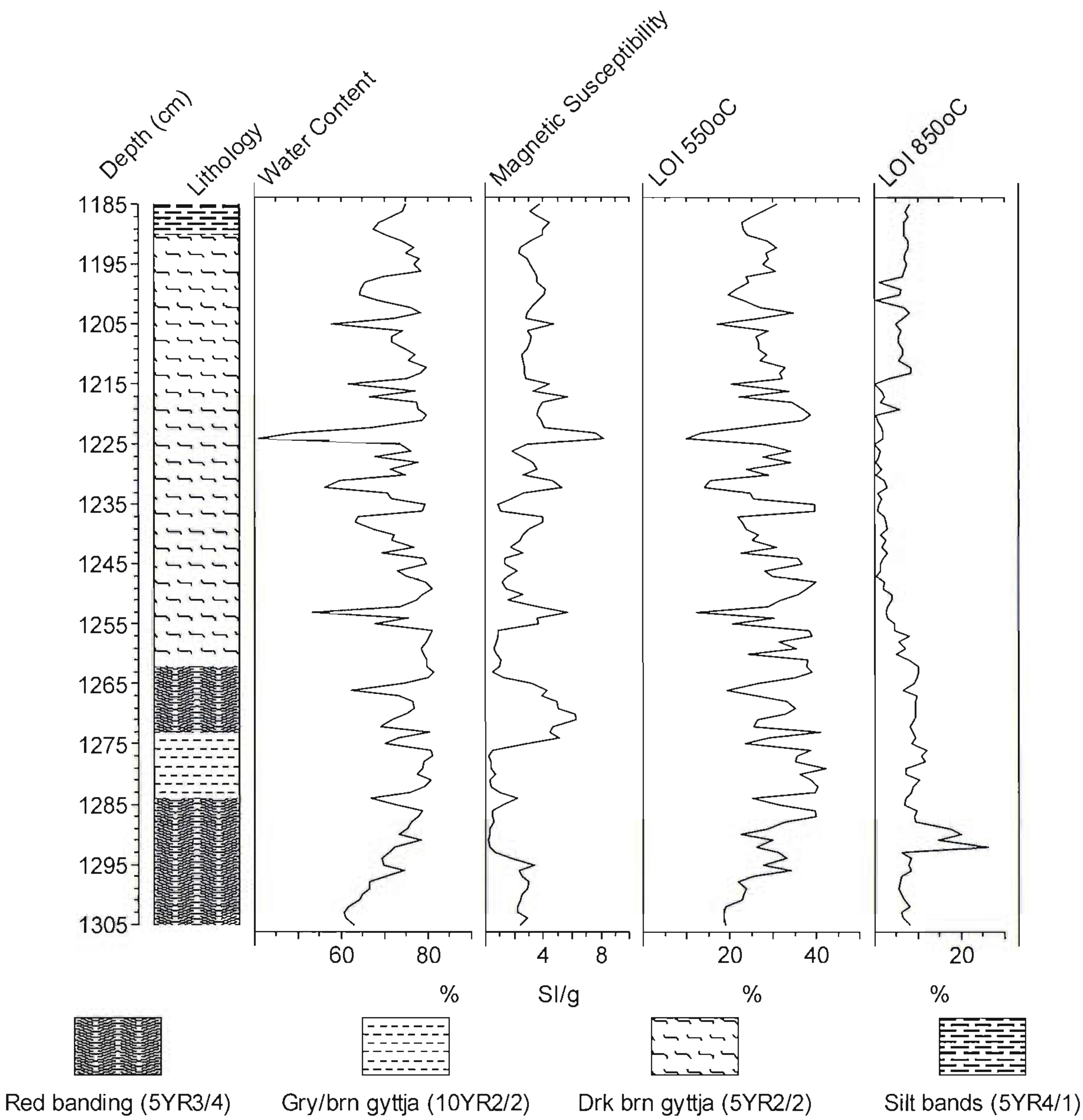


Figure 6.2: Little Harding Lake, core Z analytical window showing stratigraphy, magnetic susceptibility, water, organic and carbonate content. ***Note*** *the y-axis reflects 120cm of sediment (1185-1305cm), stretching the curves vertically compared to Figure 6.1.*

Organic content is low in the basal sediments (Figure 6.1), rising steadily from <5% at 1400cm to ~40% at 1280cm. From this point upwards, organic content is highly variable probably due to the presence of frequent silt bands within the core. In general the organic content remains above 15%, with the highest levels reached between 1030cm and 1000cm.

Closer inspection of the analytical window (Figure 6.2) shows that carbonate production overall declines from *c.* 9 - <1% (1305-1212cm), with one short-lived peak where carbonate content exceeds 20% (1292-1289cm). There appears to be a step-like increase in carbonate production at 1213cm, where average values are *c.* 7%, to the top of the analytical window.

Organic content is much more variable than carbonate production overall averaging at *c.* 30%. Generally the organic content curve reflects the inverse of changes in the magnetic susceptibility, which is reflective of the periodic silt influx into the lake.

6.2.1.3 Magnetic Susceptibility

The basal sediments show variable susceptibilities (Figure 6.1), with a sharp decrease in values from the base (>20 SI/g) to 1385cm (~5 SI/g). This sharp decline in the basal section is indicative of lake establishment, with high values probably signifying in-wash of large quantities of ferromagnetic material, most likely bedrock (late Precambrian to early Paleozoic metamorphic rock), and lower values reflecting the increase in productivity and organic input into the system. A more gradual decreasing trend in values is observable from 1385cm to 1279cm, at which point a relatively large spike is present, possibly a transient high magnitude in-wash event (reflected in the carbonate curve as discussed in 6.2.1.2). From 1279cm to 1230cm values rise to *c.* 8 SI/g.

Magnetic susceptibility is relatively constant from 1230cm upwards, with a slight increasing trend at the very top of the core. The samples show positive susceptibility in this section and are probably dominated by paramagnetic material, with a proportion of canted antiferromagnetic materials (Dearing, 1999).

The analytical window (Figure 6.2) highlights the fluctuations in the magnetic susceptibility curve. The fluctuations in susceptibility are exaggerated in Figure 6.2 due to the shortened length of core represented, and most probably reflect the banded-nature of the sediments. There is no noticeable trend, with moderate positive susceptibilities indicating a predominance of canted antiferromagnetic materials in this section of the core.

6.2.1.4 Water Content

The water content curves are presented in Figure 6.1 and 6.2 above. The overall trend for the whole core is of increasing water content up-core towards the surface. The basal sediments have low water content levels ~20%, and they increase relatively rapidly to ~80% by 1285cm. There is observable short-term variability within the curve, reflective of the banded nature of the sediments in this core. Reduced water content is associated with silt bands. On average, the water content of 80% is maintained up-core to ~850cm, where it increases to near 100%.

Within the analytical window the curve reflects the end of the initial increase from 1305-1285cm. The curve pattern remains between 40 and 80% water content, with periods of lowered water content reflecting the inverse of the magnetic susceptibility curve associated with silt bands.

6.2.1.5 Pollen Diagrams

Presentation of the full pollen diagrams is restricted to the appendices (see Appendix 1); here key taxa, critical for the interpretation of vegetation history with regard to the fire regime, are represented in pollen percentage (Figure 6.3) and influx diagrams (Figure 6.4).

The pollen assemblage zone LHLZ-1 (1305 -1245cm) is dominated by *Betula* (*c.* 65-80%). *Salix* (*c.* 10%), *Populus* (*c.*2-3%) and *Juniperus* also contribute significant amounts of pollen to the assemblage, although abundances decrease to very low amounts by the end of the zone. Poaceae dominates the herbs (*c.* 5-10%), with *Artemisia* (*c.* 2-3%), however, the proportions decline to *c.* 1% by the end of the zone. Monolete spores are also relatively abundant (*c.*5-8%), declining to very low levels at the very end of the zone. The influx of *Salix, Populus, Artemisia,* Poaceae pollen and monolete spores are at the highest in the deciduous zone, with more gradual and sustained decreases through the transition into the spruce zone, than is suggested by the percentage diagram.

The transition between the zones LHLZ-1 and LHLZ-2 is highlighted by the decline in *Salix, Populus Artemisia*, Poaceae and monolete spores, and the increase *Picea* and Alnus pollen, and the overall tree and shrub proportions.

The pollen assemblage zone LHLZ-2 (1244 – 1125cm (topmost level counted)) is clearly demarcated by the sustained increase in *Picea* pollen percentages (≥10%) and influx, and the sustained contribution of *Alnus* pollen to the record. The *Betula* proportion is relatively unchanged; however there is a marked increase in the influx values.

In the influx diagram one section of the record (1175-1155cm) (Figure6.4) shows overall pollen influx values that are much lower than the levels above and below. This section is subsequent to a significant silt band (1180-1190cm), much of which is a continuous block with no visible gyttja, indicating a single input event. It seems likely that a short-lived transient erosional event deposited the silt, which caused a dilution of incoming pollen.

6.2.1.6 Vegetation History

The pollen percentage diagram (Figure 6.3), and the pollen influx diagram (Figure 6.4) show the transition from a pollen assemblage characterised by deciduous woody taxa to one including up to 20% spruce. In LHLZ-1, *Betula* and Poaceae contribute the highest proportion to the pollen record (60-70% and 5-10% respectively) indicating a sparse herbaceous landscape. The pollen assemblage in this zone is consistent with other regional deciduous zone records (for example Ager, 1975; Bigelow, 1997; Edwards *et al.*, 2000), and from here will be referred to as the deciduous zone. It is probable that the *Betula* present was shrub (*Betula nana* and/or *glandulosa*) birch, with increasing proportions of tree birch (*Betula papyrifera*, paper birch). after the transition to zone LHLZ-2. Although grain size determinations were not conducted on *Betula* grains in this study, there is evidence for local tree presence around the Fairbanks area at this time from macrofossils (Hopkins *et al.*, 1981) and seeds at Wein Lake (Hu *et al.*, 1993).

Picea pollen gradually increases in abundance through the end of the deciduous zone, reaching co-dominance with *Betula* by the beginning of LHLZ-2, which will be referred to from here as the spruce zone. Overall spruce pollen percentages remain below 20%, which is lower than other records (Ager, 1975; Bigelow, 1997). The spruce zone is characterised by the decline in *Populus, Salix, Poaceae* and monolete spores, indicating more tree-dominated closed vegetation.

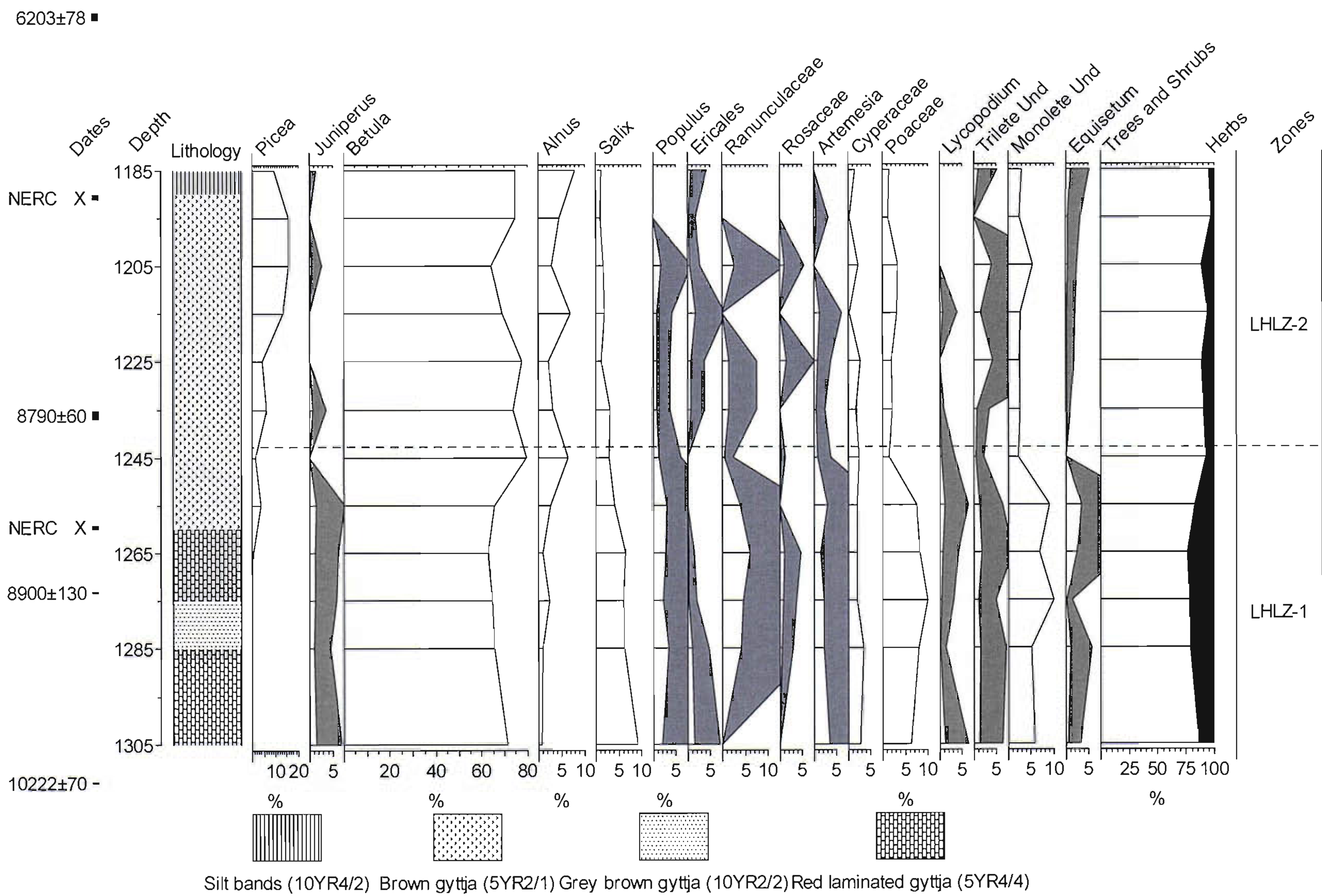


Figure 6.3: Pollen Percentage diagram for the analytical window covering the spruce rise at Little Harding Lake, Alaska (Core Z). Key taxa represented only, full pollen diagram is included in Appendix 1. Shading represents a 5x exaggeration (Analyst: LMFS)

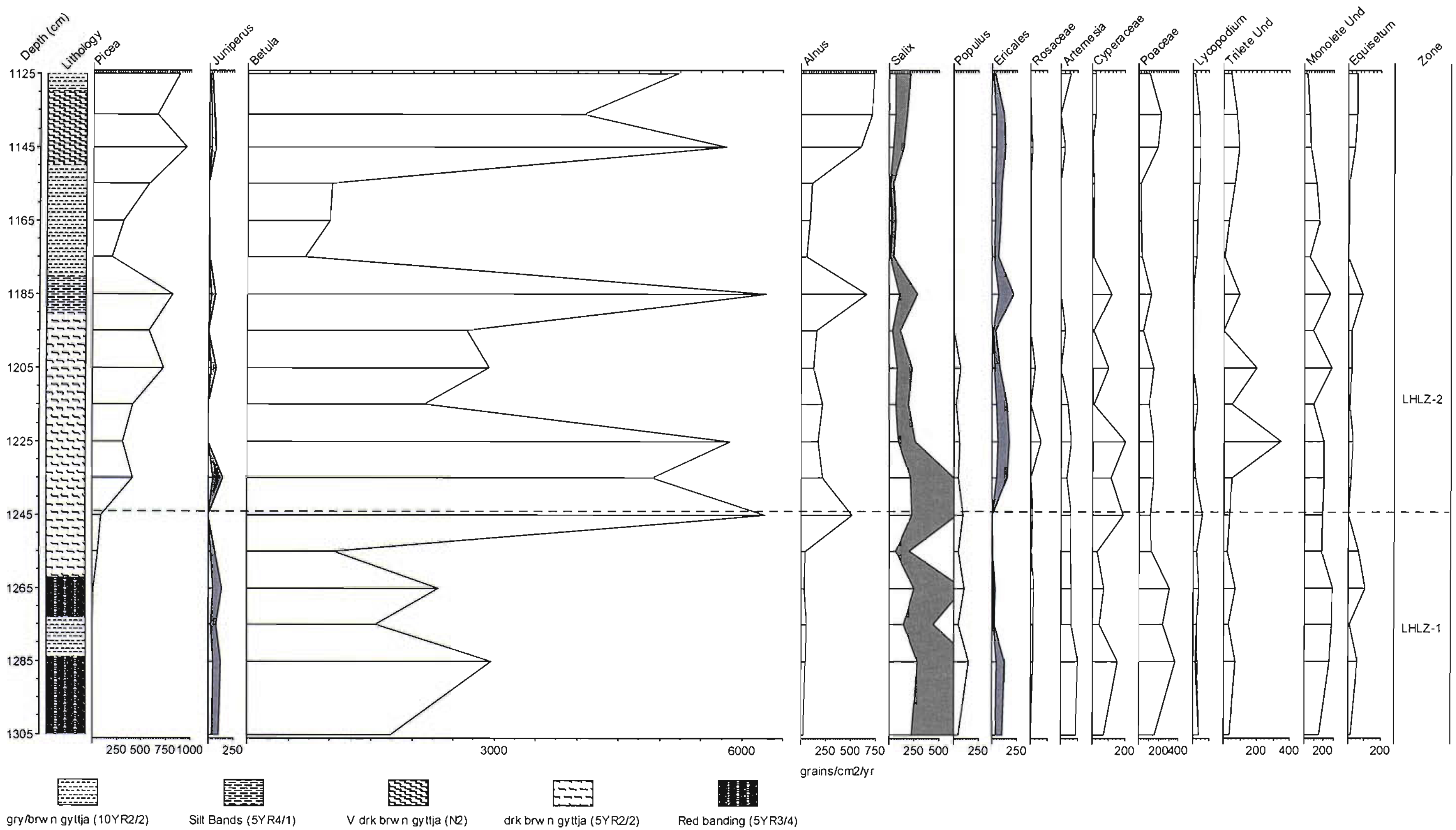


Figure 6.4: Little Harding Lake, Core Z, showing pollen influx values for key taxa, representation of all species is contained in Appendix 1. **Note:** *Grey shading indicates a 3x exaggeration factor applied to relevant woody taxa to keep a constant x-axis scale. Herbaceous taxa have a different x-axis scale. (Analyst: LMFS).*

6.2.2 *Birch Lake*

6.2.1.7 Previous Results and Interpretation

Birch Lake was sampled and analysed firstly by Ager (1975)as part of a Ph.D. project to reconstruct the environmental history of the area, and subsequently by Bigelow (1997) who reconstructed late Quaternary vegetation in relation to lake level changes. The cores taken by Bigelow were used by this study. In the original study, a suite of over 10 cores were collected for pollen analysis and lake level reconstruction. A composite core (Birch G/H) provided a record from present to ~14 ^{14}C yr BP (Bigelow, 1997). This project utilised a section of core 97b, which covers the period *ca.* 6800 - ~10300 cal yrs BP.

The sediments are typically lacustrine gyttja with varying amounts of silt and lamination (Table 6.1). A near continuous input of eolian silt is indicated via the presence of mica flecks through the core length. The analytical window for charcoal analysis ranges from 1600 – 1700cm (drive 4), and in-depth interpretation of the sediments and pollen are restricted to the window.

Drive start (cm)	**Drive end (cm)**	**Sediment Description**
1500	1600	1500-1510 cm: un-laminated olive black (5y 2/1). 1510-1588 cm: light lams olive gray (5y 4/1). 1588-1600 cm: laminated (5y 3/2).
1600	1700	1600-1626 cm: slightly darker olive gray (5y 4/1). 1642-1666 cm: thinly laminated. 1666-1682.5 cm: olive gray (5y 3/2). 1682.5-1685 cm: brown 5yr 4/4.
1700	1800	1700-1800 cm: silty, olive grey (5y 5/2 or 5y 4/1) with mica flecks throughout.
1800	1900	1800-1900 cm: silty dark yellowish brown (10yr 4/2) to olive gray (5y 4/1).
1900	1986	1900-1907 cm: dark yellowish brown (10yr 4/2). 1907-1927 cm: silty olive gray and light olive gray (5y 4/1 to 5y 6/1) with darker lams. 1927-1961 cm: silty olive gray (5y 3/2) with olive black (5y 2/1) striations. 1961-1986 cm: grayish olive green (5 gy 3/2), higher conc. of mica flecks than in rest of drive.
1986	2016	1986-2011 cm: grayish olive green (95gy 3/2), mica flecks throughout. 2011-2016: gravelly sediments.

Table 6.1: Sediment description of Birch Lake, core 97b (Bigelow, pers. comm.)

The magnetic susceptibility values are high (averaging *c.* 20%) and relatively stable through the analytical window (Figure 6.5). These levels indicate a relatively constant input of ferromagnetic material. There is an episode of heightened magnetic susceptibility values (to ~40%) between 1640 and 1625cm. This episode probably reflects a period of increased sediment influx into the basin, as suggested by the more grey appearance of the sediments.

Organic content is low (~5%) from core base to ~1700cm (Bigelow, pers. comm.), from which point it increases to ~10%. From here to the top of the analytical window the organic content ranges between 9-13%, with a slight increasing trend. Carbonate content is generally low (<5%) throughout the whole record, with very little variation.

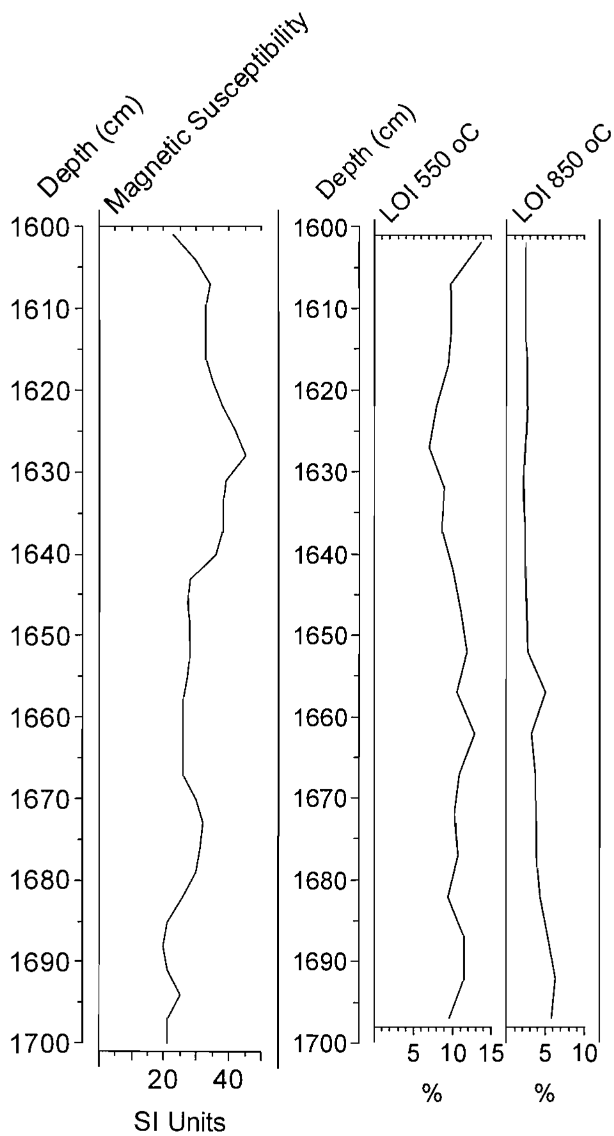


Figure 6.5: Birch Lake, core 97b, showing magnetic susceptibility, organic and carbonate content for the analytical window ***1600 – 1700cm*** *below water surface.*

6.2.1.8 Pollen Diagrams

Previous pollen analysis of core 97b produced intermediate resolution (5cm) pollen data for a generalised vegetation reconstruction (A. Krumhardt, B. Finney, and M. Edwards, unpublished data). High resolution pollen analysis (1cm contiguous samples from 1644-1664cm see Table of Analyses page xii) was completed for this project, to pinpoint exactly the transition from deciduous- to spruce dominated vegetation. Amalgamations of both analyses are represented in Figures 6.6 (percentages) and 6.7 (influx).

The pollen assemblage zone BL97b-1 extends from 1900 to 1876cm and is characterised by low *Betula* and high Poaceae, Cyperaceae and *Artemisia* percentages. These pollen spectra and influx values are consistent with other herb-zone records at Birch Lake (see Ager, 1975; Bigelow, 1997).

Pollen assemblage zone BL97b-2a ranges from 1875cm to 1655cm (~9200 cal yrs BP), and is characterised by the rapid increase and dominance of *Betula*, in addition to decreases in *Salix, Artemisia,* Cyperaceae and Poaceae percentages. These combinations are indicative of the "deciduous" or birch zones, in other regional records, including Birch Lake (see for example, Bigelow, 1997).

A sub-zone, BL97b-2b was identified between 1700 and 1655cm, which reflects a period with similar characteristics to the main deciduous zone, but with the addition of *Populus*, which constitutes 10% of the pollen sum at this point.

The rise in spruce occurs in pollen zone BL97b-3, between 1655 – 1640 cm (~9200 cal yrs BP) exceeding 15% by 1650cm. Continued dominance of *Betula*, in the absence of *Populus*, accompanies the appearance of spruce.

This core has not been analysed for pollen above 1620cm. However, it is reasonable to assume that the characteristics of the spruce zone (pollen zone BL-5 in core G/H used by Bigelow (1997)) will continue above this level. In any case, the analytical window for charcoal ends at 1600cm. With a secure radiocarbon date and observable stratigraphic integrity it seems unlikely that any unconformities exist within the core area of interest. In core G/H the remainder of the core to the surface (~6900 ^{14}C yr BP to present) sees the arrival of *Alnus*, and co-dominance forming between *Picea, Betula* and *Alnus* in the pollen assemblage.

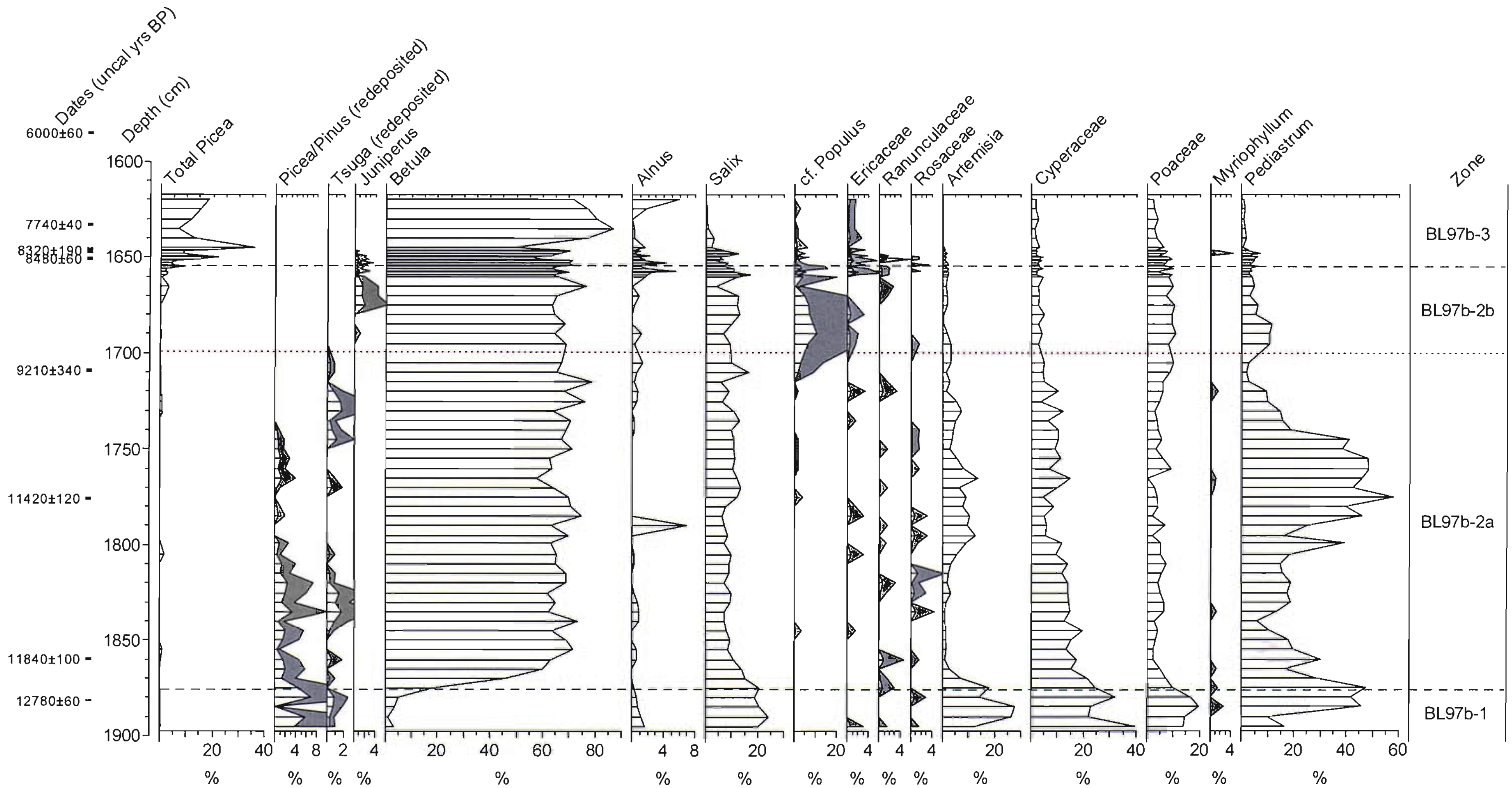


Figure 6.6: Pollen percentage diagram for Birch Lake, core 97B. Key taxa only represented, full pollen diagram is located in Appendix 1. Shading represents a 3x exaggeration. The red dotted line indicates the start of the analytical window, in addition to the transition between zones 2a and 2b. Analysts: LMFS and NHB.

6.2.2.3 Vegetation History

The earliest vegetation documented at this site is a sparse assemblage of herbaceous plants and *Salix* (Figures 6.6 and 6.7), BL97b-1. There is relatively little sediment preserved in this zone in core 97b, but there are distinct similarities with an earlier record of Ager (1975), which indicates that the landscape was sparsely covered with *Salix*, grasses, sedges and *Artemisia*.

The transition to BL97b-2a indicates an abrupt transition to shrub *Betula*-dominated vegetation, although the composition of herbaceous taxa remains relatively unchanged over the transition, The fact that the increase in *Betula* was so rapid and abrupt suggests that the species was probably present in the preceding herb zone, but in low densities, and probably in scattered locations. It is thought that the expansion of *Betula* was probably triggered by a combination of higher summer temperatures and precipitation (Bigelow, 1997). Summer insolation was rapidly increasing during this time (Berger, 1978), and lake level analyses at Birch (Abbott *et al.*, 2000) and Windmill (Bigelow & Edwards, 2001) Lakes suggest that levels were increasing during this period, indicating higher effective moisture.

In sub-zone BL97b-2b *Populus* increases to ~10% of the terrestrial pollen (Figure 6.6). As *Populus* pollen does not preserve well, nor travel far from its source (Edwards & Dunwiddie, 1985), this indicates a significant change in the local vegetation around the lake. It is probable that the *Populus* found here is *P. balsamifera,* as it has been identified in the central Brooks Range at this time period (Brubaker *et al.*, 1983). This species is shade-intolerant, and a pioneer species, common on well-drained sites (Viereck *et al.*, 1992), and shows a positive growth response to warm early summer temperatures (Lev, 1987). This suggests that *Populus* expansion is linked with summer temperature and also the availability of habitat for such an early successional species.

At ~9500 cal yrs BP, *Picea* appears on the landscape, marking the start of zone BL97b-3, resulting in a mixed vegetation with *Picea* and *Betula* co-dominant (Bigelow, 1997). This date is slightly earlier than that of Ager (1975, 9800 cal yrs BP) but consistent with that of Bigelow (1997, 9500 cal yrs BP) As at Little Harding Lake, the increase in *Betula* influx probably indicates the appearance of tree birch in the region. The increase in spruce coincides with an increase in water level at Birch Lake (Abbott *et al.*, 2000), the dynamics of which will be discussed further in Chapter 8.

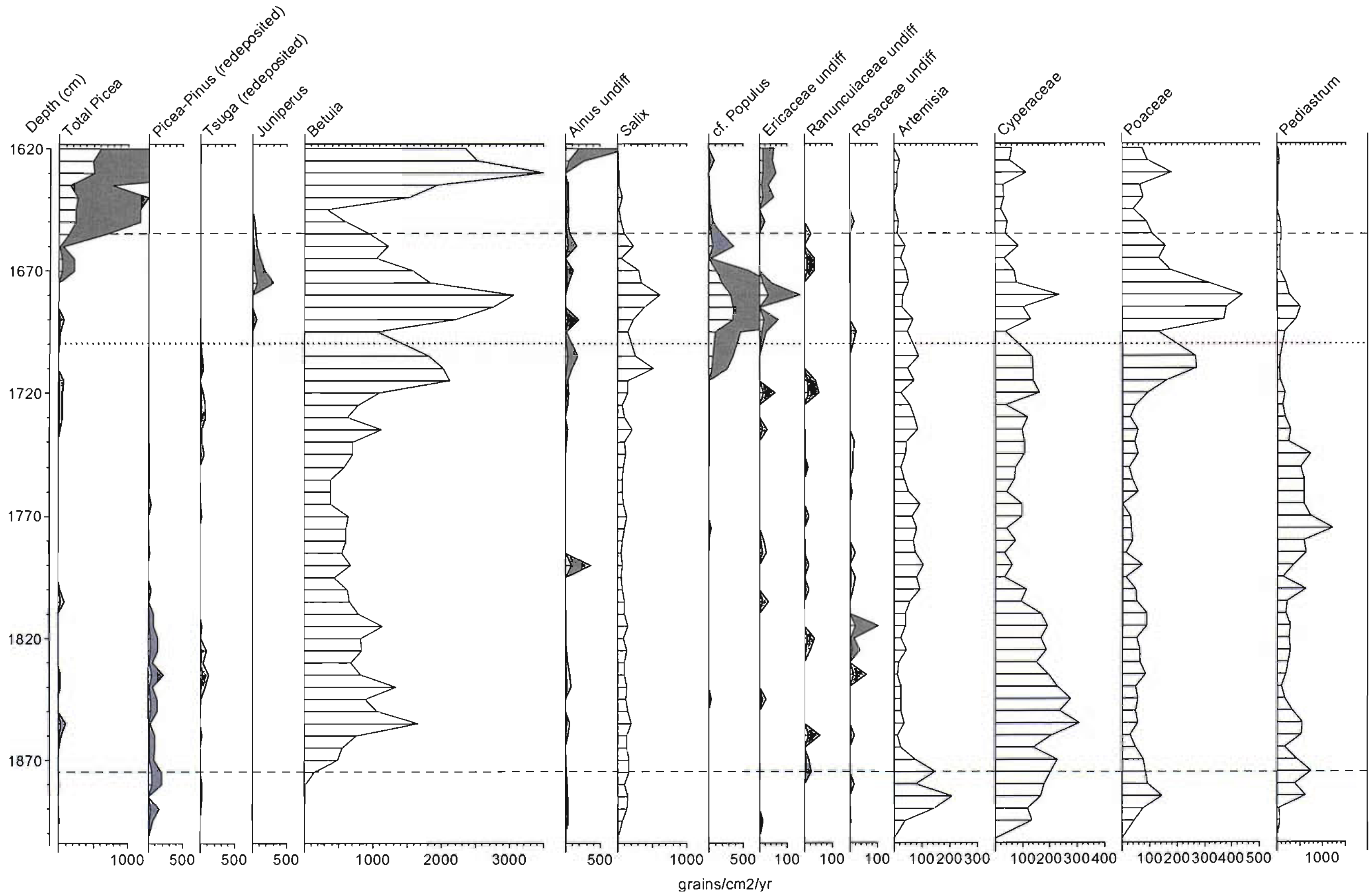


Figure 6.7: Pollen influx diagram for Birch Lake. Grey shading indicates a 5x exaggeration factor. ***Note:*** *scale change for herbaceous and aquatic taxa.*

6.3 Yukon Lakes

6.3.1 Salmo Lake

6.3.1.1 Stratigraphy

The total core length retrieved was 260cm at a water depth of 783cm, with the sediment lithology represented in Figure 6.8. The transition from coarse gravel to lacustrine gyttja is abrupt at 1050cm, probably demarcating the onset of the lacustrine depositional environment. From 1050cm upwards the sediments grade from dark brown gyttja (1049-42cm) into coarse drier red/brown gyttja (1042-37cm) which contains large fragments of aquatic mosses. Between 1037cm to 965cm the sediments become lighter in colour, interrupted by a block of dark, banded gyttja (997-1016cm). The dark banded gyttja occurs again in an extensive and continuous block from 965-876cm.

At 876cm the sediment type shifts again to a dark brown gyttja without banding. This type continues to the surface of the core, interrupted by the WRA tephra between 828 and 831cm.

6.3.1.2 Loss on Ignition

The basal sediments display a rapid rise in organic content from 1055-1040cm (Figure 6.8), which coincides with the transition from gravel to gyttja observed in the stratigraphy. A rapid transition to a lacustrine environment is consistent with these results. The organic content gradually decreases up-core until approx. 955cm, where an increasing trend begins. Between 855 and 833cm there is a period of elevated organic content, which rapidly decreases from 833cm to 830cm, coincident with the WRA position. From this point to the core surface, organic content increases overall from 5 to 40%.

Carbonate production increases rapidly from lake-establishment (~1050cm) to maximum levels at 1010cm. Carbonate production gradually decreases overall from this point to ~900cm, where the levels decrease significantly, and remain below ~10% to the core surface.

6.3.1.3 Magnetic Susceptibility

Basal sediments are characterised by decreasing susceptibility, and probably indicate a transition from paramagnetic to diamagnetic materials, as the lake-environment

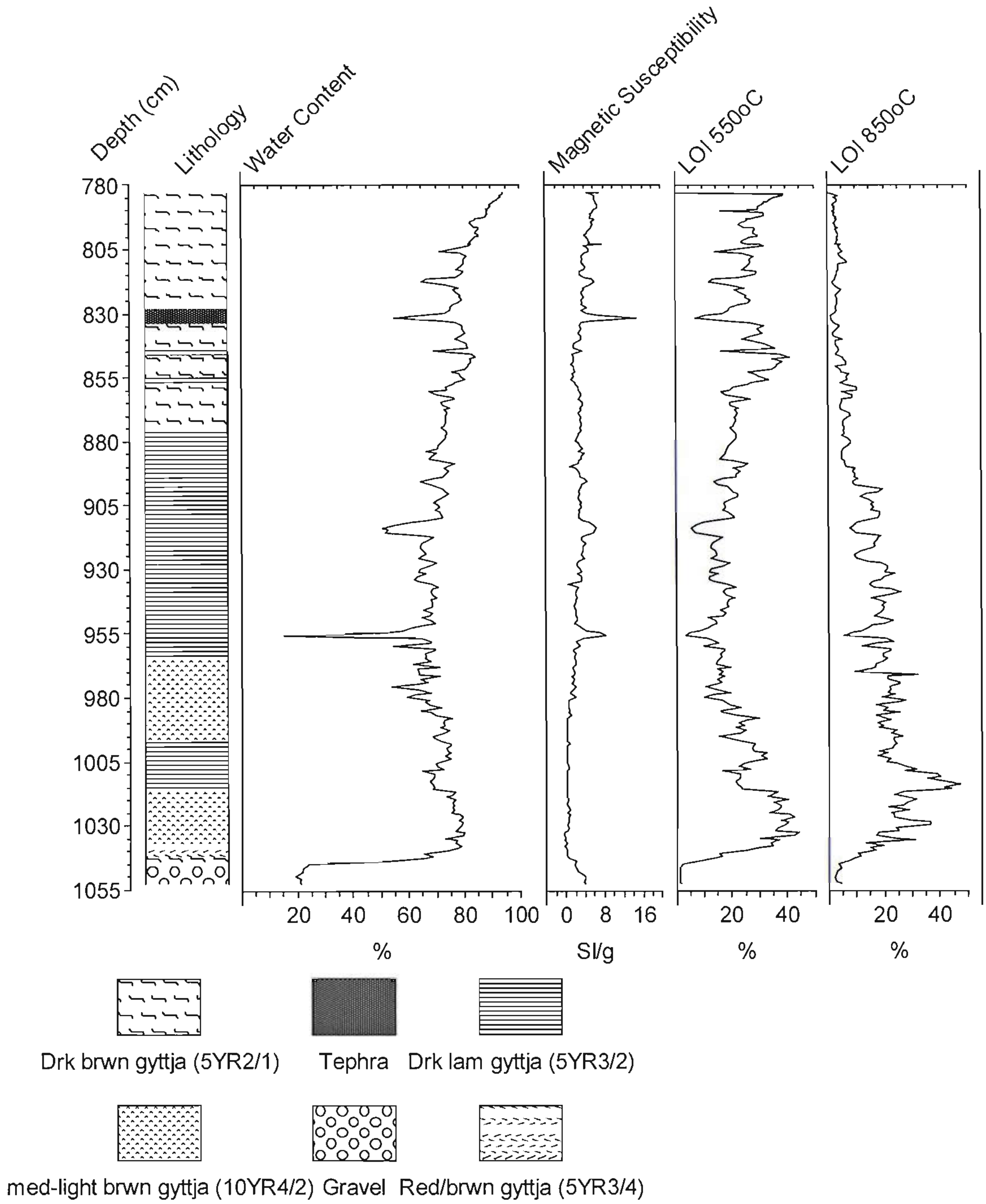


Figure 6.8: Salmo Lake, core A, showing stratigraphy, magnetic susceptibility, organic and carbonate content

formed. Diamagnetic characteristics are short-lived and coincide with the rapid onset of organic and carbonate production (Figure 6.8). From 1035cm up-core the susceptibility values gradually increase, with paramagnetic and canted antiferromagnetic materials predominating. There are two prominent peaks in magnetic susceptibility observed in the record. The peak at 832cm coincides with the position of the WRA tephra. The second peak occurs at 955-956cm, coinciding with a major silt band, indicating a temporary erosional event in the catchment.

There are a further three minor peaks in magnetic susceptibility, which are made more obvious when taken into consideration with the LOI graphs. The peaks in magnetic susceptibility are mirrored by contrasting short-lived decreases in both LOI proxies. The peaks at 914-915cm, 817-819cm and 802cm do not coincide with visible silt bands.

6.3.1.4 Water Content

The record begins with very low water content (20%). This increases rapidly to ~80% by 1040cm, and the establishment of organic gyttja deposition. The water content remains high (ranging between ~65 and 80%) until ~955cm, where there is a brief but dramatic reduction in water content to ~20%. This occurs within a section of laminated gyttja, and is likely caused by an influx of silt at this point. From 955cm the water content values display an overall increasing trend ranging from ~60-75%, with minor fluctuations to 830cm. At 830cm there is a decrease in water content associated with the WRA tephra. Above the WRA tephra the values increase to the core surface where values are ~95%.

6.3.1.5 Pollen Diagrams

The pollen percentage and influx diagrams are presented in Figures 6.9 and 6.10 respectively. A full Holocene record is present, but due to the relatively late deglaciation of the Whitehorse area the pre-spruce zone is very short.

The pollen assemblage zone SLA-1 (1053-1031cm) is characterised by the rapid increase of *Betula* pollen abundances (Figure 6.9), and initial high levels of *Populus, Artemisia,* Cyperaceae and trilete spores, and a peak in aquatic pollen. The influx diagram reflects similar patterns, with rapid increases in *Juniperus, Betula, Alnus, Salix, Artemisia,* Cyperaceae pollen and monolete spores. These characteristics are generally representative of the deciduous zone in other regional records (for example Cwynar, 1988) and are also consistent with initial lake and vegetation establishment. Due to the short length of this zone and the sharp transition from basal gravels (at 1050cm) to lacustrine gyttja (see lithology Figure 6.8) it is probably more accurate to consider this zone as a phase of lake and catchment establishment. *Populus* is present at 5 ~10% of the terrestrial pollen.

Pollen assemblage zone SLA-2 (1031-960cm) is characterised by a rapid increase in *Picea* percentages, with a coincident decrease in *Betula*, and general stabilisation of all other species abundances. The influx diagram (Figure 6.10), indicates more clearly the decrease in *Betula, Alnus, Salix* and *Artemisia* pollen, and associated increase in *Picea* pollen influx.

The zone SLA-3 (960cm-865cm) was identified due to the first appearance of *Pinus* stomata (see section 6.3.1.5 below), and very low levels of *Pinus* pollen. This is coincident with increases in the influx values of *Betula* and *Alnus* pollen, and the establishment of *Ericales* pollen consistently in the record. Coincident with the stomata and *Pinus* pollen, there appears to be a switch in the species of *Picea* dominating the landscape at this point. Gallagher (2005) found that at this time the proportions of *Picea glauca* pollen decreased, and *P. mariana* pollen contributed increasing proportions to the pollen record.

The pollen assemblage zone SLA-4 (865-783cm) is demarcated by the main rise in *Pinus* pollen influx and abundance, associated with a significant decrease in *Picea*, and an increase in Cyperaceae. The WRA tephra is present in this record at 831cm, as a deposit 1.5 cm thick, and it is highlighted by the short-lived decrease in the influx values of all species.

6.3.1.6 Stomata Analysis

The results of the stomatal analysis are represented on the pollen percentage diagram (Figure 6.9) as presence/absence data. The first appearance of coniferous stomata in the record is at 1003cm, in the spruce zone. These stomata were not identifiable as *Pinus* stomata and were probably *Picea* stomata. The first positive identification of *Pinus* stomata was at 963cm, coincident with the appearance of very first, albeit very low levels of *Pinus* pollen in the record. The stomata were consistently found upwards of this level, throughout the "pine tail" of low-level influx. Analysis was not conducted on every sample above the main *Pinus* increase; however, in every sample analysed *Pinus* stomata were identified.

6.3.1.7 Vegetation History

The earliest zone, SLA-1, contains a pollen assemblage consistent with regional deciduous zone records (Terasmae, 1966; Cwynar, 1988); however, due to the short length, and the sharp transition from basal gravels (at 1050cm) to lacustrine gyttja (see lithology Figure 6.8) it is probably more accurate to consider this zone as a lake-establishment phase. It is probably fair to consider the landscape as sparsely covered with *Salix*, grasses, sedges and *Artemisia,* with increasing proportions of *Betula. Populus* constitutes a relatively large proportion of the vegetation in this zone, and is found in these proportions at this time period in many regional records (Edwards & Dunwiddie, 1985), and is thought to represent a slight warming trend.

The transition to SLA-2 at ~9900 cal yrs BP displays an abrupt increase in *Picea* and *Betula* abundances, indicating mixed forest vegetation, with *Picea* becoming dominant through time. The *Picea* present consists mostly of *P. glauca* (Gallagher, 2005), at this stage in the record.

By ~6000 cal yrs BP (zone SLA-3), the *Picea* assemblage has shifted to increased dominance by *P. mariana* (Gallagher, 2005), In the Yukon this switch has bee attributed to increased effective moisture (Cwynar & Spear, 1995; Anderson *et al.*, 2005b; Anderson *et al.*, In Press), which is supported by data from lake-level reconstructions at Marcella Lake (ref and Chapter 2). Coincident with this is the first appearance of *Pinus* stomata within the catchment. Stomata indicate local presence of pine trees, however it is unlikely trees were present in significant numbers at this time. Migration of lodgepole pine is thought to be via establishment of isolated satellite populations far ahead of the main population front, with expansion and back-filling of populations over time (Wheeler & Guries, 1982; MacDonald & Cwynar, 1985; Peetet, 1991; Johnstone & Chapin, 2003). Thus the pattern observed here is consistent with a few isolated trees establishing in the lake catchment.

The transition to SLA-4 at ~2500 cal yrs BP is characterised by the abrupt increase in *Pinus* pollen, which more likely indicates the regional expansion of pine populations contributing pollen to the catchment. There is some evidence of a further period of increased effective moisture within the Yukon Territory at this time (Anderson *et al.*, 2005a; Anderson *et al.*, In Press).

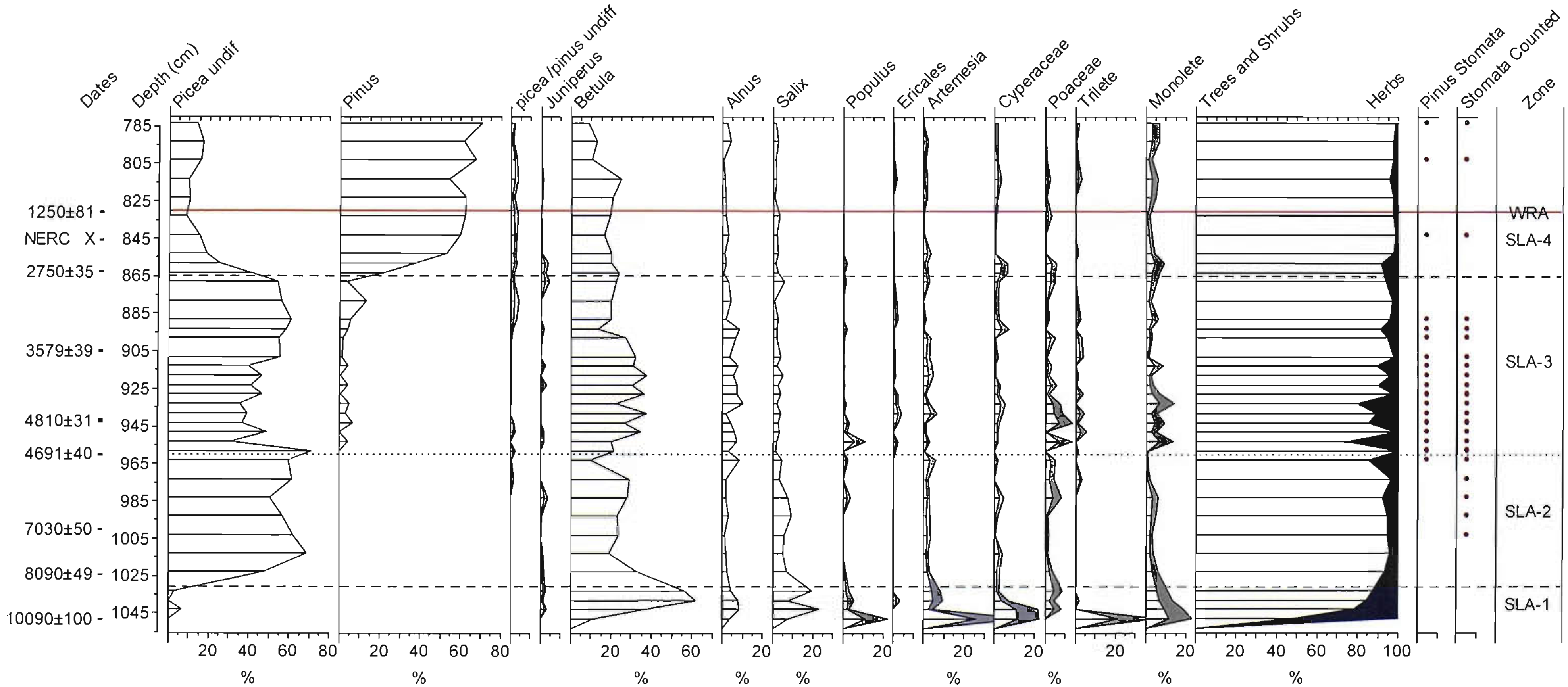


Figure 6.9: Pollen percentage diagram for Salmo Lake, core A, representing key taxa only (full pollen diagrams located in Appendix 1). Shaded graphs indicate a 5% exaggeration factor applied for ease of interpretation of low abundances. Analysts: LMFS and SMH

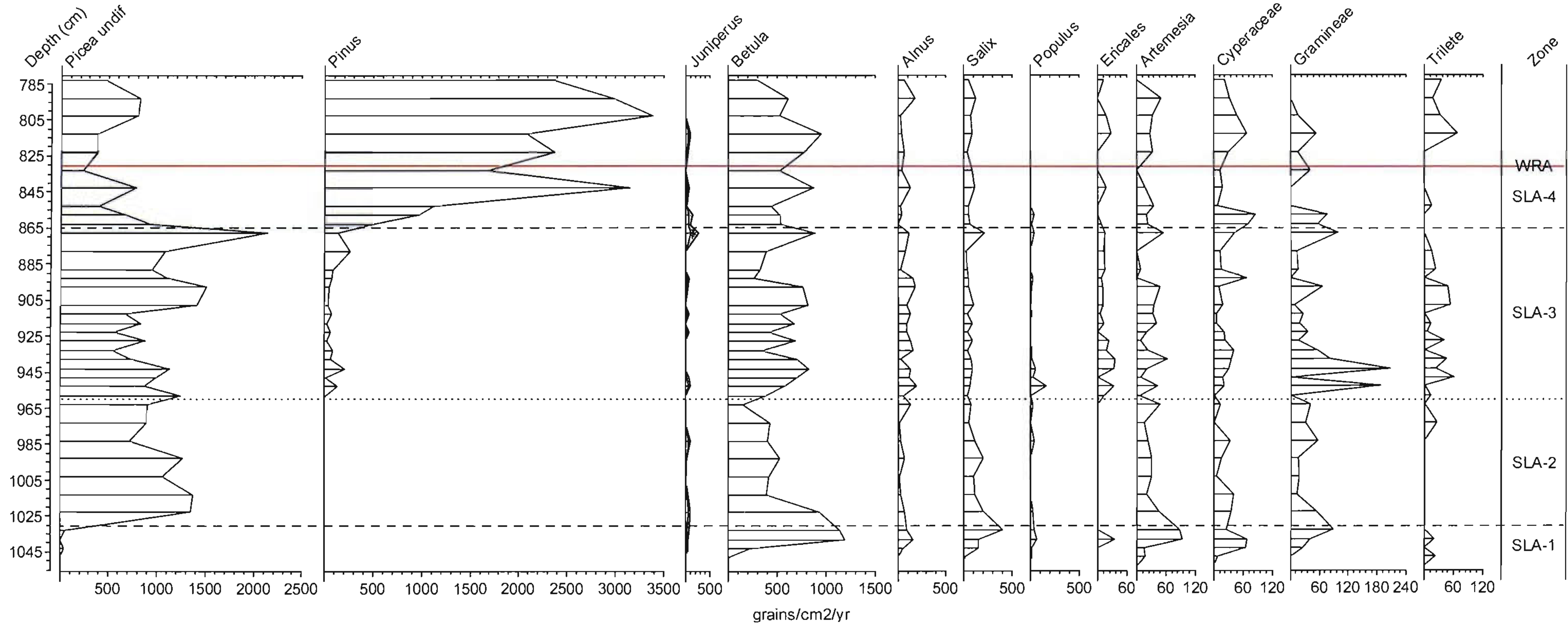


Figure 6.10: Pollen influx diagram for Salmo Lake, core A, representing key taxa. Full pollen influx diagram located in Appendix 1. The position of the WRA tephra is demarcated by the red line, and pollen zones are assigned in accordance with the pollen percentage diagram (Figure 6.9).

6.3.2 *Dragonfly Lake*

6.3.2.1 Stratigraphy

The total core length retrieved for Dragonfly Lake, core A, was 468cm, in a water depth of 211cm. The lithology is represented in Figure 6.11. The core was extremely moist, and elastic from surface to base, with relatively little variation in physical appearance. The base of the core consists of coarse gravel (687-681cm), which changes abruptly to lacustrine gyttja at 681cm. The majority of the core consists of medium to light brown gyttja, with relatively short-lived intervals of lamination or input shell-rich sediment (ostracod shells). The WRA tephra appears in the record from 306 to 303cm, overlain with a shell-rich layer.

6.3.2.2 Loss on Ignition

Basal sediments are characterised by rapidly increasing levels of organic material (Figure 6.11). Organic content remains high, averaging *c.* 67%, with short-lived fluctuations up core to 3.20m. At this point organic content decreases sharply, to minimum values at the point of the WRA deposition (3.01m). From the WRA to the core surface (2.11m), organic content is more highly variable, but generally increasing from *c.* 25 to 60%.

Carbonate content is relatively high (*c.* 20%) at the core base, but sharply decreases to <1% within 5cm. In general carbonate production is very low through the main body of the core, with values on average <10%. Short lived peaks in carbonate content in the mid-core section (e.g. 4.20m) are associated with presence of gastropod and bivalve shells in the core. Rapid onset of carbonate production is observed coincident with the WRA deposition at 3.01m. From this level to the core top carbonate content is on average >20%.

In stratified lakes carbonate is preserved only above the thermocline, and in shallow water depths (Anderson, 2004); therefore the percentage of $CaCO_3$ is controlled by lake level. In Dragonfly Lake the organic matter and $CaCO_3$ are inversely related as both are products of within lake algal productivity. The unknown variable then is the proportion in which they were being preserved as they settled through the water column, as $CaCO_3$ dissolves when pH gets too low, which often occurs either in the water column or at the sediment water interface. In this case, the WRA may have altered the pH of the lake, causing a shift in carbonate production, or the lake level may have become shallow enough to cause a shift in carbonate production coincidentally with the WRA deposit.

6.3.2.3 Magnetic Susceptibility

The axis for magnetic susceptibility in Figure 6.11 has been truncated due to the two surface core values being >70 SIunits/g, which obscures the vast majority of the core's values if plotted (see Figure 6.11 annotation). The main core susceptibility is weakly positive, most probably consisting of paramagnetic materials. With minor fluctuations, the susceptibility remains relatively constant from base to surface, with the exception of the uppermost two values. The WRA tephra susceptibility peak is subdued due to the presence of the two high surface values. The surface values are probably affected by the extremely high water content of the sediments at this point, and the fact that the catchment is known to have been burnt over in the Takhini burn of 1958. The influx of burned soils into the lake system is a likely explanation for the extremely high surface susceptibility.

6.3.2.4 Water content

The water content values are extremely high in Dragonfly Lake. The basal material rapidly increases from 70 to 90% within the first 5cm of the core (684-679cm). After this initial increase, the water content values remain stable up-core to ~310cm, where there is a dramatic and short-lived reduction in water content to ~55% associated with the deposition of the WRA at this point. After this event, the water content increases again to ~90%, increasing further to >99% in the very top sediments.

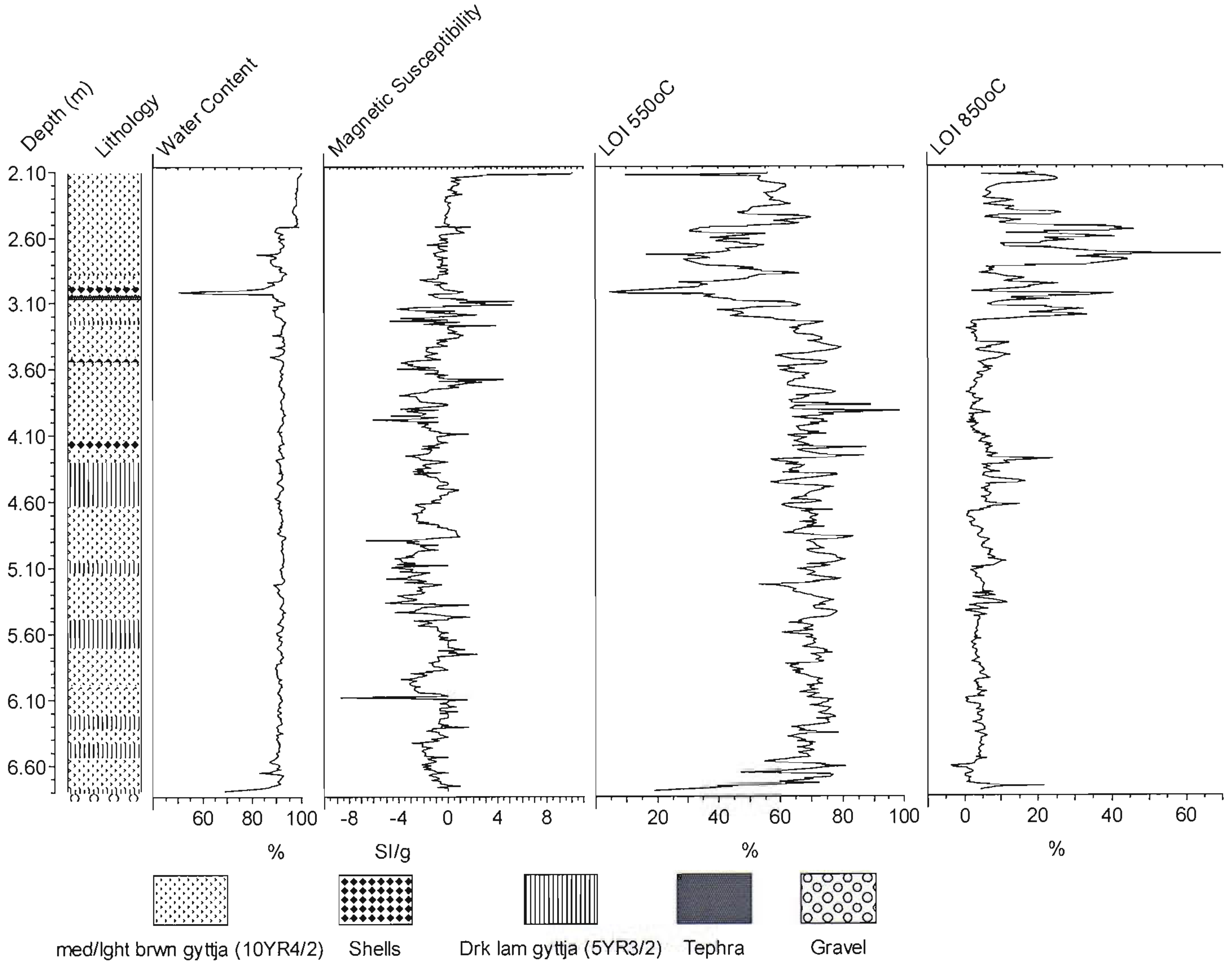


Figure 6.11: Dragonfly Lake, core A, showing stratigraphy, magnetic susceptibility, water, organic and carbonate content. The x-axis of the magnetic susceptibility curve has been truncated to 10 SI units/g in order to observe the fluctuations contained within the main body of the core. The top-most 2 magnetic susceptibility values were 76 (212cm) and 111 SI/g (211cm).

6.3.2.5 Pollen Diagram

Pollen analysis was conducted at 10cm resolution, to construct a general overview of the vegetation history of the area surrounding Dragonfly Lake. The pollen percentage diagram (Figure 6.12) and the pollen influx diagram (Figure 6.13) are presented below for key taxa. Full pollen diagrams are included in Appendix 1.

Zone DFLA-1 (684-635cm) is similar to other regional deciduous zones (for example Cwynar, 1988). Although this zone is relatively short in this record, it is possible to identify typical deciduous zone characteristics, such as the dominance of *Betula* pollen, in association with significant *Salix* and *Alnus* percentages. Abundances of *Betula, Alnus* and *Artemisia* are highest in the deciduous zone, and rapidly decline through the transition to remain low for the remainder of the record.

The pollen assemblage zone DFLA-2 (635-450cm) is characterised by the rapid increase in *Picea* pollen abundance, and associated decline in *Betula* percentages. *Alnus* abundance also rapidly decreases at the transition from the deciduous to spruce zones.

Zone DFLA-3 (450-285cm) is demarcated by the appearance of *Pinus* stomata in the record (see section 6.3.2.5), associated with the first appearance of *Pinus* pollen, albeit at very low levels. The proportion of *Betula* increases into this zone; however, the total number of grains deposited in the sediment remains low.

The rapid increase in *Pinus* pollen (285-211cm) characterises pollen zone DFLA-4, coincident with a decrease in *Picea* pollen proportions.

The record is also characterised by the continuous presence of high proportions of *Pediastrum* nets. *Pediastrum* is an aquatic plant that is thought to be indicative of shallow, and/or warm waters (Anderson & Brubaker, 1994)

6.3.2.6 Stomata Analysis

The stomatal analysis for Dragonfly was conducted in conjunction with the pollen analysis. *Pinus* stomata were found coincident with the start of the “pine tail” (Figure 6.12), and not below, indicating that pine was present in the catchment even while the pollen percentages were relatively low (<1%). Other coniferous stomata not identified as pine, were present in the lower half of the core, during the period of spruce dominance, and suggest that pine stomata are genuinely absent from the record, rather than that they failed to be preserved for some reason.

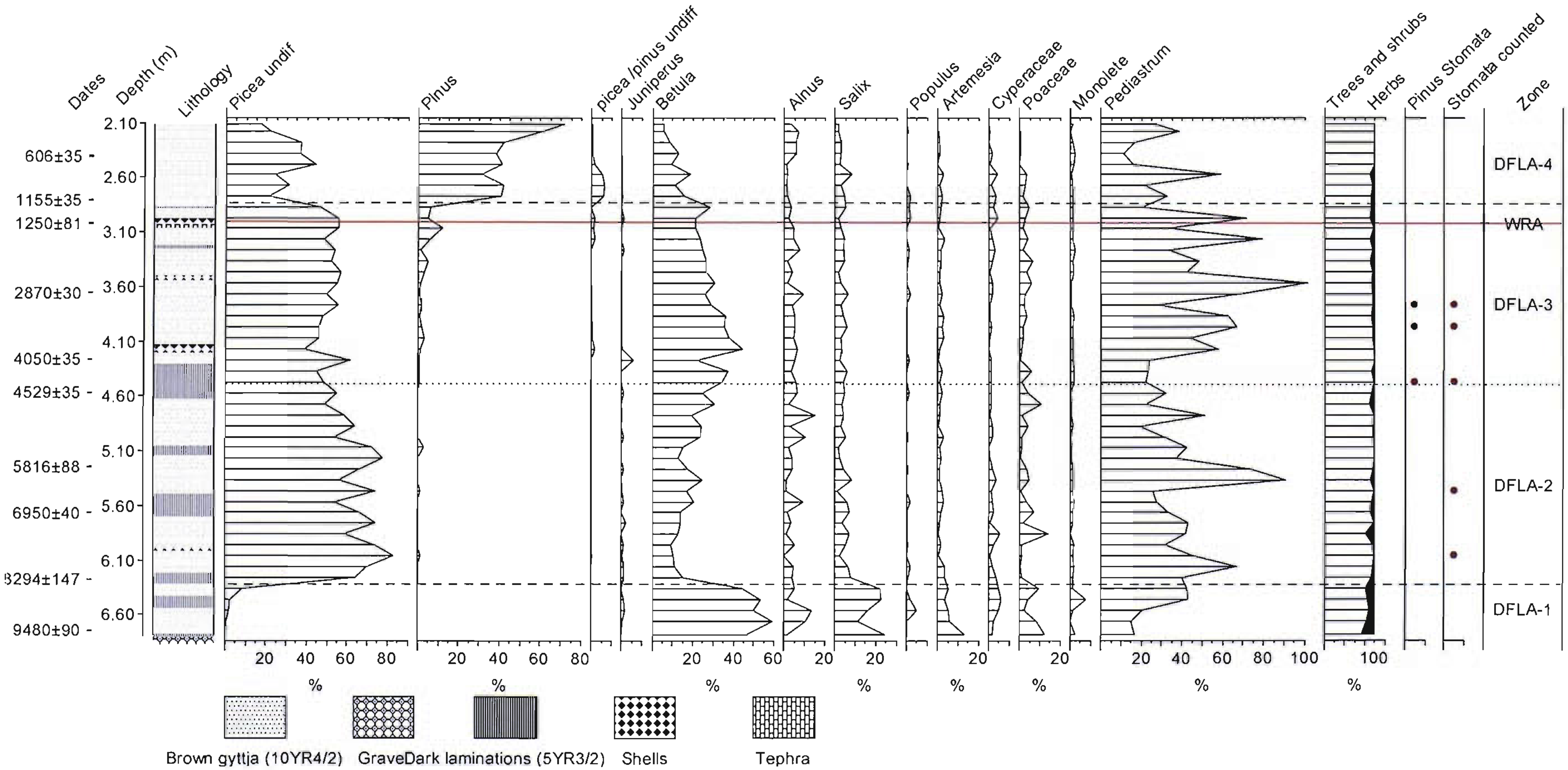


Figure 6.12: Pollen percentage diagram for Dragonfly Lake, core A. Analyst: LMFS and KJG (spruce determination).

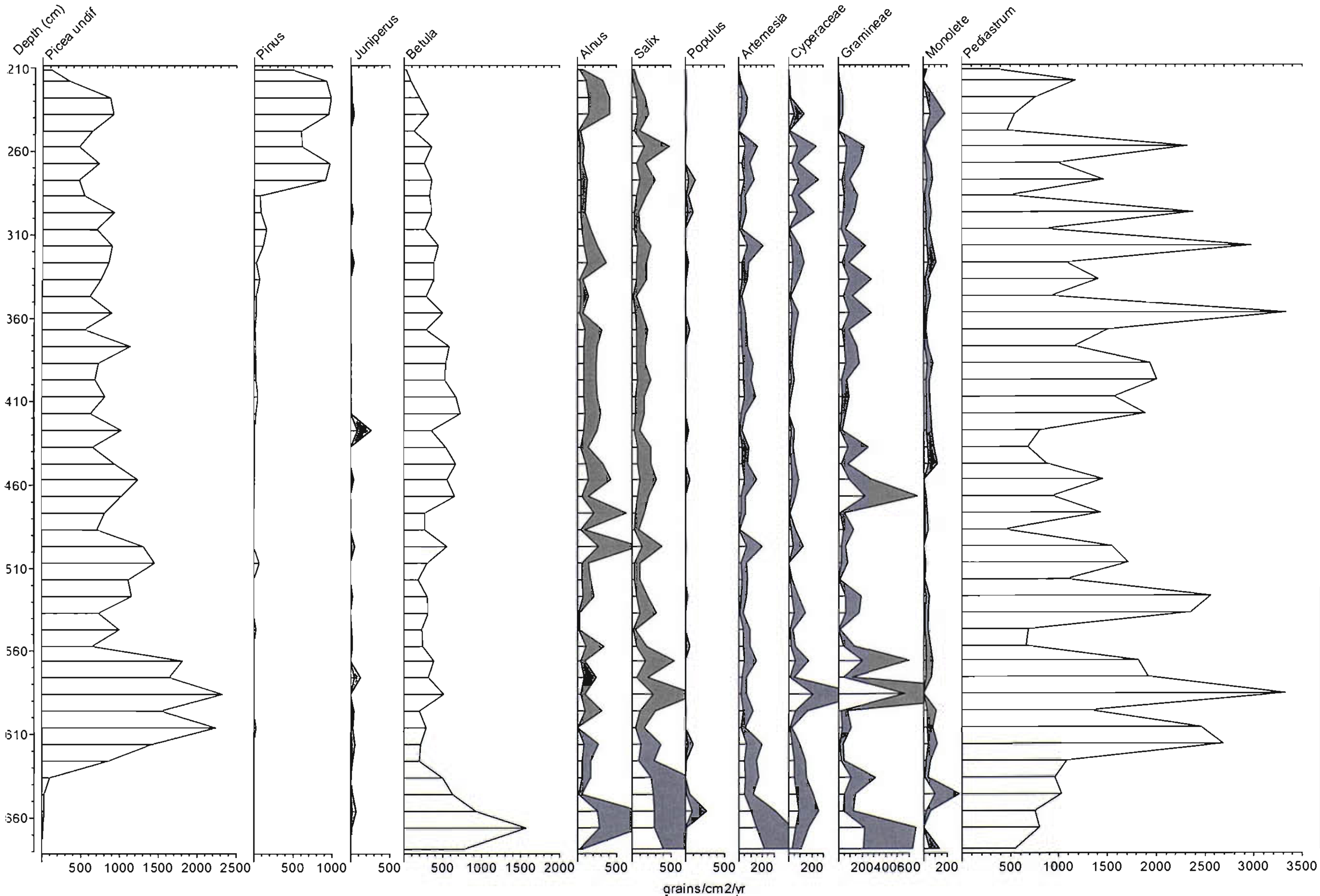


Figure 6.13: Pollen influx diagram for Dragonfly Lake, Core A. Grey shading indicates a 5x exaggeration factor applied.

6.3.2.7 Vegetation History

The earliest zone, DFLA-1, contains a pollen assemblage consistent with regional deciduous zone records (Terasmae, 1966; Cwynar, 1988). However, as at Salmo Lake, the record is very short in length. This is despite Dragonfly lake being the most northerly site, which might be expected to have been deglaciated relatively early. The influx values suggest an established *Betula* dominated community, with probably a sparse cover of *Salix*, grasses, sedges and *Artemisia.*

The transition to DFLA-2 at ~9600 cal yrs BP displays an abrupt increase in *Picea* pollen abundance, which reflects a spruce-dominated forest. The *Picea* present at this time was mostly *P. glauca* (Gallagher, 2005).*Pediastrum* is present in relatively high abundances throughout the record; however, the transition is marked by a definite increase in *Pediastrum* influx. The significance of *Pediastrum* is not entirely clear, but many *Pediastrum* species prefer lakes with high conductivity and pH (Crisman, 1982), suggesting it may reflect episodes of increased sediment and/or nutrient in-wash. On the modern landscape air-fall loess can be rich in carbonates and other cations (Reiger *et al.*, 1979), and increases in loess deposition may have led to *Pediastrum* blooms, although other factors, including summer water temperature and lake productivity, may also be important.

By ~5000 cal yrs BP (zone DFLA-3), *Picea* is represented by increasing proportions of *P. mariana* pollen grains (Gallagher, 2005), as at Salmo Lake, if slightly later.. This switch in spruce type is coincident with the first appearance of *Pinus* stomata within the catchment. The stomata likely represent isolated trees as discussed above (6.3.1.7).

The abrupt increase in *Pinus* pollen associated with the transition to DFLA-4 at ~1100 cal yrs BP, more likely indicates the regional expansion of pine. The pine zone is relatively short in this record compared with the other Yukon lakes studied, and the difference in timing likely reflects the migration northward through the Territory in the last ~1000 years. Dragonfly Lake is the most northern of the Yukon sites, and is relatively close to the current northern limit of the geographic range of lodgepole pine.

6.3.3 Haircut Lake

6.3.3.1 Stratigraphy

Haircut Lake yielded 153cm of relatively dry, non-elastic sediment cored in a water depth of 865cm. The lake was cored to refusal; however, the transition from sediment to gravel was not found, and a full Holocene record was not expected. The core lithology is represented in Figure 6.14.

Overall the core is made up of dark gyttja, with the mid-section of the core displaying regular laminations (983-904cm), and substantial silt bands.

6.3.3.2 Loss on Ignition

Organic content is highly variable throughout the core (Figure 6.14), but average values are relatively high at *c.* 35%. There are three main areas with decreased levels of organic content, which coincide with elevated magnetic susceptibility values, and suppressed water content indicating these are associated with transient in-wash events in the catchment identified as silt bands in the core stratigraphy.

Carbonate content is relatively steady throughout the core (Figure 6.14), with average values quite low at *c.* 5%. There appears to be very slightly increased carbonate production at the core top, above the WRA tephra (897cm), where values are sustained above 5% until the surface. It is thought that this is a result of a shift in the bicarbonate equilibrium brought about by the deposition of the ash layer into the system, similar to that observed in Dragonfly Lake.

6.3.3.3 Magnetic Susceptibility

The magnetic susceptibility record for this core is highly variable (Figure 6.14), with moderate to strongly positive values. The core consists mostly of canted antiferromagnetic material with a minor component of paramagnetic materials. The areas with highly elevated susceptibility values (1002cm, 906cm and the surface above 870cm) are coincident with major, or multiple-minor, silt bands, indicative of transient erosional events in the catchment, with the values being comparable to that of burned soils, igneous or basic rocks (Dearing, 1999). The WRA tephra is identified by the peak at 896-897cm.

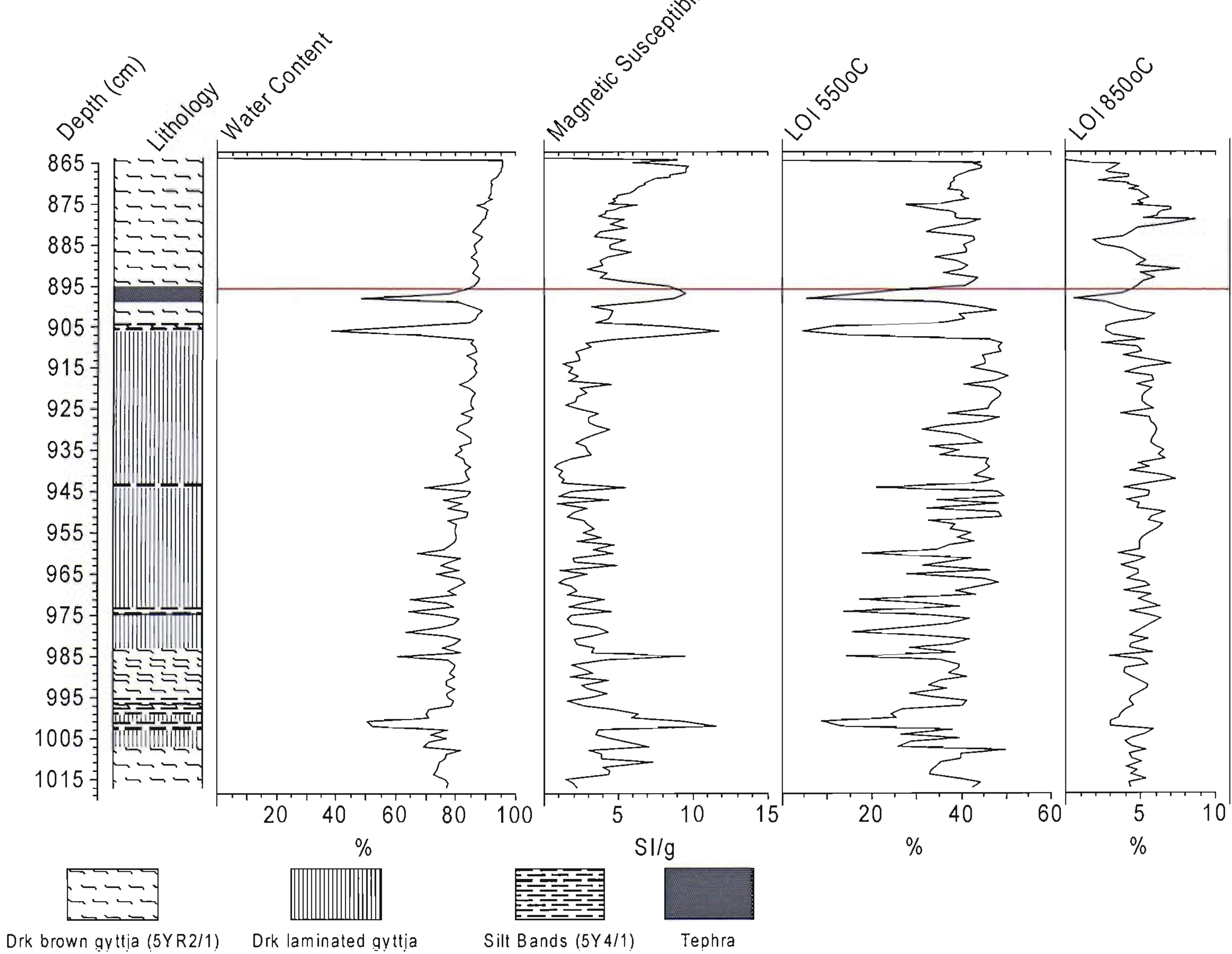


Figure 6.14: Haircut Lake, core A, showing stratigraphy, water content, magnetic susceptibility, organic and carbonate content. Red line denotes the position of the WRA tephra.

6.3.3.4 Water Content

The water content curve shows an overall increasing trend up-core, from~70-95%. There are periods of dramatically reduced water content, mostly associated with silt bands in the sediment stratigraphy and peaks in magnetic susceptibility. The most extreme reduction in water content is at 905cm (~45%) immediately prior to the WRA deposit which also reduces the water content to ~50%.

6.3.3.5 Pollen Diagrams

Haircut Lake pollen analysis was conducted at 5cm sampling resolution due to the length of the core, and the basal age (Figures 6.15 and 5.6). A full Holocene record was not obtained from this lake. The pollen diagram has been split into three zones.

The record starts in pollen assemblage zone HCLA-1 (1017-990cm), typified by the high proportions of *Picea* pollen, in addition to relatively high abundances of *Betula*, Poaceae and Cyperaceae. *Picea* had presumably been established in this catchment for some time prior to the beginning of this record, as the regional pattern of spruce establishment dates the spruce rise to *c.* 10,000 cal yrs BP.

Zone HCLA-2 (990-940cm) is demarcated by the appearance of *Pinus* stomata in the record. The stomata first appear in conjunction with the initial influx of *Pinus* pollen. There is a slight increase in the proportion of *Betula* pollen, and a decrease in Poaceae and *Picea* pollen in this zone. The influx diagram (Figure 6.16) indicates a probable shift in sedimentation rate at 975cm, highlighted by the peak of influx for most species. This alteration is not reflected in the pollen percentage diagram (Figure 6.15). A sample for radiocarbon determination is currently being processed near this event, to further inform the record (see section 5.3.6).

The transition to zone HCLA-3 (940-864cm) is characterized by a rapid increase in the proportion of *Pinus* pollen, and coincident decreases in *Picea* and *Betula* pollen. The WRA tephra is located at 897cm, and does not appear to impact the influx rates in this record.

6.3.3.6 Stomata Analysis

The stomatal analysis performed on Haircut lake shows that the first appearance of *Pinus* stomata in the record (at 990cm) is coincident with the first appearance of *Pinus* pollen (Figure 6.15). Stomata were consistently found throughout the “pine-tail” of low *Pinus* abundances, and in every sample analysed in the main pine pollen zone. Stomata found below the "pine-tail" were not identified as *Pinus* stomata, however, they were coniferous stomata.

6.3.3.11 Vegetation History

The earliest zone in this record, HCLA-1, contains a pollen assemblage consistent with an established spruce-dominated forest. As the record does not extend back past ~8000 cal yrs BP, the regional spruce rise at ~10,000 cal yrs BP is not covered. The *Picea* present at this time is mostly *P. glauca* (Gallagher, 2005). *Betula* and *Salix* are also key contributors to the vegetation assemblage at this time. It is probable that the *Salix* component was mostly located in the valley containing the inflowing stream.

By ~6000 cal yrs BP (zone HCLA-2), the first appearance of *Pinus* stomata is recorded within the catchment; however, as discussed previously, it is likely that only isolated trees were present to start. The *Picea* determinations were not made at this site, although it seems probable considering the other records that *P. mariana* would have increased around this time also..

The abrupt increase in *Pinus* pollen marking the transition to HCLA-3 at ~3400 cal yrs BP is probably indicative of the local expansion of pine within the catchment. Due to the very well defined catchment, extended slightly by the inflowing stream, and the deep basin that Haircut Lake sits in, it is likely that the local increase in pine trees occurred over a relatively short period of time. Haircut Lake is the only site with a catchment entirely dominated by LPP on the present landscape. There is little spruce within the catchment today, and it is likely that spruce became increasingly marginalised once pine appeared in the catchment.

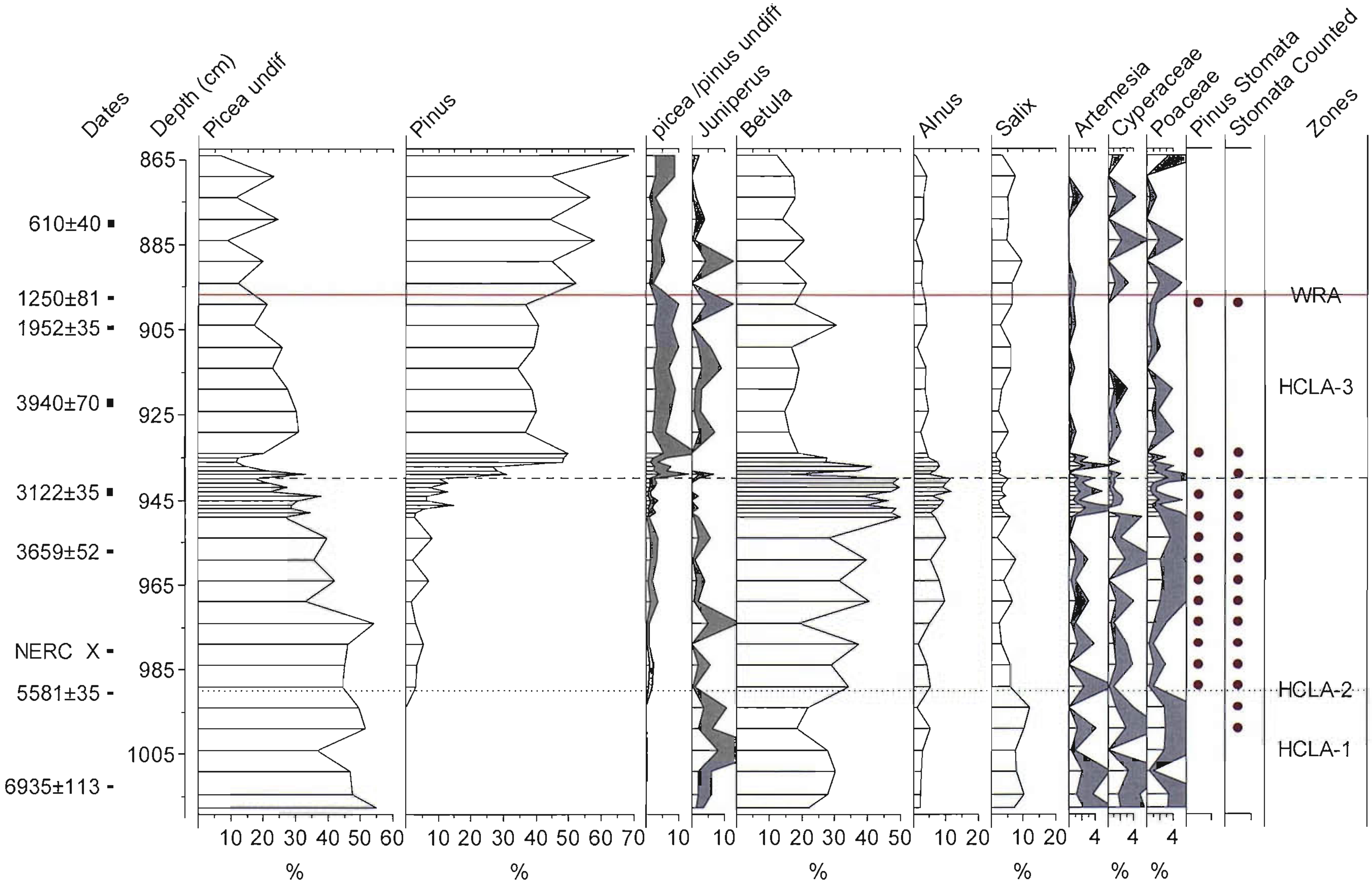


Figure 6.15: Pollen percentage diagram for Haircut Lake, core A. Grey shading indicates a 3x exaggeration factor applied. Analysts: LMFS and JRB.

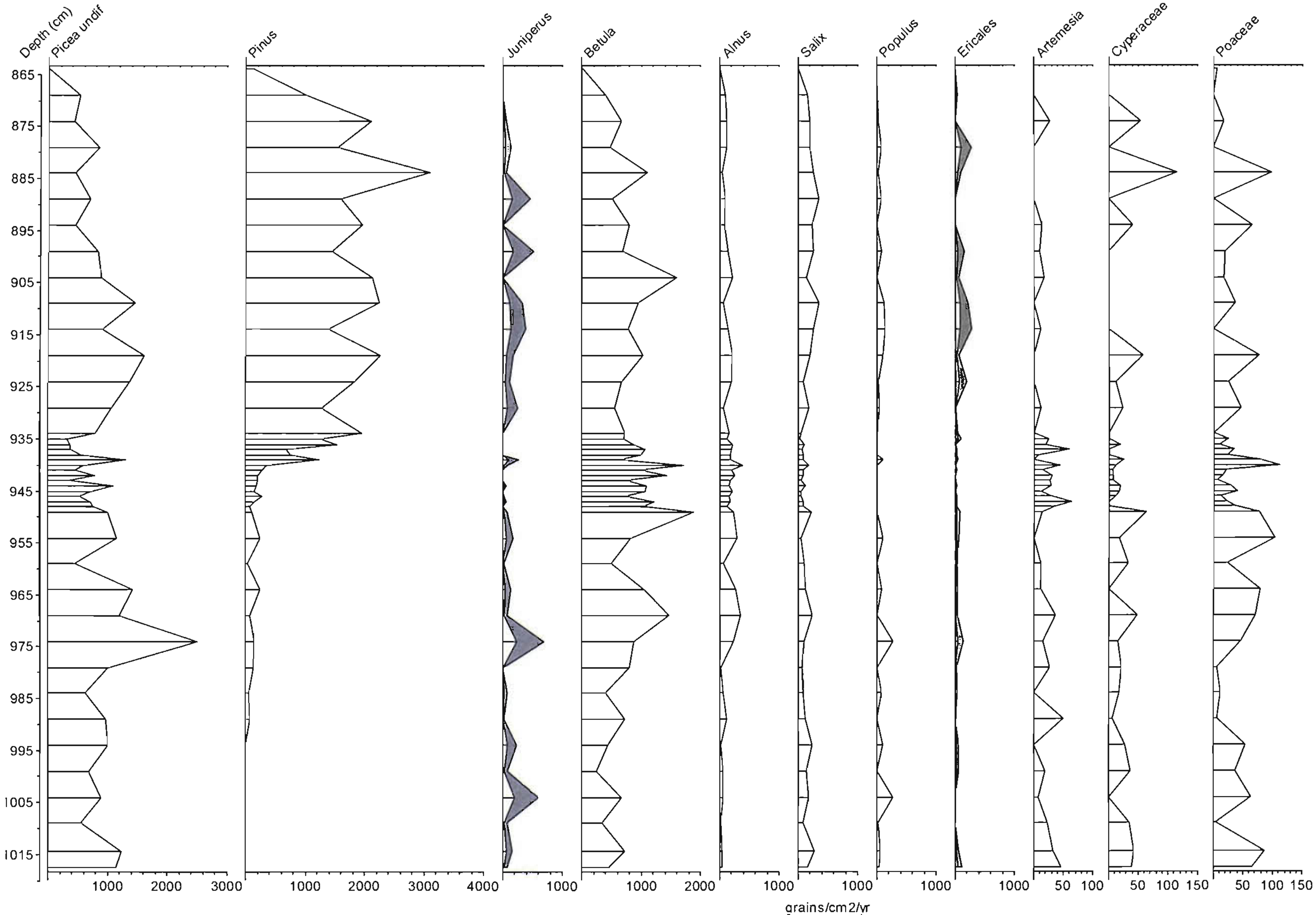


Figure 6.16: Pollen influx diagram for Haircut Lake, core A. Depth is measured from water surface.

6.3.4 Marcella Lake

6.3.4.1. Previous Results and Interpretation

Data from Marcella Lake have been published and interpreted for the reconstruction of Holocene palaeoclimatic variations in the northwest sub-arctic (Anderson *et al.*, 2002; Anderson, 2004; Anderson *et al.*, 2005b; Anderson *et al.*, In Press). The main findings relevant to this project will be outlined here, and the reader is directed to the publications listed for further details.

Lithologic unit 4 is made up of gravel (Figure 6.18). Unit 3c consists of grey silty marl, with relatively high magnetic susceptibility, low organic content and high calcium carbonate content. This unit is thought to consist of reworked shallow lake deposits. The transition to unit 3b is abrupt, with decreased magnetic susceptibility, increased carbon content and highly variable carbonate content. Presence of distinctly laminated sediment structure indicates development of a deeper water lake (Anderson, 2004).

The transition to unit 3a is visibly abrupt, expressed in all of the sedimentary measures (Figure 6.18) and the pollen record (Figure 6.17), and is interpreted as a sedimentary discontinuity, indicating low lake levels (Anderson, 2004). This discontinuity ranges from 429 to 260cm core depth, with the sediments containing disturbed textures, and otherwise rare aquatic mosses, and yielding suspiciously old radiocarbon ages (Anderson *et al.*, 2005b). From ~300-260cm, the pollen curves are largely identical to those of the early spruce zone, and this repetition was taken to indicate the uppermost portion of the displaced material. The radiocarbon ages from depths immediately above and below are statistically identical, and provide a further constraint on the thickness of the displaced material and the timing of the event (~7850 cal yrs BP). Figure 6.18 shows the stratigraphy and bulk sedimentary data after removing the suspect sediment (unit 3a), and adjusting the depths for samples below 260cm.

Unit 2b contains the highest organic carbon content values in the core, and the sediments become gradually more olive brown, with faint laminations. Well preserved micro-scale laminae and a gelatinous texture define unit 2a, with the laminae shown to consist of alternating layers of diatoms and organic matter. This is presumed to account for the decrease in organic carbon content via an increase in the biogenic silica component (Anderson, 2004).

The transition from unit 2a to unit 1 is delineated by the 1.0cm thick WRA deposit. The organic content is slightly lower, while the calcium content increases. These surface sediments are faintly colour banded, bioturbated, uncompacted, gelatinous organic muds.

The pollen record reported in Anderson (2004; 2005b)(Figure 6.17) shows a relatively short pre-spruce zone, dominated by Cyperaceae, *Artemisia* and *Alnus* species, with an abrupt transition to the early spruce zone at ~310cm adjusted (notation of Anderson *et. al.* (2005b)) demarcated by a rapid increase in spruce pollen, and contemporaneous decreases in Cyperaceae and *Artemisia.*

The first appearance of pine is at ~245cm, with the main increase in pine pollen occurring significantly further up-core at ~120cm, coincident with the start of the decline in spruce.

This pollen record is broadly similar to that of Cwynar (1988), who conducted a pollen study on Kettlehole Pond (aka Marcella Lake). Chronological differences between the records exist, which may result from the use of bulk sediments for radiocarbon dating in the earlier study.

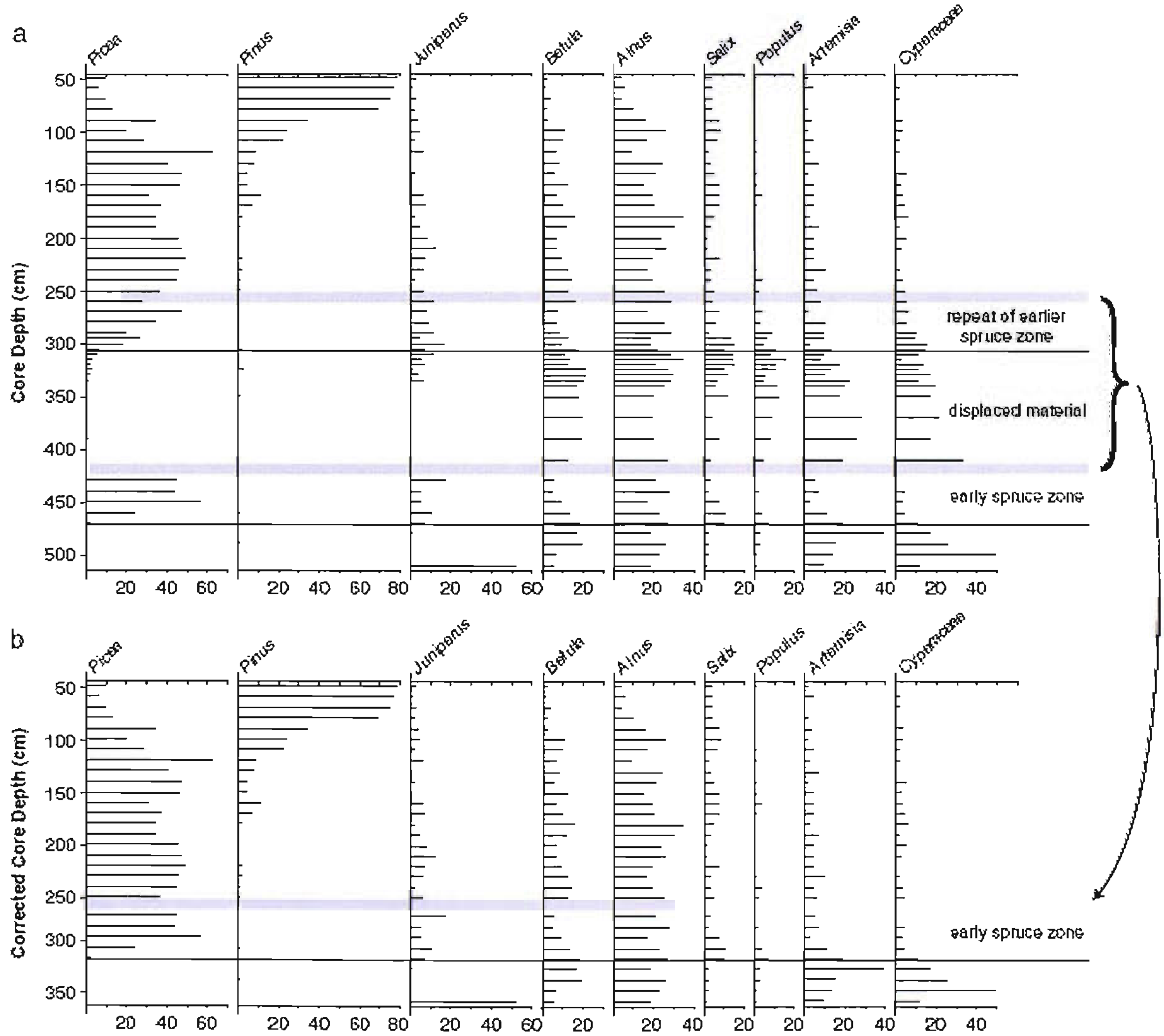


Figure 6.17: Marcella Lake pollen percentages of key taxa with depth for: (a) the whole core without adjustment and (b) after adjustment by removing suspect material between 429 and 260cm depth. Reproduced with permission from L. Anderson (Anderson, 2004).

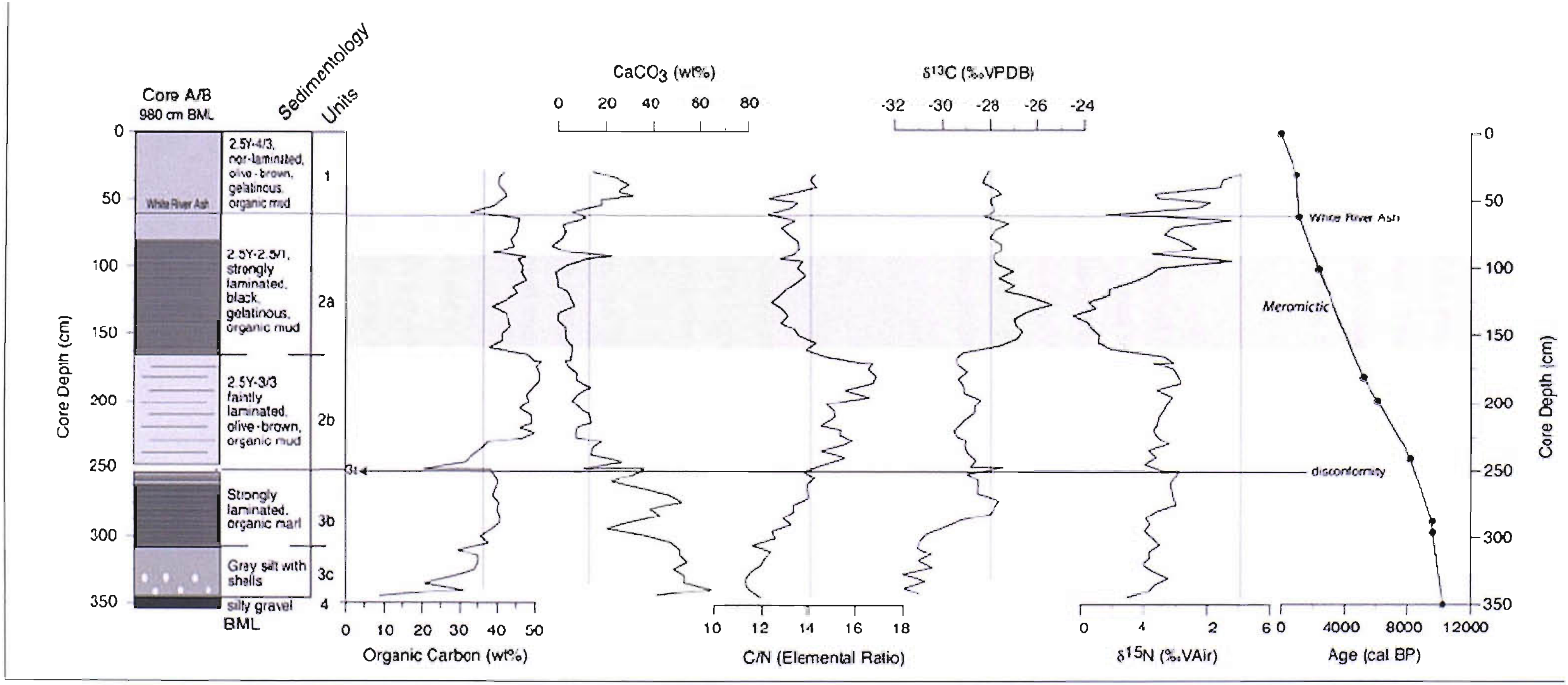


Figure 6.18: Marcella Lake, core A - stratigraphy, organic carbon, calcium carbonate, C/N, δ13C, δ15N and median calibrated radiocarbon ages by depth (reproduced from with permission from L. Anderson (Anderson et al., 2005b)

Anderson's core A is amalgamated with Franklin-Smith's core Z, in order to have a complete Holocene record for the reconstruction of the charcoal record. Below the results of amalgamated core A-Z are presented, and the joined records will subsequently be interpreted as one sequence. The cores were correlated via the LOI and magnetic susceptibility curves for each core, the WRA was used as a stratigraphic pinning point, and radiocarbon determinations supported the correlation (see Appendix 2).

6.3.4.2. Stratigraphy

The lithology of core A-Z remains largely as described for core A in section 6.3.4.1. The amalgamated core A-Z has the addition of unit 2a, a section of dark brown non-laminated gyttja immediately below the WRA. There is a strong correlation of the organic and carbonate curves about the WRA. Unit 1c consists of olive or light brown organic gyttja, which is interrupted temporarily by unit 1b, a further section of dark brown gyttja. Unit 1c is a reversion to the olive/light brown organic gyttja of unit 1a, which persists to the core top.

6.3.4.3. Loss on Ignition

It should first be noted that the apparent increase in variability in the LOI results above the WRA is due to the higher sampling resolution conducted on core Z (1cm) compared with core A (5cm). For description of the LOI below the WRA the reader is referred to section 6.3.4.1.

After an initial decrease in organic content as a response to the ash deposit, the organic content of the core increases to between 50 and 60%, until approx 30cm from the core surface. The uppermost sediments display an increase in organic content to a maximum of 80%.

The carbonate content of the core generally increases above the WRA ranging between 20 and 40%, with a marked decrease in the uppermost sediments. Marcella lake is strongly stratified, and there is currently little or no carbonate accumulation in the deep hole to the west side of the lake basin (Anderson, Pers comm.), and this area was not cored for this study. Carbonate in Marcella is only preserved above the thermocline, in water depths of less than ~5-7m (Anderson, 2004), and therefore at any point the percentage $CaCO_3$ accumulation is strongly controlled by lake level.

6.3.4.4. Magnetic Susceptibility

The magnetic susceptibility shows very little variation up core after the initial basal peak in susceptibility. The WRA is marked by the high peak at ~87cm, after which the magnetic susceptibility returns to between 0 and 1 SI units. A second smaller peak in susceptibility is noted between 18 and 20cm, and is associated with a discrete silt band.

6.3.4.5. Pollen Diagram

Key pollen taxa for Marcella Lake, core A-Z are represented as percentages (Figure 6.20) and influx (6.21) below. Full pollen records with pollen sums are available in Appendix 1. Below ~330cm the record is dominated by the presence of Cyperaceae, *Artemisia*, *Alnus* and *Betula*, which is characteristic of other regional records as the deciduous zone, and is identified by pollen assemblage zone MLAZ-1. The transition to MLAZ-2 is marked by the decline in Cyperaceae and *Artemisia*, and the abrupt increase in *Picea* pollen after ~330cm.

The pollen assemblage zone MLAZ-2 is marked by the continued high percentages of *Picea* pollen, ranging between 40 and 60%. The zone is sub-divided by the appearance of pine stomata and very low levels of pine pollen (<1%) after ~270cm and is identified as MLAZ-3. Pine percentages increase very gradually up-core, with the main increase in pine pollen at ~145cm marking the transition to MLAZ-4. Pine pollen percentages reach their maximum prior to the WRA deposition. However, above the ash deposit there appears to be a slight alteration in the pollen deposition. *Alnus* pollen declines immediately prior to the WRA deposition, and remains at very low levels to the core surface. There is also a slight increase in *Betula* pollen. The influx diagram (Figure 6.21) indicates an alteration in the influx between ~80 and 60cm depth. This is due to the effect of the WRA deposit on the age-depth model (see section 5.3.7), which increases the sedimentation rate from >20 yrs/cm2 to ~5 yrs/cm2. The effects of this alteration in deposition rate are further heightened by the proximity of this section to the surface of the core. The uppermost sediments experience a dilution-effect due to the relatively uncompacted nature of the surface sediments, and therefore the influx values for all taxa decrease. It should also be noted here that the topmost sample contained very little pollen, and therefore the surface samples should in reality be excluded from the interpretation, or at very least be interpreted with extreme caution.

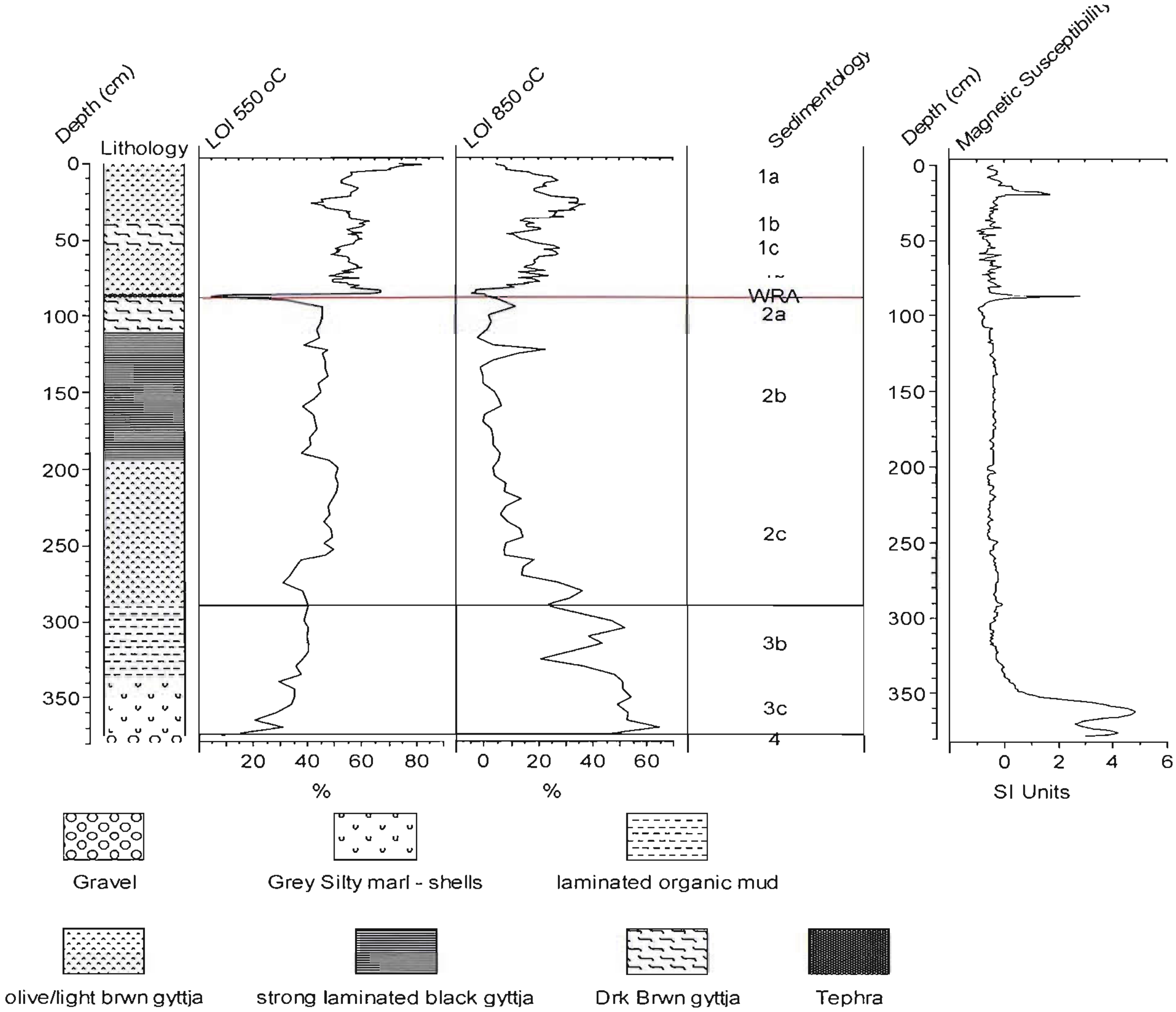


Figure 6.19: Marcella Lake, core A-Z, showing stratigraphy, organic content, carbonate content and magnetic susceptibility. Depth is measured from core surface, and stratigraphic units follow that of Anderson (2004) for ease of comparison..

6.3.4.6. Stomata Analysis

The pine stomata results are presented in Figure 6.20, and it is apparent that the pine stomata are identifiable in layers with very low pine pollen percentages (from 270cm up core). Samples below 270cm were also screened for the presence of stomata; however, no pine stomata were identified.

6.3.4.7 Vegetation History

The earliest zone, MLAZ-1, contains a pollen assemblage consistent with another record produced on this site (Cwynar, 1988); and is considered to be representative of the deciduous zone. The landscape was most likely quite sparsely covered with *Salix*, grasses, sedges and *Artemisia,* with *Betula* and *Alnus* the dominant species.

The transition to MLAZ-2 at ~9800 cal yrs BP shows an abrupt increase in *Picea* and abundance, indicating a relatively rapid transition from a *Betula* dominated shrub-land, to a coniferous dominated forest. The *Picea* present at this time is mostly *P. glauca.*

By ~6000 cal yrs BP (zone MLAZ-3), the *Picea* assemblage has shifted to increased dominance by *P. mariana* (Gallagher, 2005). Coincident with this is the first appearance of *Pinus* stomata within the catchment. Stomata indicate local presence of pine trees, however it is unlikely trees were present in significant numbers at this time. and today, lodgepole pine is only found on the south side of the lake.

The transition to MLAZ-4 at ~3000 cal yrs BP is characterised by the abrupt increase in *Pinus* pollen, which more likely indicates the regional expansion of pine populations contributing pollen to the catchment, and is coincident with another period of increased effective moisture in the interior of the Yukon Territory (Anderson *et al.*, 2005a; Anderson *et al.*, In Press).

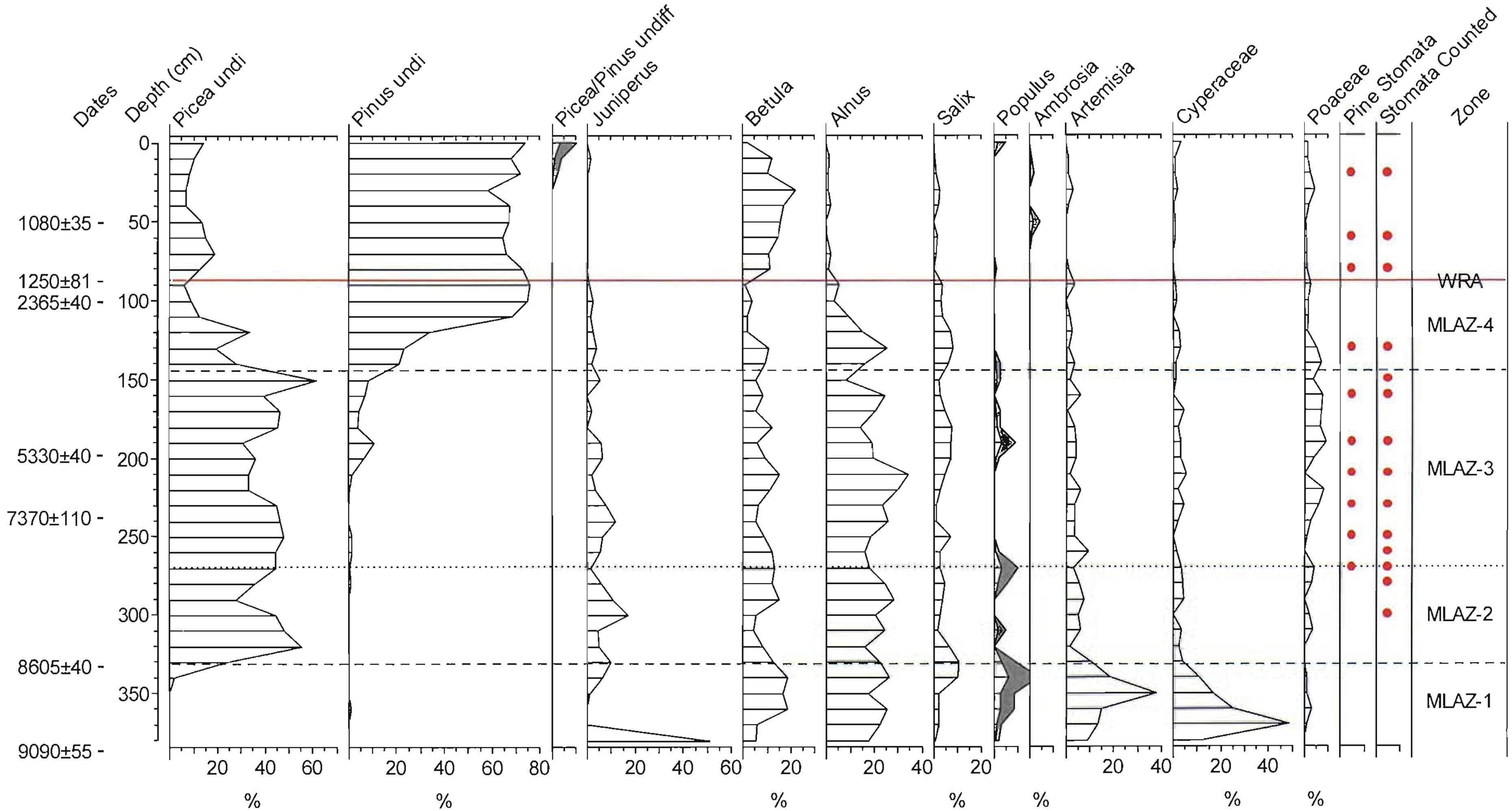


Figure 6.20: Marcella Lake pollen percentages for key taxa for amalgamated core A-Z. Depths are reported from ore surface. For full pollen diagram please refer to Appendix 1. Grey shading indicates a 3x exaggeration factor. Analysts: LMFS, APK , JRB and KJG for spruce determinations..

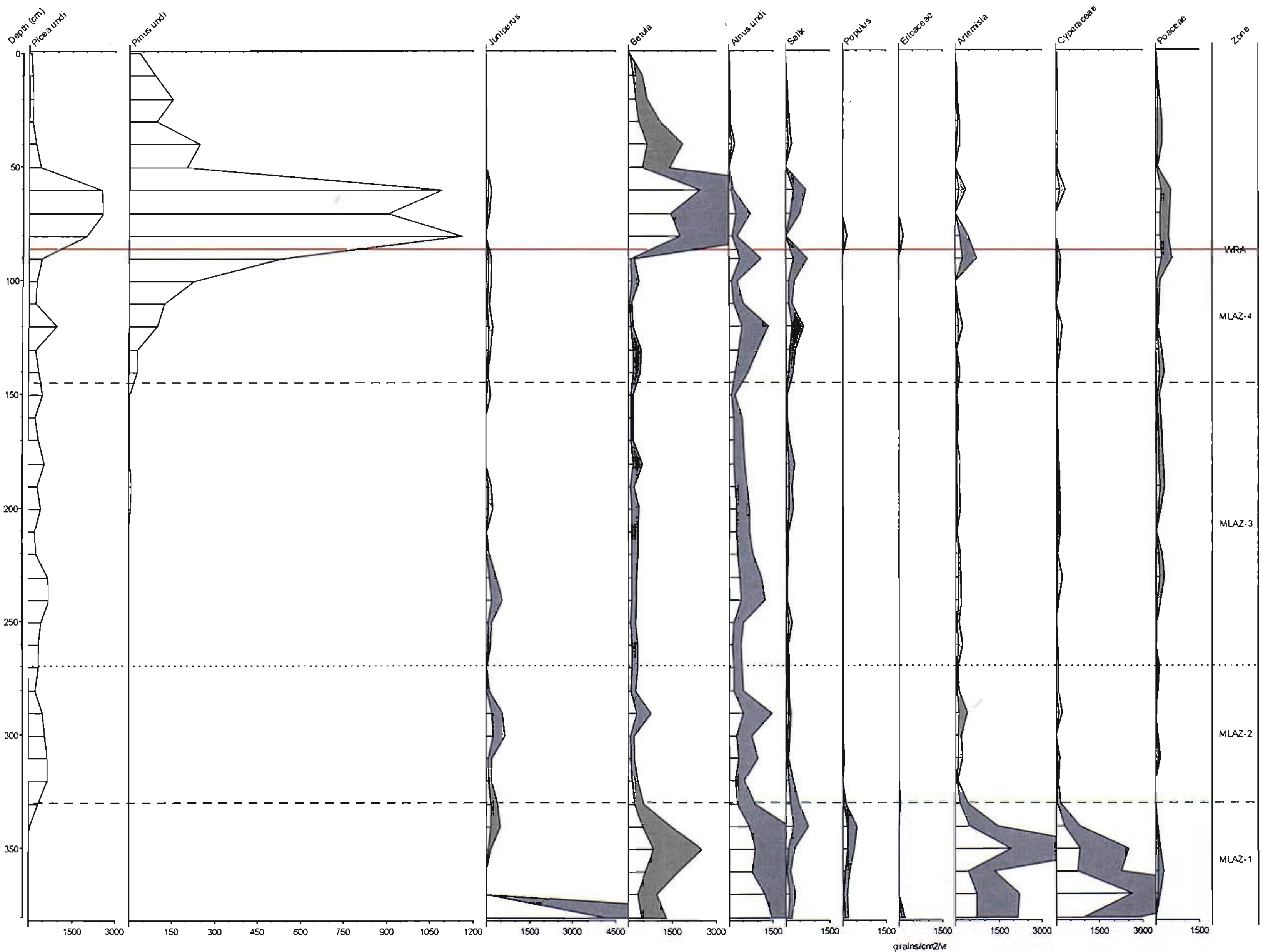


Figure 6.21: Marcella Lake pollen influx diagram for amalgamated core A-Z. Depths reported are from core surface. Grey shading indicates a 3x exaggeration.

6.3.5 Jellybean Lake

6.3.5.1 Previous Results and Interpretation

Jellybean Lake sediment and isotope data have been published and interpreted for the reconstruction of Holocene palaeoclimatic variations in the northwest sub-arctic (Anderson, 2004; Anderson *et al.*, 2005a). As with Marcella Lake, the main findings relevant to this project will be outlined here, and the reader is directed to the publications listed for further details.

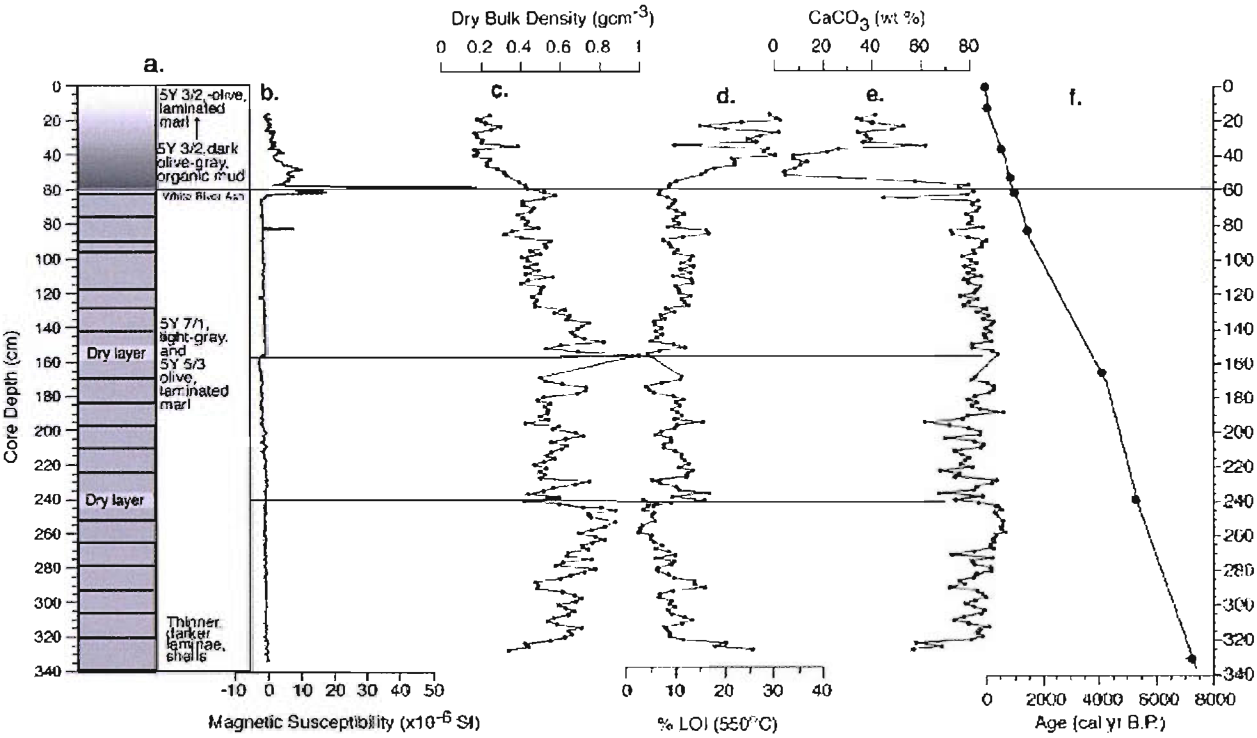


Figure 6.22: Jellybean Lake core C – stratigraphy, magnetic susceptibility, dry bulk density, organic content and median-calibrated ages plotted against depth. Reproduced with permission from L. Anderson (Anderson et al., 2005a)

The sedimentary results (Figure 6.22) show that Jellybean Lake has generally low magnetic susceptibility, with the exception of the WRA at 60cm. Organic content ranges between 5 and 15%, and the dry bulk density is generally low. The carbonate values range between 80 and 85% for most of the core, changing to between 5 and 10% above the WRA. Sediments overlying the WRA are distinctly darker (Anderson, 2004). Carbonate production in Jellybean is not affected by changes in lake level as there is no thermocline (Anderson, Pers comm.). Therefore, it is more likely to be changes in the lake pH, and/or dissolved CO_2 and subsequent shifts in bicarbonate-carbonate equilibrium in response to the ash deposition, that alters the $CaCO_3$. If the WRA altered the pH of the lake, shifting the carbonate-bicarbonate equilibria, an increased proportion of organic matter and a decreased proportion of $CaCO_3$ would be preserved in the mud. Organic and carbonate are usually inversely related as both are a product of within-lake algal productivity at this site.

The basal date on this core is ~7000 cal yrs BP, and a full Holocene record is not available. There are no pollen data for this site at present; however, regional vegetation history is well established (see above). Jellybean Lake provides a high-resolution isotope record of climatic variability, which will inform the fire record at this site, and provide information on the regional precipitation pattern. Results indicate that the oxygen isotope composition of the water reflects the composition of mean-annual precipitation (Anderson *et al.*, 2005a), and changes in North Pacific atmospheric circulation inferred from the down-core record corresponds with late Holocene glacial advances in the St. Elias Mountains (Denton & Karlen, 1977; Calkin, 1988), changes in North Pacific salmon abundance (Finney *et al.*, 2000; Finney *et al.*, 2002), and shifts in atmospheric circulation over the Beaufort Sea (Dyke & Savelle, 2000).

Summary of Core suitability for Charcoal Analysis

Through the establishment of vegetation histories, and analysis of physical sediment properties all of the cores, with the exception of the discontinuity section in Marcella core A (Anderson, 2004), have demonstrated their integrity. The vegetation histories are consistent with the regional records discussed in Chapter 2, and supported by the chronologies presented in Chapter 5 it can be concluded that all sites are suitable for charcoal analysis, and a reliable representation of local burning events can be expected.

Chapter 7. Charcoal Analysis

7.1 Introduction to CHAPS

Charcoal abundances were converted to charcoal concentration (particles/cm^3), from which a charcoal accumulation rate (CHAR) (charcoal particles/cm2/yr) time-series was created using CHAPS (**CH**arcoal **A**nalysis **P**rogram**S**), a Fortran-77 based program (Long *et al.*, 1998). The program interpolates the charcoal concentration values and sedimentation rates (cm/yr) to a constant time interval. The time interval for all sites was set at 10 years as this represented the shortest deposition time of all the lake records. This strategy preserves the total number of particles accumulated over time and allows for presentation of the data at equally spaced time intervals (Long *et al.*, 1998; Mohr *et al.*, 2000; Brunelle-Daines, 2002). Long *et al.* (1998) describe the rationale behind the decomposition technique, which has been used by many other studies reconstructing long-term fire histories (see for example: Hallett & Walker, 2000; Millspaugh *et al.*, 2000; Mohr *et al.*, 2000; Lynch *et al.*, 2002; Gavin *et al.*, 2003; Hallett *et al.*, 2003; Brunelle *et al.*, 2005).

The objective of CHAPS is to decompose the charcoal influx time series into two components: i) a background component, which represents trends in the charcoal data that are most likely related to processes of production and deposition, and which is slowly varying over time; and ii) a peaks component, which represents the CHAR above the background level. This is rapidly varying over time, representing short-term variations in charcoal influx, of which the largest positively correlated deviations are registered as fire episodes (Long *et al.*, 1998). As charcoal production varies over time (for example due to changes in fuel availability) and charcoal inputs can alter if the amount of charcoal entering the lake from terrestrial or littoral sediment storage changes, the background component reflects changes in vegetation (fuel availability), slope-wash, and surficial delivery to the lake in addition to sediment deposition within the lake itself (Bradbury, 1996; Whitlock & Millspaugh, 1996).

The term 'fire episode' differs from that of 'fire event'. In dendrochronological studies, a fire event is defined as either a single fire, or a series of fires within a defined area at a particular point in time, whereas a fire episode describes fire data that do not have annual resolution, or which have the potential to reflect several years of fire events (Barrett *et al.*, 1997). As the majority of sedimentary records of fire do not have annual resolution (with the exception of annually laminated sites), and charcoal peaks may represent more than one event, "fire episode" is determined as an appropriate term to describe the sedimentary charcoal record of fire occurrence (Brunelle-Daines, 2002; Brunelle & Whitlock, 2003).

A fire episode is identified when the CHAR exceeds the background component by a predefined threshold ratio. The threshold ratio is assigned specifically for each site in relation to the influence of site characteristics (basin size, shape and hydrology, vegetation history, and surrounding topography for example), and is determined by the point at which the number of fire episode identifications becomes insensitive to further increases in the threshold ratio (Gavin *et al.*, In Press). As no tree-ring data are available at present to calibrate the charcoal record with known fires, the charcoal analyses were informed by the mean fire return interval for modern forest types (for example Viereck, 1973; Van Cleve *et al.*, 1986; Johnson, 1992), as a "reality-check" on the model outputs.

7.2 CHAPS parameters

In addition to the user-defined parameters described below, CHAPS also calculates time series of various summary statistics such as the local mean number of peaks per unit time, or the time since the last peak. In relation to peak identification, CHAPS provides information on the point at which a peak exceeds the assigned threshold (peak start), and when it passes back below the threshold (peak end). The peak start point is used to identify fire episodes, and as such may be offset from the peak maxima on the diagrams. For core sections where the sedimentation rate is quite high and the temporal resolution is low (e.g. >50 years/sample), charcoal peaks have a tendency to become broad due to the procedures employed in CHAPS, and the displacement caused by this can be quite marked in places.

7.2.1 Moving Average

In order for a data series to be decomposed using CHAPS, the user has to define two key parameters which affect the model output. The first parameter specifies the width of the temporal window used to calculate the locally weighted mean, which ultimately sets the smoothness of the fitted trend line and the resulting local mean. This smoothness is especially important when analysing transitions in a time series environments, as it defines the background above which the threshold for peak identification will be set. If the mean is too smoothed, any sudden alterations in the mean (for example during vegetation transitions) would be obscured by the mean changing over much longer periods, and this would lead to false-negative peak identification. If the mean is too reactive to the influx data, it would be probable for false-positive peak identifications to be made, as the fitted trend ultimately follows the influx pattern and identifies every spike in influx values. In this study, the data have been normalised and interpolated to a constant 10-year time interval, and the window width was set to 100 samples, creating a moving average of 1000 years. This window width is considered to be suitable to characterise an established regime which is in equilibrium with the vegetation and climate (Brunelle-Daines, 2002).

7.2.2 Threshold Determination

The second parameter to be defined is the threshold used to identify a peak in the charcoal influx, above the background component. As previously stated this should be defined on a site-specific basis due to individual catchment characteristics influencing the charcoal background component. If a CHAR record is accurately recording fires then the peak component should comprise a small population of high values, which represent local fires, and a large population of low values, which represent distant fires, analytical noise and charcoal redeposition either within the watershed, or the lake (Clark *et al.*, 1996). A residual CHAR record was produced for each site by subtracting the running mean (as a function of background/noise in the data) from the influx. The resulting frequency distribution was plotted (example of which is presented for Salmo Lake in Figure 7.1), and used to inform the evaluation of appropriate thresholds. The threshold was ultimately defined using a method described by Gavin *et al.* (In Press), following the general framework of Clark *et al.,* (1996) and Long *et al.,* (1998) (see Chapter 3), and was assigned to the value where further increases in the threshold produced no further effect on the number of fires identified in the record.

As an example, the threshold determination process for Salmo Lake is described here (Figure 7.1). The record was split in accordance with the pollen assemblage zones presented in Chapter 6. Each zone was analysed with systematically increasing thresholds. The final threshold was assigned to the highest plateau value, which produced a realistic estimation of peak frequency. For Salmo Lake, the influx values in the deciduous zone are very low, but the peaks above the background are relatively large in magnitude. When the threshold was increased, three plateaus were achieved (Figure 7.1a) at 1.1-1.2x, 1.6-1.7x and 1.8-2.4x the background. If the threshold is assigned to the highest range, only the 2 largest peaks are identified. Although it may be considered the most conservative approach to identifying fire episodes, it produces a mean peak frequency in excess of 1100 years. It is highly unlikely that this is truly representative of the regional fire regime, even in the early Holocene. The small plateau between 1.6 and

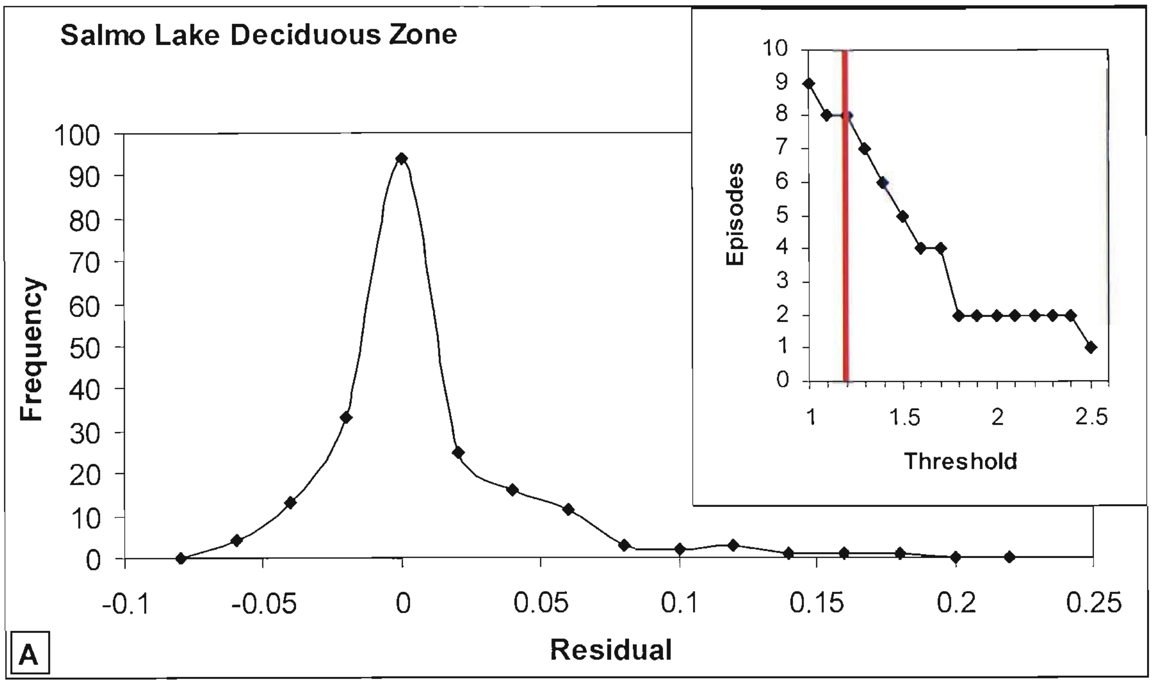
1.7x the background still produces an unlikely peak frequency of ~550 years using the four highest peaks.

Other studies covering the post-glacial periods have utilised low thresholds (1.1-1.15x) to reconstruct fire regimes (Brunelle-Daines, 2002; Brunelle & Whitlock, 2003; Brunelle *et al.*, 2005), using information on the mean fire return interval of modern forest types as a template. It is for these reasons that the threshold of 1.2x was assigned to the deciduous zone at Salmo Lake, producing an estimation of fire return interval at ~275 years (identifying 8 episodes in ~2000 years).

The remaining three fossil pollen assemblage zones (spruce, stomata and pine) produced threshold curves which displayed a low value plateau (between 1.1-1.2x), and a high value plateau where only the highest peaks were recognised (between 2.0-2.6x). These three sections of the record display higher mean influx values (0.5-1.25 particles/cm2/yr), and increased numbers of peaks than the deciduous zone, with a variety of peak magnitudes. It was considered un-representative to assign the threshold to the high value as it produced a peak frequency in the range of ~600-1200 years, considering the modern fire return frequency (FRF) of 100-120 years (Barrett & Arno, 1991; Johnson, 1992). Although modern estimates are not representative of the exact climate and/or vegetation assemblages found in the early- to mid-Holocene, they can and should be used as a guide, especially for the late-Holocene zones.

Assigning the low value threshold produced a peak frequency between ~115-170 years, which is considered to be more representative than >600 years. Analysis of macro charcoal produces a background with relatively little noise from regional sources of charcoal, and therefore the background component which is present, and above which the threshold is assigned, should encompass the majority of any remaining analytical noise. Thus the threshold is set to represent peaks significantly above the background that will likely represent fire episodes. If we accept that once the background component is subtracted from the record the positive residuals that remain are representative of the fire regime, then assigning a threshold above this residual, even a low one is a robust method of identifying fire episodes . Thresholds as low as 0.8-0.9x have been used in the published literature (e.g. Gavin *et al.*, In Press) and have demonstrated sufficient robustness in reconstructing fire regimes.

This process and rationale was applied to every zone at each site to assign a threshold; for brevity the details are not included here.



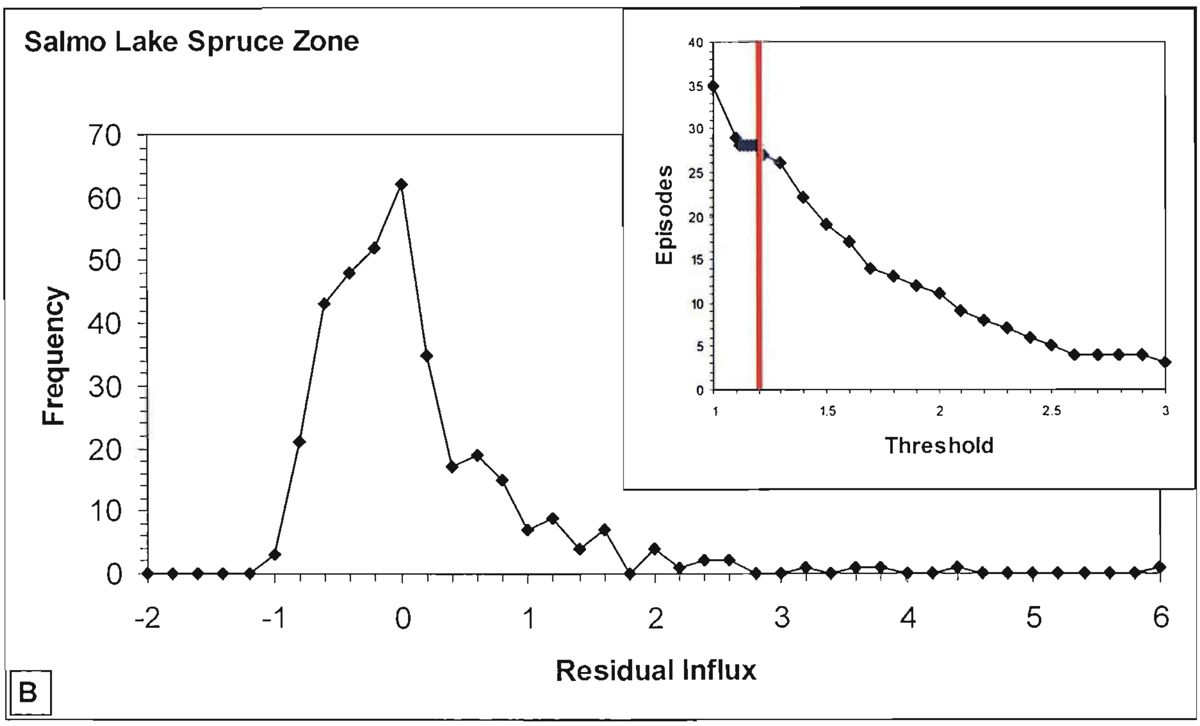
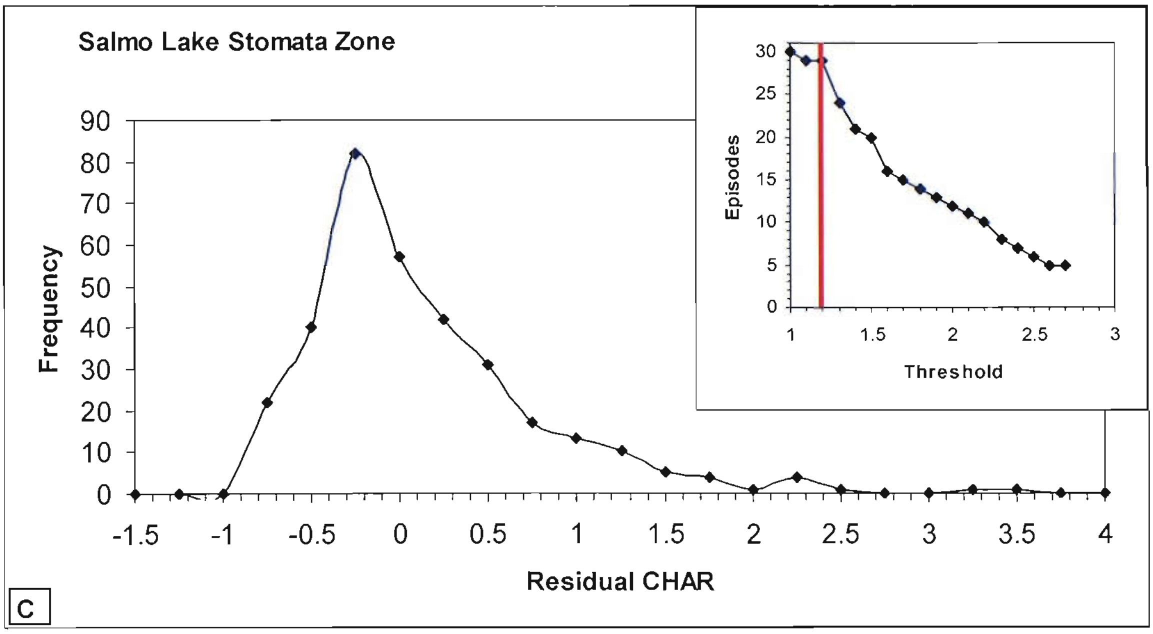


Figure 7.1: Frequency distribution of the CHAR residual for the a) deciduous b) spruce c) stomata and d) pine zones of Salmo Lake. Inset in each shows the sensitivity of the number of CHAR peaks detected using a range of threshold values. The red line shows the threshold chosen to identify charcoal peaks most likely due to local fire episodes (continued overleaf).



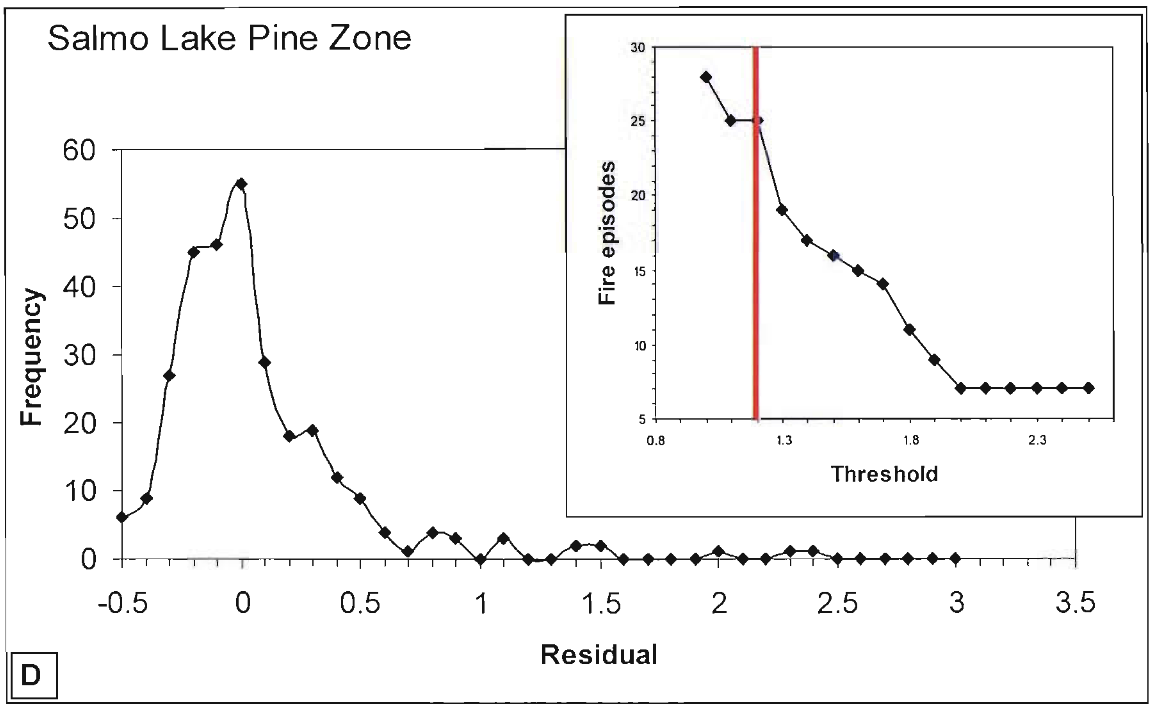


Figure 7.1 continued: Frequency distribution of the CHAR residual for the a) deciduous b) spruce c) stomata and d) pine zones of Salmo Lake. Inset in each shows the sensitivity of the number of CHAR peaks detected using a range of threshold values. The red line shows the threshold chosen to identify charcoal peaks most likely due to local fire episodes.

7.3 Zonation

In order to define the fire regime accurately under the different phases of vegetation history of each site, the records were split into vegetation zones as defined by the pollen diagrams (see Chapter 6). Each vegetation zone was then run through CHAPS separately to identify the background, threshold and peak frequency, etc. Splitting the records into vegetation zones was decided on as the most appropriate approach given that the study is focussed on whether fire regimes are affected by vegetation type. The exact position of the splits in the record were dictated by the pollen abundance of dominant tree species (spruce or pine) exceeding a 10% threshold, or the appearance of pine stomata within the record (see Chapter 6). Therefore, each zone provides the fire record of a specific vegetation type on the landscape, and the effects of increasing or decreasing different species abundances, and their possible effect on the fire regime will be interpreted separately in Chapter 8. As each pollen assemblage zone is considered separately, the major shifts in charcoal influx/ peak frequencies that occur during the transition phases are isolated into a single zone for interpretation. Due to the fact that in the early Holocene the vegetation transition occurs very close to a major climatic shift, it is necessary to analyse the data in such a way as to allow identification of climate-related shifts in the fire regime.

It should be noted here that the frequency of peaks is of primary interest to this study as this provides an estimate of fire regime activity or FRF. The influx values provide an indication of the amount of available biomass for burning, which is important in the interpretation of these records in terms of vegetation type and its effects on fire regime. Rapid alterations in the background influx values can have dramatic effects on the running mean. As these step-like shifts are generally associated with vegetation transitions, the zone division reduces the impact of the step-increase in influx elevating the running mean in the lower influx zones, and masking any patterns of peaks within them. The data are normalised prior to analysis with CHAPS to reduce the magnitude of shift changes as far as is reasonable practicable, but extreme peaks do still occur. Another way to deal with this would have been to reduce the window width for the running mean; however, this would have the effect of introducing more noise into the data.

Zones are shaded on all diagrams in a colour-coded fashion. Brown indicates the deciduous zone, green shading demarcates the spruce zone, blue shaded areas represent the stomata zone, and orange for the pine zone.

7.4 Alaskan Lakes

7.4.1 Little Harding

The analytical window for Little Harding Lake, core Z, covers the spruce transition at ~10 ka, and the charcoal influx data are represented in Figure 7.2 below. The overall influx values are presented, and in general the influx level is very low (mean ≤1 particle/cm2/yr); however, there are very few levels with zero, or near zero influx values in the original data.

The vegetation transition period (from Deciduous- to spruce-dominated forest) ~10100 to 9800 cal yrs BP is associated with an apparent abrupt increase in charcoal influx and peak frequency. This is probably a reflection of the age-depth model, as it stands at the time of writing. The author is awaiting the results of further dates (see Chapter 5), which are expected to constrain the inflection of the curve, which experiences two alterations in sediment accumulation rate in this period. It is unlikely that the pattern of charcoal influx as it is presented here is an accurate representation of the fire regime at the time. This site will be re-interpreted when a more robust chronology and age-depth model are secured.

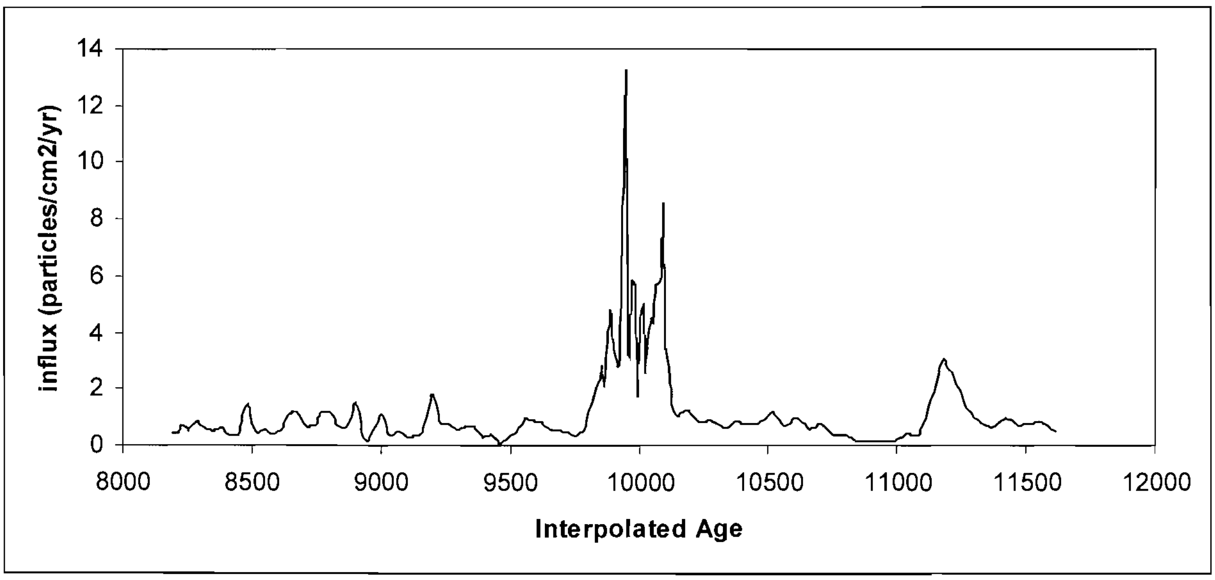
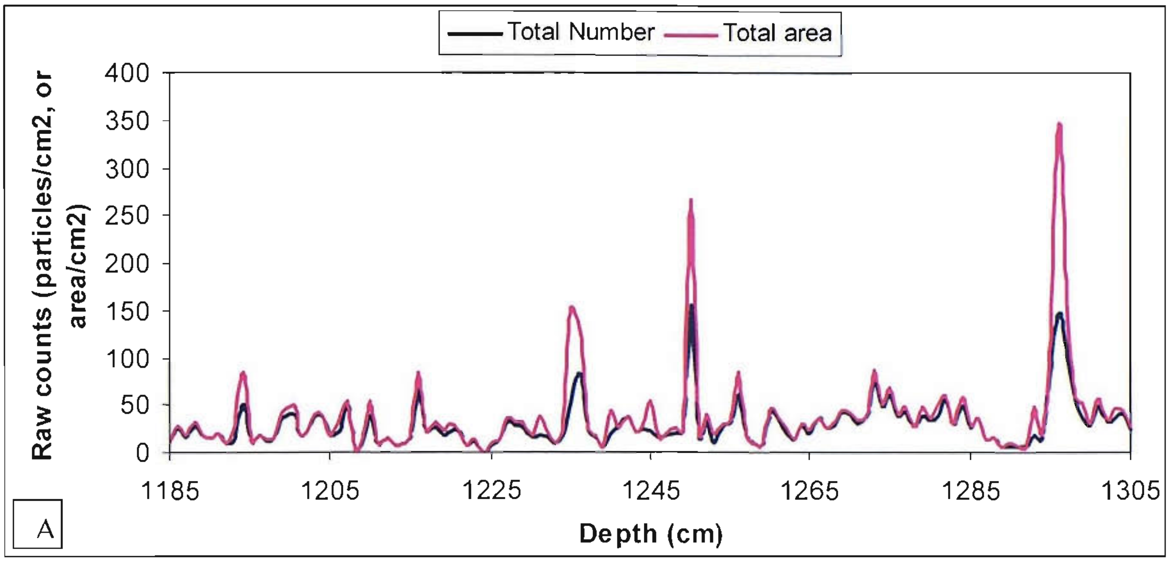


Figure 7.2: Little Harding Lake, core Z, showing charcoal influx for the analytical window, using a linearly-interpolated age model.. Ages shown here are interpolated to 10 year intervals by CHAPS.



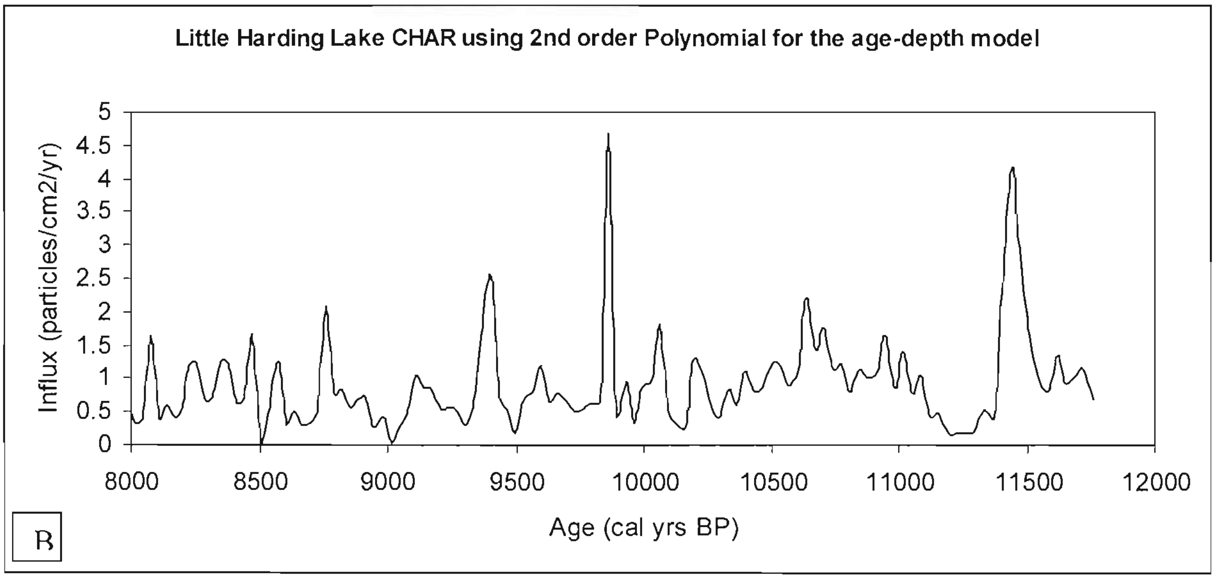


Figure 7.3: Little Harding Lake, core Z, showing charcoal influx data for the analytical window as a) raw counts comparing particle number and area and b) using a second order polynomial age model to interpolate influx to equal time slices (particle number).

The raw charcoal influx data are presented in Figure 7.3a above. The raw data display relatively little difference between particle number and particle area, except for the height of the biggest peaks and very slight differences within the spruce zone. The record is then interpreted with a second order polynomial used to inform the age-depth model (Figure 7.3b). Although this type of regression is not thought to be an accurate representation of sediment accumulation as it does not pass within the 2σ range of most of the calibrated radiocarbon determinations, it is presented to highlight fact that the original data have the potential to provide an interpretable charcoal record, and that the raw data retain their basic pattern after interpolation and normalisation within CHAPS.

For explanatory purposes, the CHAPS output for this site using the second order polynomial is presented (Figure 7.3), although it is not used to provide an interpretation of the fire regime at present. The record was split into two, dictated by the rise in spruce pollen (see Figure 6.3, 1245cm *ca.* 9900 cal yrs BP). It can be seen that when using this age-model there is relatively little difference in peak frequency between the two vegetation zones although There is an overall increase in mean fire episode frequency from the deciduous to spruce zones.. The estimates of fire regime statistics are presented in Table 7.1.

Vegetation	Mean Time Between Peaks	Fire Frequency (#/1000 yrs)	Fire Episodes in zone	Average Influx (particles/cm2/yr)
Deciduous	269.4	4.5	8	1.0
Spruce	201.4	6.0	13	0.78

Table 7.1: Little Harding Lake, showing estimates of fire regime statistics for all vegetation zones using a second order polynomial for the age-depth model. Threshold for both zones was set at 1.1x.

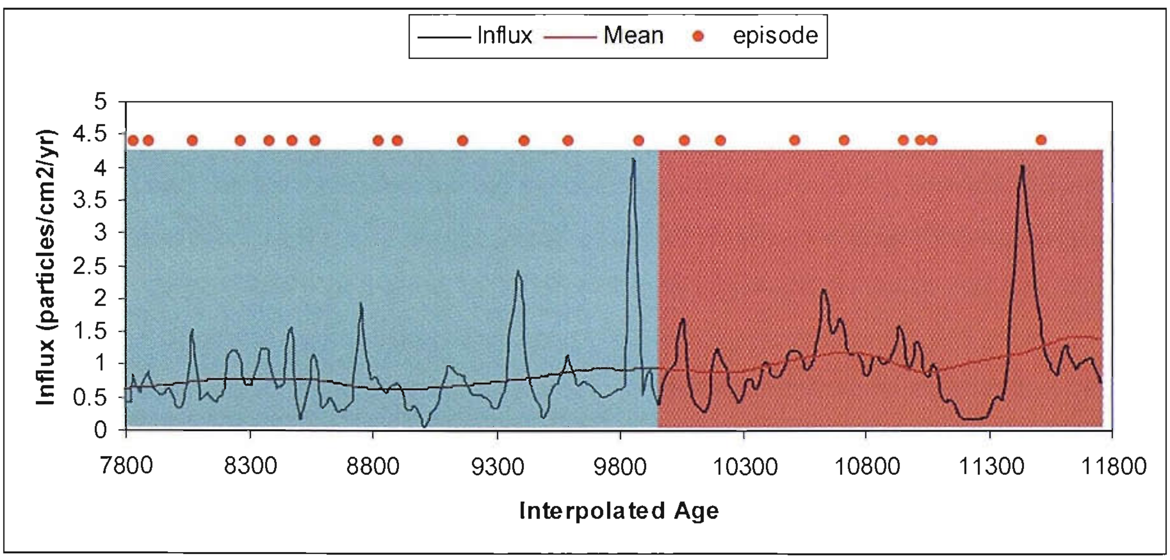


Figure 7.4: Little Harding Lake, core Z, showing charcoal influx when using a second order polynomial. The mean is demarcated by the red line, and the point at which a peak exceeded the assigned threshold (1.1x the mean) by the orange dots. The deciduous zone is shaded brown, and the spruce zone green.

In general Little Harding Lake is a difficult site to interpret. The overall charcoal influx values are relatively low creating uncertainty with the interpretations. The CHAPS model will determine peaks in any data-set. However, the low influx values are interpretable in terms of fire occurrence if the difference between the background and peak components is of significant magnitude. Many studies have interpreted low influx values (e.g. Whitlock & Millspaugh, 1996; Millspaugh *et al.*, 2000; Brunelle-Daines, 2002; Whitlock *et al.*, 2004; Gavin *et al.*, In Press) and demonstrated statistical robustness.

It may be the case that the low influx values are reflective of the relatively large size of the lake (in comparison with the Yukon sites), and therefore the charcoal influx is diluted. Charcoal depositional processes were discussed in Chapter 2, and the time-lag between fire episode and final integration within the sediment in the deepest/central portion of the lake is a probable causal factor in reduced peak magnitude and general influx values (Bradbury, 1996; Whitlock & Millspaugh, 1996; Gardner & Whitlock, 2001). The background/noise component of such records is proportionally larger in relation to the peak component; however macro-charcoal is still considered to be reflective of local fire regimes. A reliable chronology will greatly reduce the uncertainty associated with the interpretation of the peak distribution of this record.

The influx values decrease across the transition to spruce-dominated forest, and this is possibly due to the fact that the deciduous vegetation was well established, and probably comprised birch trees and associated shrubs and herbs rather than shrub species of birch (Hopkins *et al.*, 1981). Tree birch would provide significantly higher amounts of fine fuels to the catchment in comparison to a more shrub-dominated landscape, and the switch to increasing spruce proportions would not provide a dramatic increase in particles to the catchment (Scott *et al.*, 2000; Enache & Cumming, 2006). Therefore it is possible that the low initial proportions of spruce on the landscape, and the composition of already established vegetation into which it established, led to an apparently lowered amount of biomass burning within the catchment, as reflected in the charcoal record.

7.4.2 Birch Lake

The analytical window used for Birch Lake core 97b covers a ~3000 year period across the transition from deciduous to spruce dominated vegetation (see Chapter 6). The charcoal influx record for the whole section of core analysed is presented in Figure 7.5 below. It can be observed that overall the influx values are very low, with a general decreasing trend up-core. The record displays two phases of influx, with higher values and peaks of greater magnitude in the period from ~9100 – 10300 interpolated yrs BP, and much lower influx values, and smaller magnitude peaks from ~9100 – 7600 interpolated yrs BP.

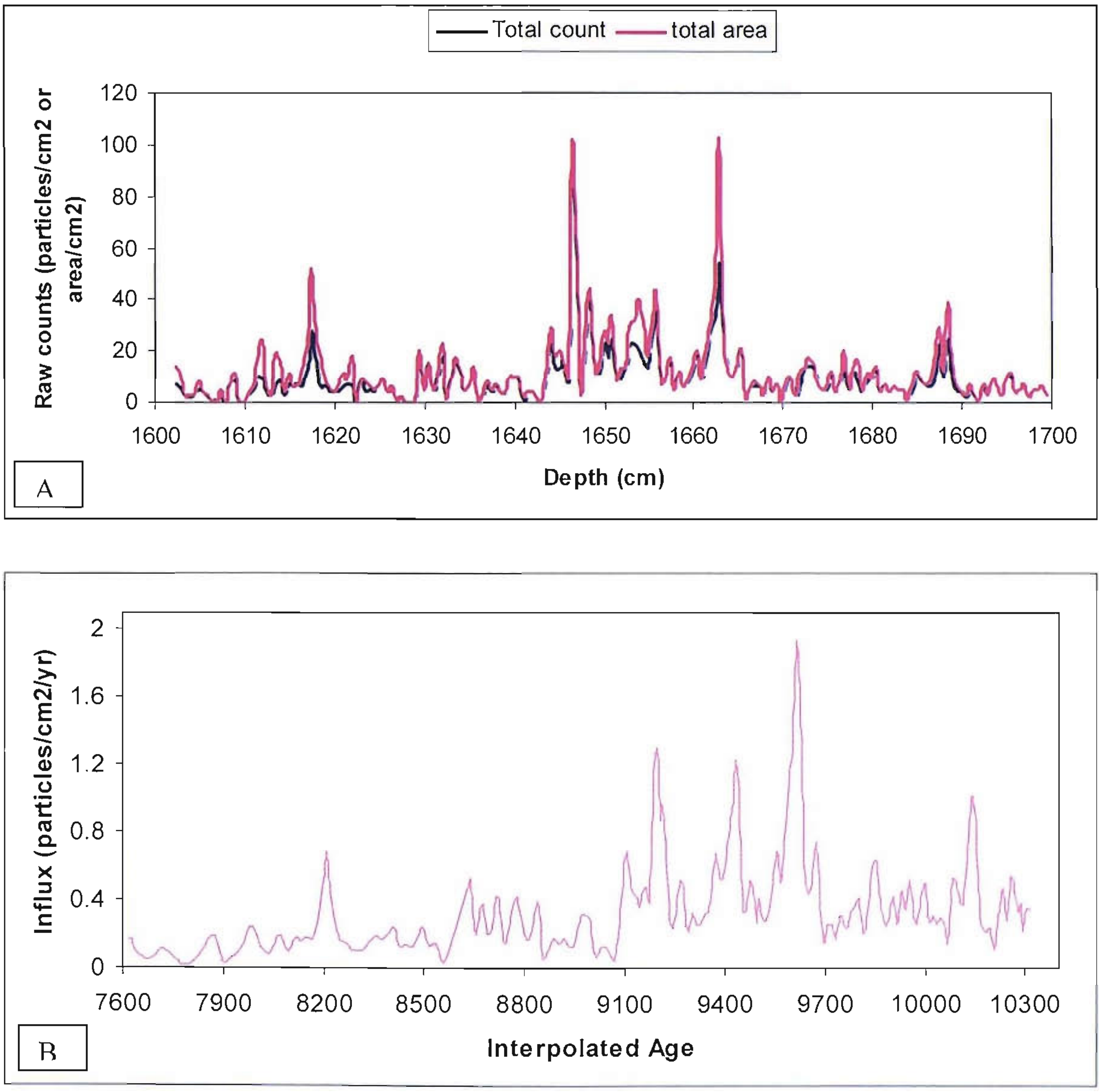


Figure 7.5: Birch Lake core 97b showing a) raw count data for particle number and charcoal area and b) charcoal influx diagram using interpolated age

The record was split into two sections determined by the pollen diagram (Figure 6.6), and the charcoal influx record for the deciduous zone is presented in Figure 7.6 below. The record displays an overall increase in mean influx from the start of the record at ~10300 interpolated yrs BP to the end of the zone *ca.* 9450 interpolated yrs BP. With low overall influx values (Table 7.2) it is difficult to interpret the record, as at Little Harding Lake above. The interpolated model data are a very close match to the raw counts in terms of trends in influx and peak magnitude (see Figure 7.5). It is probable that the majority of the peaks identified are representing a presence/absence record due to the low influx values, with some of the identified peaks actually constituting part of the background noise and in fact the 3 larger peaks are a clearer reflection of burning in the catchment, creating a mean peak frequency of ~300 years. At Birch Lake the vegetation surrounding the lake is well established prior to the start of the charcoal record, and therefore it is probable that the low influx values are a product of the large size of the lake diluting the charcoal concentration.

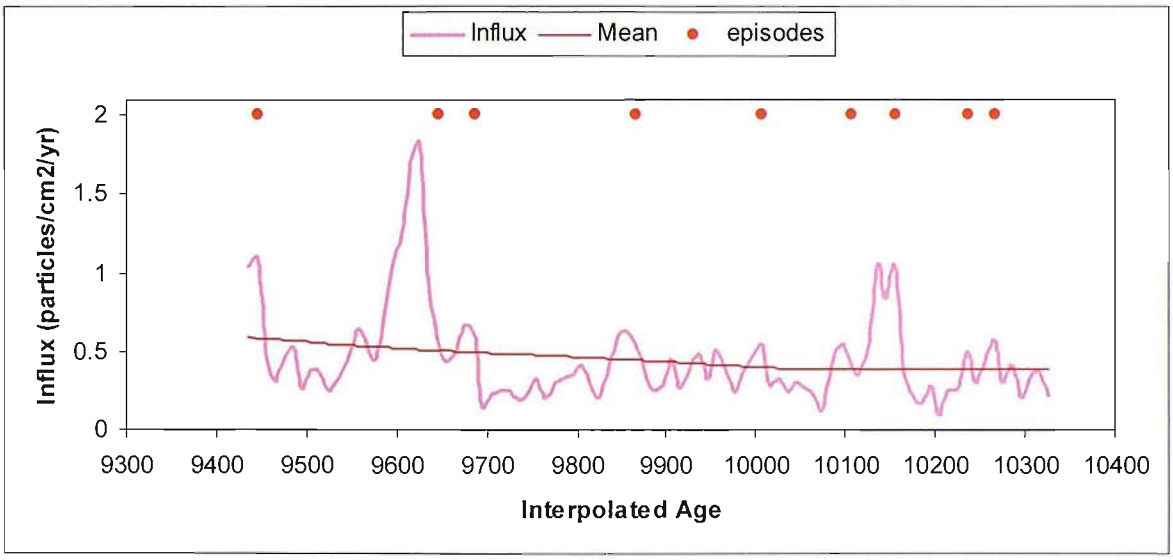


Figure 7.6: Birch Lake core 97b charcoal influx diagram for the deciduous zone. The red line represents the mean influx, and the orange dots represent the point at which a peak exceeded the 1.2x threshold applied.

The fire regime statistics calculated for this zone are very much influenced by the low influx values (Table 7.2). The mean time between peaks (TBP) is extremely low when compared to the Little Harding Lake record (see above), and illustrates further that probably only the three largest peaks are in fact a true reflection of a fire episode within the catchment with the smaller peaks are influenced by noise created due to the dilution effects discussed above.

Vegetation	Mean Time Between Peaks	Fire Frequency (#/1000 yrs)	Fire Episodes in zone	Average Influx (particles/cm^2/yr)
Deciduous	112.2	9.1	9	0.46
Spruce	146.7	9.9	18	0.24

Table 7.2: Fire regime statistics for Birch Lake, Alaska divided by vegetation zone.

The spruce zone spans the period *ca.* 9450 interpolated yrs BP – the end of the analytical window at ~7600 interpolated yrs BP. The charcoal influx record is presented in Figure 7.7 below, displaying an overall decreasing trend in charcoal influx through the zone. Again the charcoal influx values are relatively low, which will influence any interpretation.

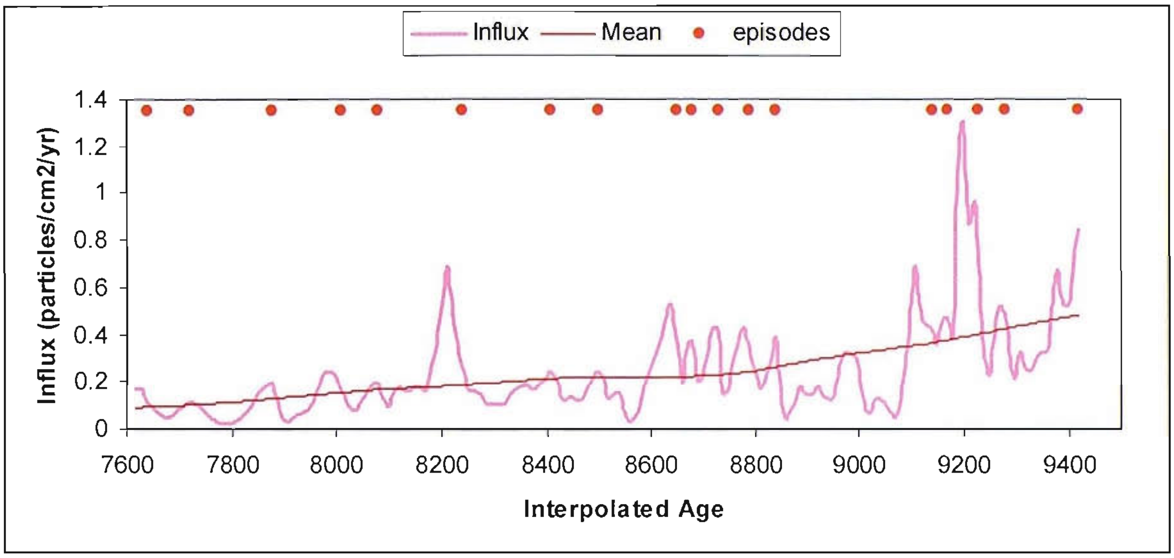


Figure 7.7: Birch Lake core 97b charcoal influx diagram for the spruce zone. The red line represents the mean influx, and the orange dots represent the point at which a peak exceeds the 1.1x threshold applied. ***Note:*** *y-axis scale change when compared with Figure 7.5 and 7.6.*

The zone begins with a series of relatively high magnitude peaks and decreasing mean influx. There is a quiescent period in the fire episode record between ~9100 and ~8850 interpolated yrs BP, which is not entirely a function of this down-trending in mean influx. The actual influx values at this point also become very low. Periods are identified as quiescent if there are no identified peaks for a time that exceeds the mean TBP for the zone. There is a general increase in magnetic susceptibility in this portion of the record (see Chapter 6), which may contribute to the lowered influx values via dilution through increased autocthonous input into the lake.

From ~8800 interpolated yrs BP, the charcoal influx continues to decrease, with an increase in peaks identified. The fire regime statistics (Table 7.2) provide figures for mean TBP and fire frequency that suggest an increase from the deciduous zone. However, the deciduous zone values, as discussed above may not be an accurate reflection of the fire regime.

Due to the low influx values and the relatively small magnitude peaks, it would be possible to fit a running mean to the entire record, without loss of peak frequency data. However, for consistency with the other sites investigated, this record was treated in the same manner, by splitting the record into two vegetative zones.

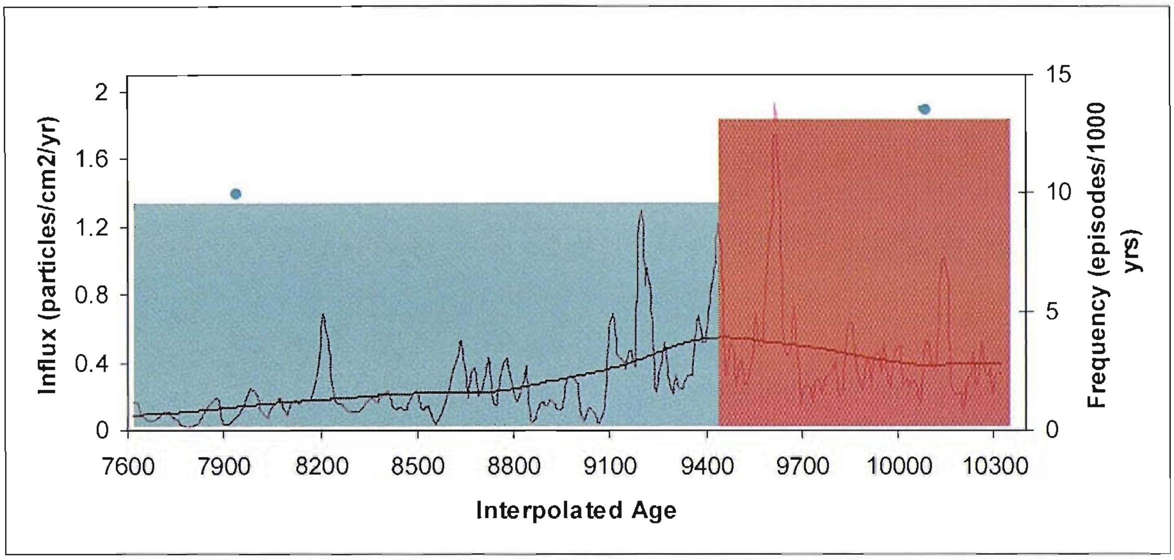


Figure 7.8: Charcoal influx diagram for the analytical window at Birch Lake, showing the mean influx (red line) and the number of fire episodes per1000 years (blue dots) for each vegetation zone.

The charcoal record for Birch Lake in relation to potential fire return frequency is presented in Figure 7.8 above. Using the mean episode frequency data for the two vegetation zones there is a decreasing trend in fire frequency from the deciduous zone to the spruce zone; however, until a full chronology is established on this core to provide an accurate representation of fire return frequency and any further interpretation is limited.

7.4.3 Summary

The low influx values create uncertainty with regard to interpretation of peaks. It is probable that the relatively large basin size of the lake produces a diluted charcoal record. The increase in spruce pollen occurs relatively rapidly, with an increase from ~5% to ~30% within *ca.* 10cm core length (~250 years). However, after the initial increase, the abundance decreases to and stabilises at ~20%, with corresponding *Betula* abundances at ~80%. This seems to be characteristic of the Alaskan sites, while the Yukon lakes display rapid and permanent switches to spruce dominance, with a coincident decrease in *Betula* abundances. Therefore it is probable that the charcoal record for the spruce zone in Alaskan sites is reflective of a mixed spruce/birch forest, and the flammability of and the available fuel load within the catchment will differ from the Yukon sites accordingly.

7.5 Yukon Lakes

7.5.1 Salmo Lake

The charcoal record for the complete core of Salmo Lake is presented in Figure 7.9 below, showing a full Holocene record. There are very low influx values from the start of the record (~12000 cal yrs BP) to *ca.* 9500 cal yrs BP. Particle number data were analysed as area data had increased peak magnitude but identical influx values. There is an abrupt increase in influx and the acuteness of the peaks at *ca.* 9500 cal yrs BP, which is sustained to ~6000 cal yrs BP. Subsequently, there is a slight decreasing trend in influx, although the acuteness is maintained.

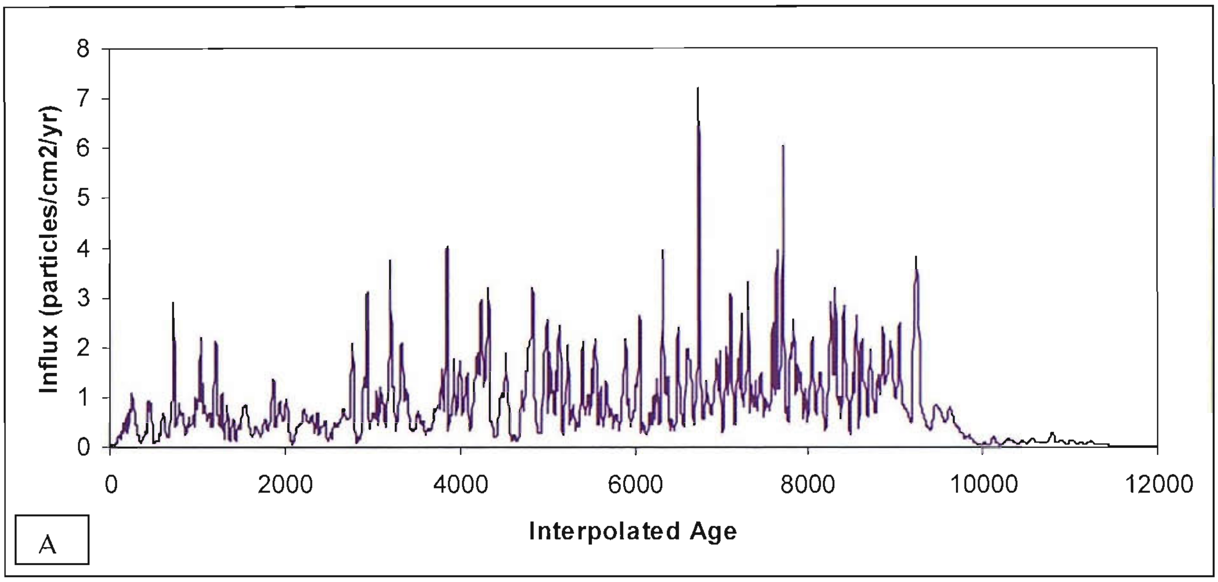
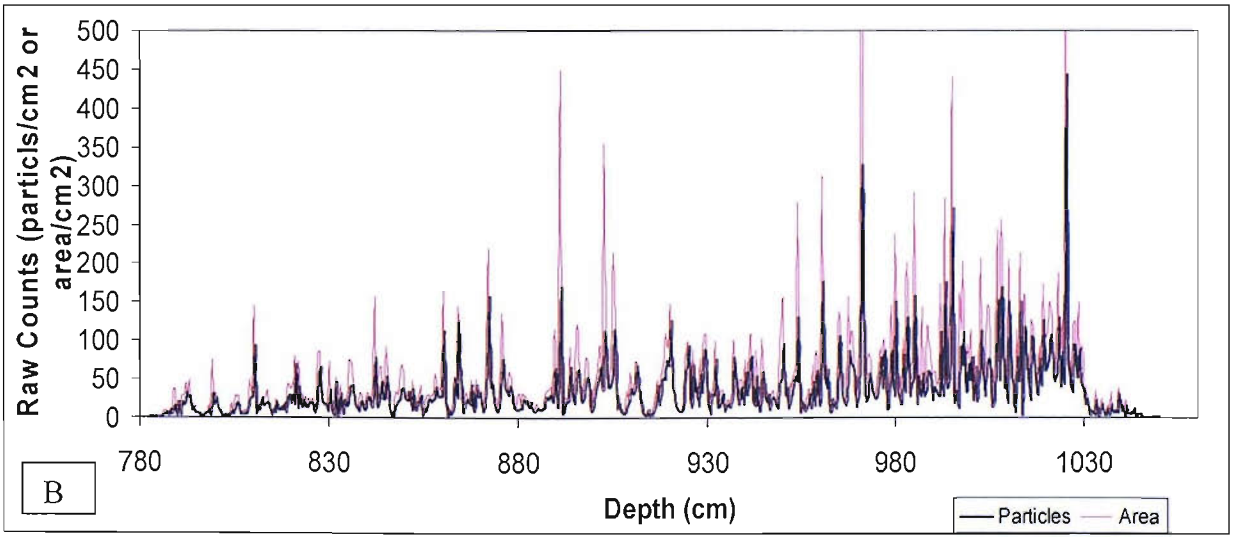


Figure 7.9: Salmo Lake, core A, showing above a) the charcoal influx data against age using interpolation of dates into 10 year increments, and below b)raw counts of charcoal particle number and area. Y-axis has been truncated to 500 for clarity.



The influx decreases further at ~2500 cal yrs BP, and the magnitude of the peaks is also reduced temporarily from this point. The lowered influx is maintained to the surface of the core; however, the magnitude of the peaks increases between ~500 and 1500 cal yrs BP. The near-surface samples illustrate a rapid decrease in influx, indicating that the charcoal at these levels is influenced by a dilution effect due to the less compacted nature of the sediments and the extended deposition time of charcoal post-fire (refer to Chapter 2).

The record has been split into four zones consistent with the pollen zones allocated in Figure 6.9. The charcoal influx and fire episodes identified by CHAPS for the deciduous zone are presented in Figure 7.10. It can be seen that the influx values in this zone are very low, although there is a general increasing trend from the start of the record to ~10800 cal yrs BP. The record consists of low influx values, from which CHAPS interprets significant variation and creates numerous identifiable peaks above the 1.1x threshold applied. However, fractional particles interpreted by CHAPS signify very small if not zero counts within the original data set (Figure 7.9b). The interpretation of such low influx values must be carefully considered. It is probable that the low influxes are a function of low raw counts divided over long deposition times. A background component must still be present, and therefore, some of the positive values must be part of the background. Due to this uncertainty, a more conservative interpretation is considered.

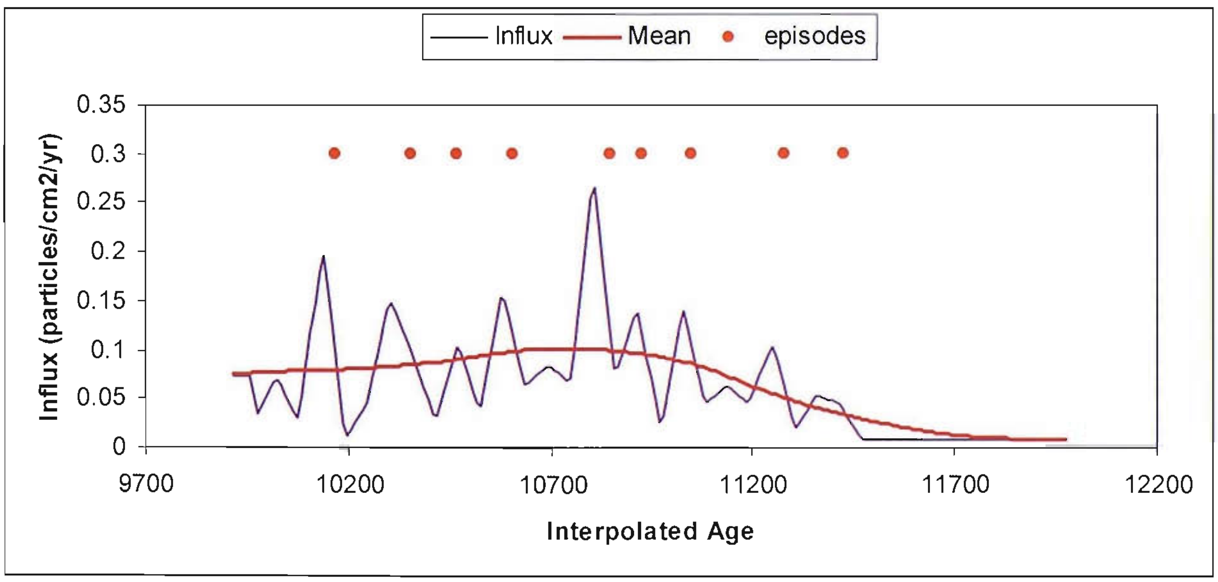


Figure 7.10: Salmo Lake, core A, showing charcoal influx for the deciduous zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

The mean influx is very low at 0.06 particles/cm^2/yr (Table 7.3), and the average TBP is calculated at ~288 years. However, due to the relatively late timing of deglaciation of this area, and thus the formation of the lake, the deciduous zone record is relatively short (~1600 years), and therefore these statistics may not represent the fire pattern of an established deciduous zone.

Vegetation	**Threshold**	**Mean Time Between Peaks**	**Fire Frequency (#/1000 yrs)**	**Fire Episodes in zone**	**Average Influx (particles/cm^2/yr)**
Deciduous	1.2	288.7	4.5	9	0.06
Spruce	1.2	174.9	8.2	29	1.25
Stomata	1.1	169.9	8.8	29	0.98
Pine	1.2	161.7	8.5	25	0.55

Table 7.3: Salmo Lake, deciduous zone fire regime statistics for each vegetation zone.

The newly deglaciated landscape surrounding this lake would have had very little if any soil cover and thus a relatively low potential to support vegetation for a period of time after the ice retreated (ecesis time (Baillie, 1995)), which is probably reflected in the near zero values for charcoal influx at the very start of the zone. The persistence of low influx values for the rest of the zone are possibly reflective of a) the newly establishing and *ergo* sparse vegetation cover, b) the vegetation mainly comprising deciduous shrub and herb species, and therefore relatively more difficult to ignite than coniferous vegetation, c) the low amount of biomass available to produce charcoal, and d) the low productive capacity of a site in close proximity to the ice edge. The area surrounding Salmo Lake (see Figure 4.10) would potentially not burn readily given the topography, geology and presence of many other water bodies nearby. (Chapter 4). The deciduous zone record will be interpreted qualitatively due to the presence of zero values in the raw counts, some of the charcoal peaks must constitute a proportion of background noise.

The spruce zone (~9900 to 6400 cal yrs BP) is represented in Figure 7.11 below. The pollen record for Salmo Lake (Figure 6.9) shows the increase in spruce abundance is sudden and abrupt at ~9800 cal yrs BP. The charcoal record displays low influx levels continuing from the deciduous zone into the start of the spruce zone, with a general trend to an increase in the mean influx values to a maximum at ~9000 cal yrs BP, where the values stabilise for the remainder of the zone.

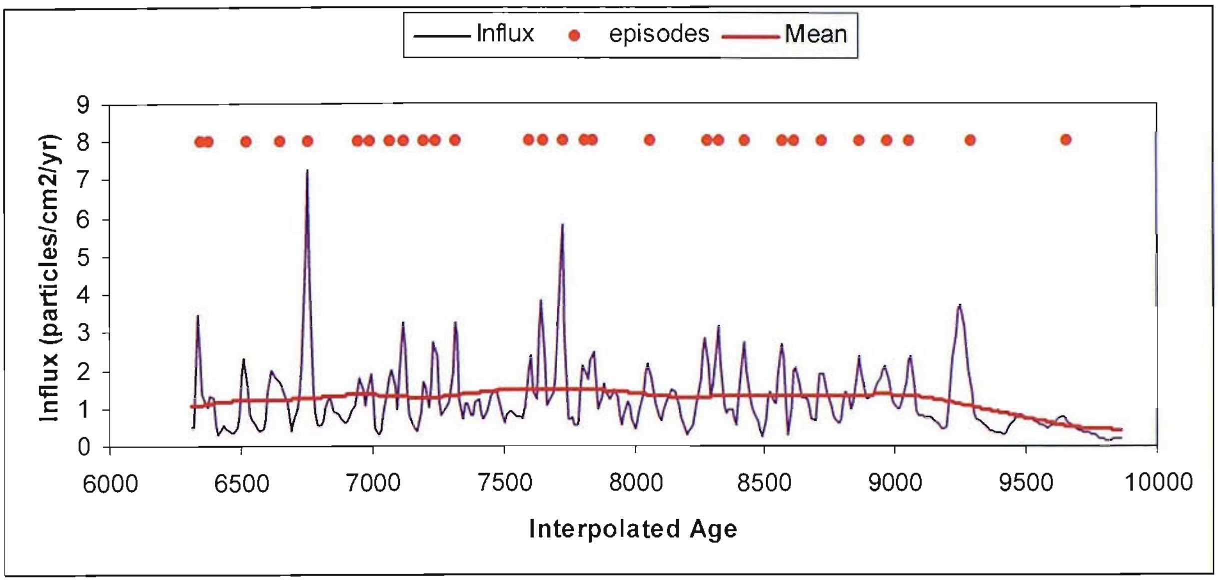


Figure 7.11: Salmo Lake, core A, showing charcoal influx for the spruce zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

The mean influx value is 1.25 particles/cm^2/yr (Table 7.3), which is a twenty-fold increase from that of the deciduous zone. A large number of fire episodes (8.2/1000 yrs) reflects the high variability in influx values. The mean TBP is ~174 years, greatly reduced from that of the deciduous zone.

The "pure" spruce zone ends with the appearance of pine stomata in the pollen record (Figure 6.9) at ~6400 cal yrs BP, indicating the beginning of the pine invasion, and the transition from white to black spruce.

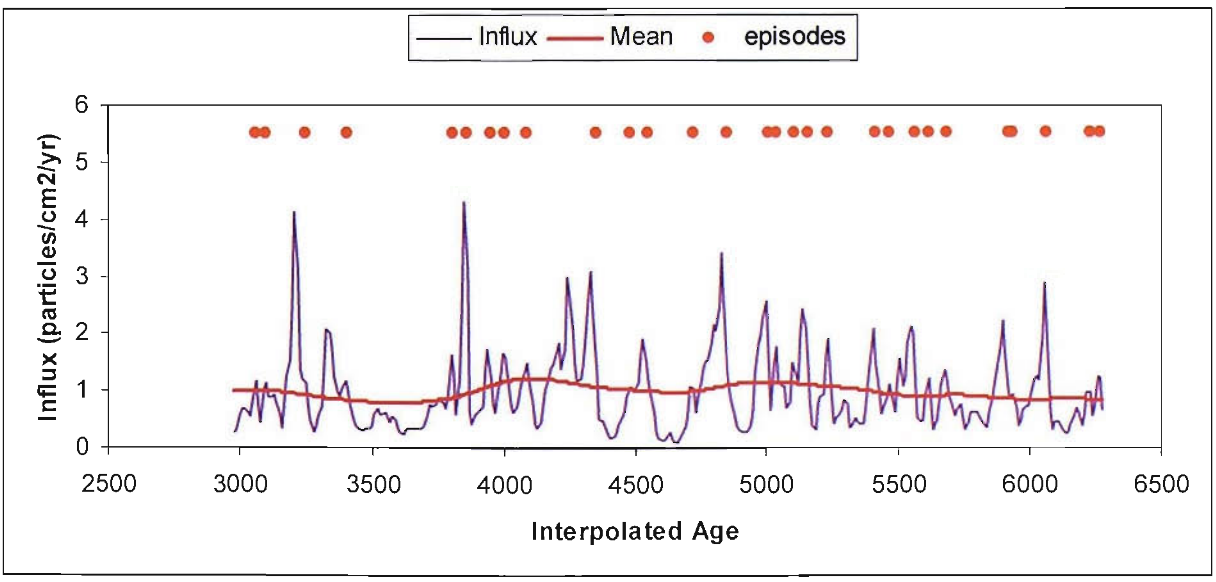


Figure 7.12: Salmo Lake, core A, showing charcoal influx for the stomata zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

The transition to the stomata zone at ~6400 cal yrs BP sees a decrease in influx values, represented in Figure 7.12. The mean influx remains relatively stable within this zone at ~1 particle/cm^2/yr (see also Table 7.3). The fire episodes appear to be relatively frequent, although more gaps are present than in the spruce zone. The most prominent charcoal low appears between ~3800 and 3500 cal yrs BP, where the overall influx appears diminished.

The charcoal data for the pine zone are presented in Figure 7.13, and reflect lower influx values, averaging ~0.55 particles/cm^2/yr (see Table 7.8). The influx is relatively variable within the zone, with a slight decreasing trend towards the surface. The highest frequency and magnitude of peaks is located between ~ 1400 and 700 cal yrs BP, coincident with the WRA.

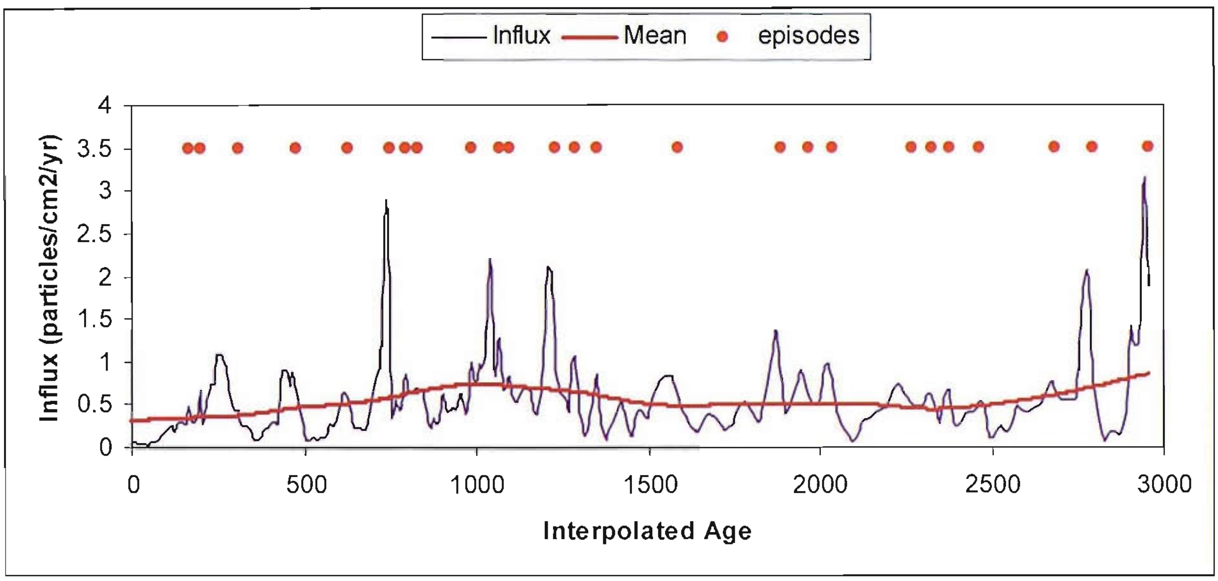


Figure 7.13: Salmo Lake, core A, showing charcoal influx for the pine zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

The influx values decrease rapidly from ~200 cal yrs BP to the surface, and are probably a reflection of the less compacted nature of the surface sediments. These top-most levels are not included in any further analyses as they are not considered representative of the pine record as a whole.

The mean TBP in this zone is 161 years (Table 7.3), which is not significantly different from the TBP of the stomata zone. The fire episode frequency for each vegetation zone is plotted against the influx data in Figure 7.14. The episode frequency becomes relatively stable once coniferous forest has been established at this site (after ~ 9800 cal yrs BP), being much lower during the deciduous zone.

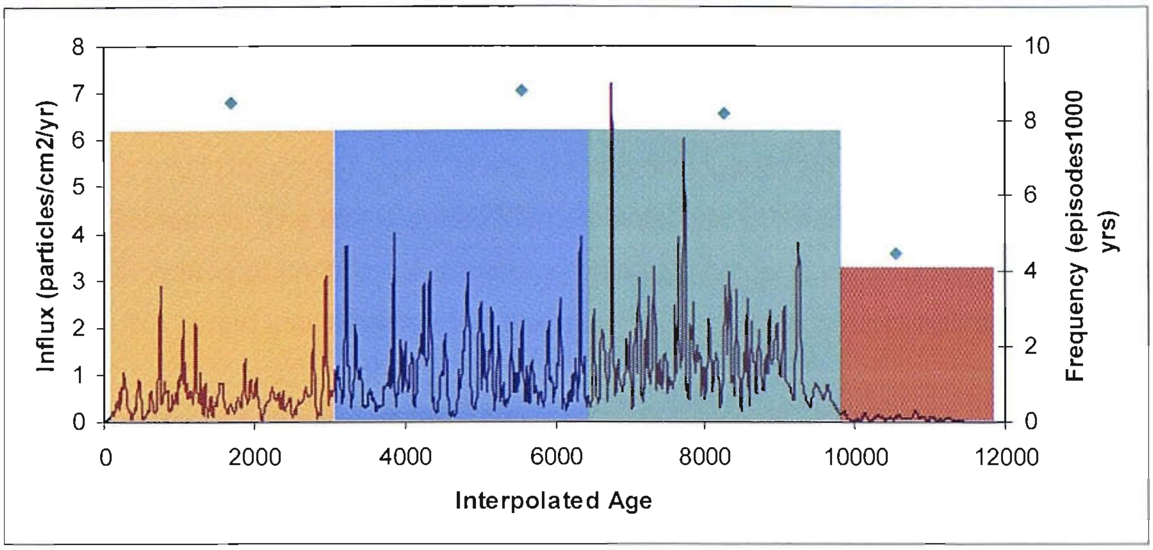


Figure 7.14: Salmo Lake, whole core charcoal influx diagram, with average peak frequency per zone plotted (blue dots). Vegetation zones are shaded brown (deciduous), green (spruce), blue (stomata) and orange (pine).

7.5.2 Dragonfly Lake

The charcoal influx data for Dragonfly Lake are presented in Figures 7.15a,b and c, showing a full Holocene record. There are two obvious abrupt increases in charcoal influx in this record. The first is at ~9500 interpolated yrs BP, and associated with the increase in spruce abundance. A secure chronology and constant sediment accumulation rate over this period indicates that this abrupt shift in charcoal influx is reflective of the actual charcoal deposition rate.

The second abrupt shift in charcoal influx values is at ~1150 interpolated yrs BP, and is associated with the WRA and an alteration in sedimentation accumulation rate created by the disparity between the age of the WRA and the adjacent radiocarbon determinations. Due to the coincidence of the increase in influx values and the alteration in sediment accumulation rate when using a linearly interpolated age-depth model (Figure 7.15b), it was necessary to test alternative age-depth models to asses the impact on the charcoal record.

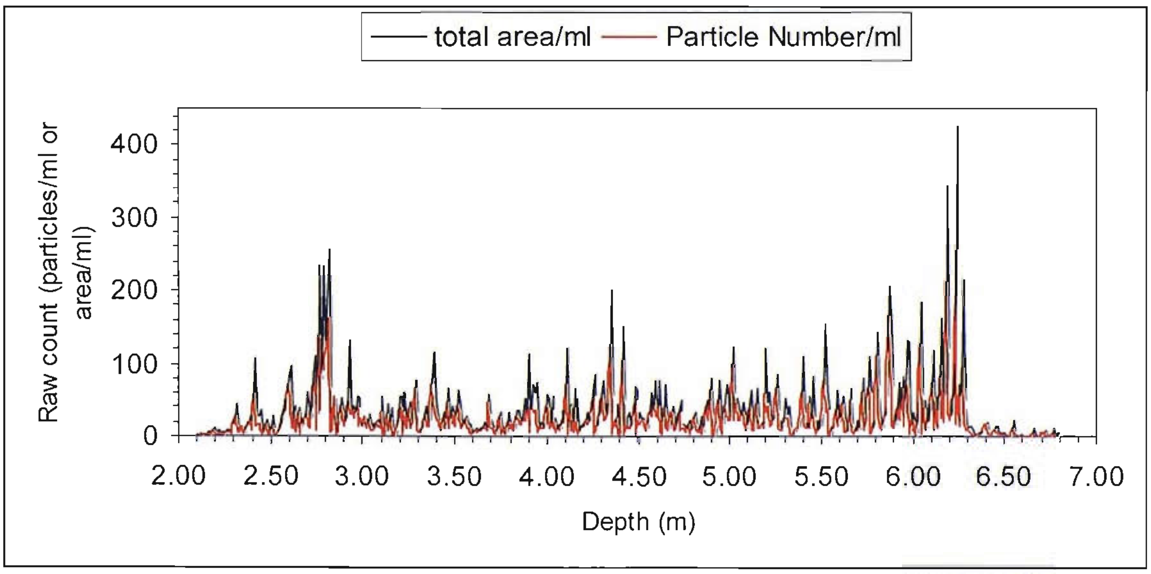


Figure 7.15a: Dragonfly Lake raw charcoal counts of particle number and total area against depth.

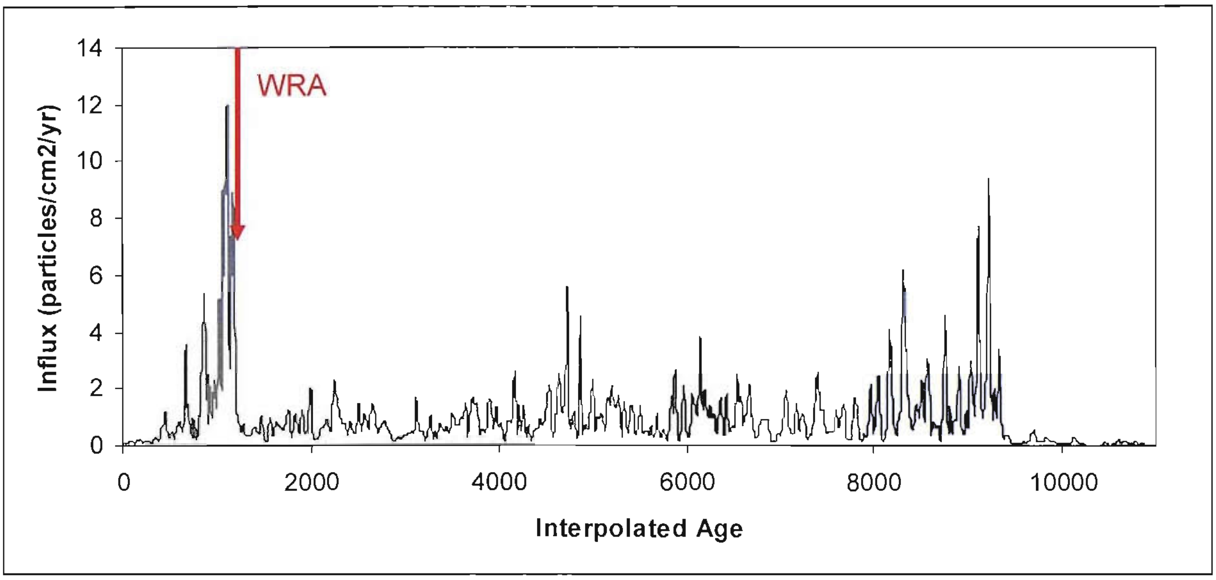


Figure7.15b: Dragonfly Lake charcoal influx diagram using a linearly interpolated chronology. Abrupt sedimentation rate change is associated with the WRA tephra coincident with the rapid increase in influx values at ~1150 interpolated yrs BP.

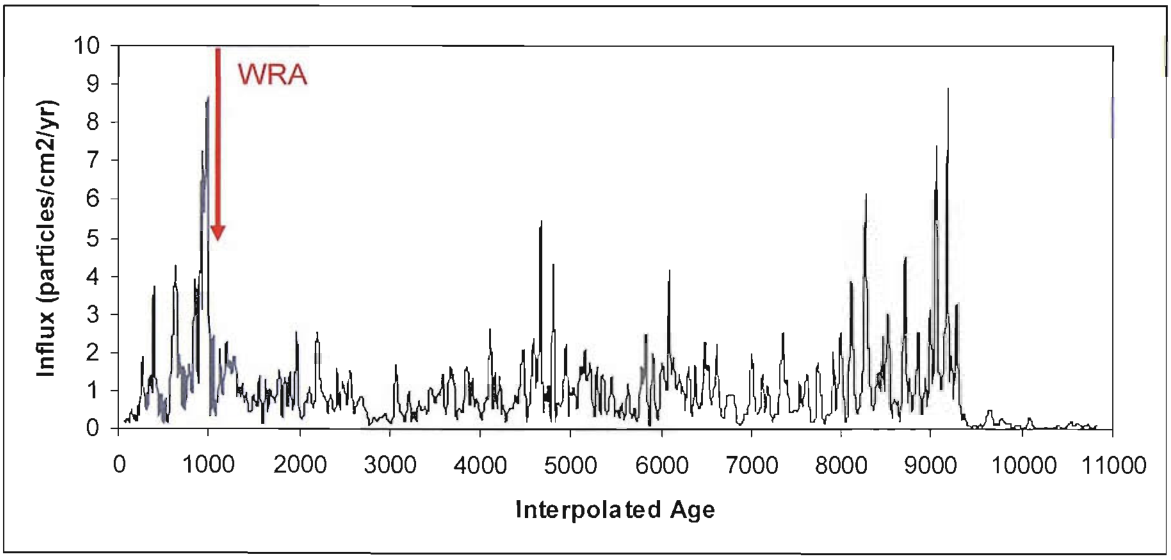


Figure 7.15c: Dragonfly Lake charcoal influx diagram using a spline interpolation chronology as an alternate chronology to investigate the effects of the abrupt change in sedimentation rate around the WRA tephra on the charcoal record.

It can be seen that using a spline interpolation to smooth the effects of the inflection in the age-depth curve a very similar pattern of charcoal influx is produced (Figure 7.14c). The increase at ~9500 interpolated yrs BP is nearly identical to the linearly interpolated model output. The increase at ~1150 interpolated yrs BP is also very similar when the alteration in the scale of the y-axis is taken into consideration. There is a slight elevation of influx values immediately prior to the main increase in influx, but the "event" still occurs, indicating that it is a reflection of the actual charcoal influx, and not a function of the chronology. The record will be interpreted using the linearly interpolated age-depth model, with results from the spline interpolation presented for the last ~5000 years for comparison.

In general the pattern of charcoal influx in Dragonfly Lake is low from the start of the record (~10900 cal yrs BP) until ~9500 cal yrs BP where there is a sudden and rapid increase in the magnitude of the peaks and the mean influx values. The influx values display a gradual decreasing trend to ~1500 cal yrs BP, where there is a second abrupt increase in both influx values and peak magnitude before the record becomes influenced by surface sediment properties, and influx values tail off to near zero levels.

The record was split into four zones determined by the pollen diagram (Figure 6.12). The charcoal record for the deciduous zone is presented in Figure 7.16. This part of the record displays very low influx values (<1 particle/cm^2/yr), with the average influx at near zero values (0.11 particles/cm^2/yr) (Table 7.4). The running mean displays a general increasing trend through the zone to ~9600 cal yrs BP.

Vegetation (age model)	**Threshold**	**Mean Time Between Peaks**	**Fire Frequency (#/1000 yrs)**	**Fire Episodes in zone**	**Average Influx (particles/cm 2/yr)**
Deciduous	1.4	207.8	5.6	7	0.11
Spruce	1.3	146.6	8.0	35	1.17
Stomata (linear)	1.3	191.3	7.0	27	0.92
Stomata (spline	1.3	189.2	7.1	27	0.91
Pine (linear)	1.4	292.6	4.3	5	1.5
Pine (truncated linear)	1.4	198.0	5.8	5	2.2
Pine (truncated spline)	1.4	174.8	6.9	7	1.71

Table 7.4: Dragonfly Lake fire regime statistics for each vegetation zone

peaks are identified above the 1.4x threshold, which produced a mean TBP of ~200 years (Table 7.4).With such low influx values it is probable that the record may not reflect accurately catchment burning at this time. The raw counts contain zero values in this zone, and as it is unlikely that the background component is completely absent, some of the peaks identified must be part of the background.

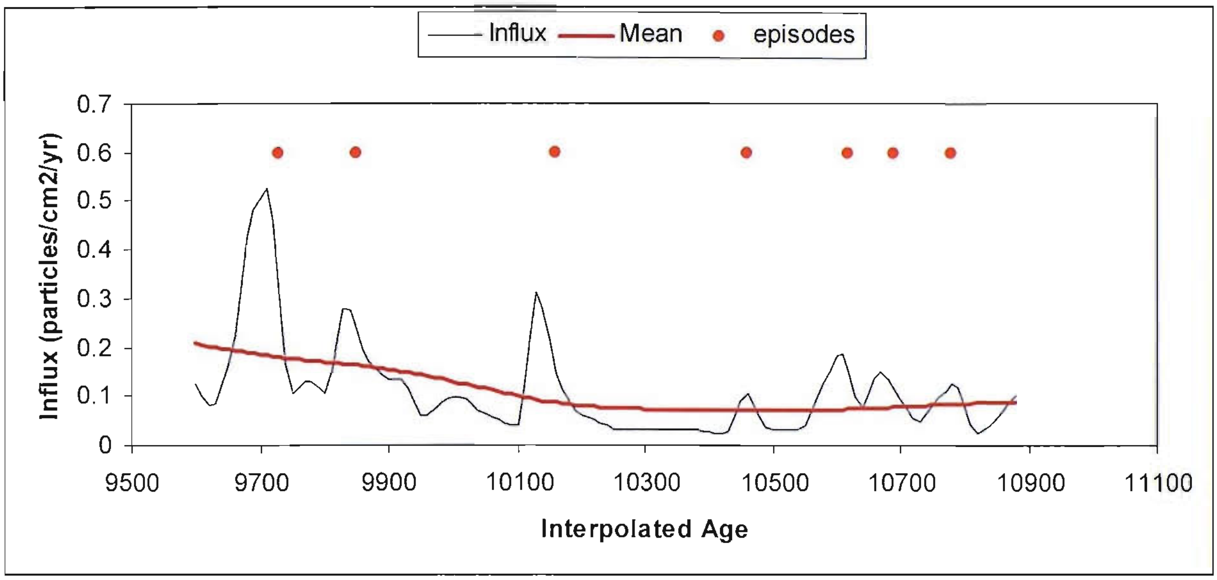


Figure 7.16: Dragonfly Lake, showing charcoal influx and fire episode for the deciduous zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

The deciduous zone is relatively short at Dragonfly Lake, as was the case for Salmo Lake. Dragonfly Lake is the northernmost of the Yukon sites, and would have been locally deglaciated earliest. However, the zone spans only ~1300 years.

The spruce zone ranges from ~9600 - ~5000 cal yrs BP and is represented in Figure 7.17. The influx values increase significantly at ~9500 cal yrs BP, with a coincident increase in the magnitude of the peaks identified. The peak frequency is relatively high at 8 episodes per 1000 years (Table 7.4) and remains consistently high through the whole zone. The influx values, after the initial increase through the transition from the deciduous zone, stabilise at 1.17 particles/cm^2/yr; however the magnitude of the peaks decreases with time through the zone.

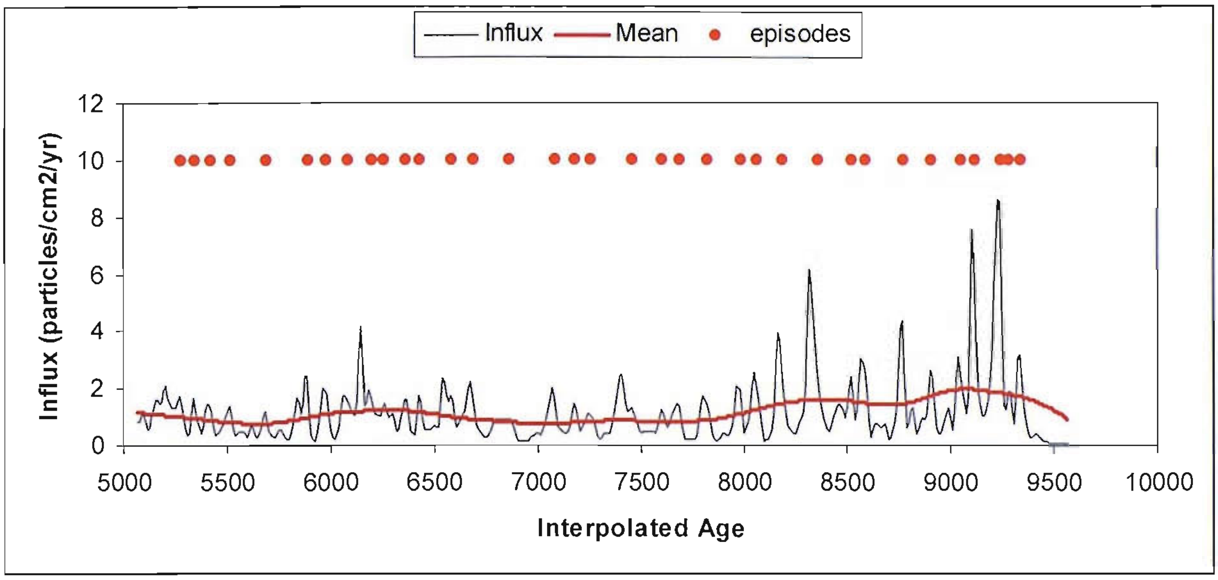


Figure 7.17: Dragonfly Lake, showing charcoal influx and fire episodes for the spruce zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

The mean TBP is ~146 years (Table 7.4), and the charcoal record displays a relatively stable fire episode frequency throughout the zone. It is probable that the higher influx values and the peak magnitude are linked to both the increased flammability of the catchment and the increased biomass available to burn once spruce trees are present.

The spruce zone is followed by the stomata zone, the charcoal influx and fire episode data of which are represented in Figure 7.18a. The influx data using the spline interpolated age-depth model is presented for comparison in Figure 7.18b. It is apparent that both fire episode records are identical, although the use of a spline interpolation has led to a slight decrease in influx level.

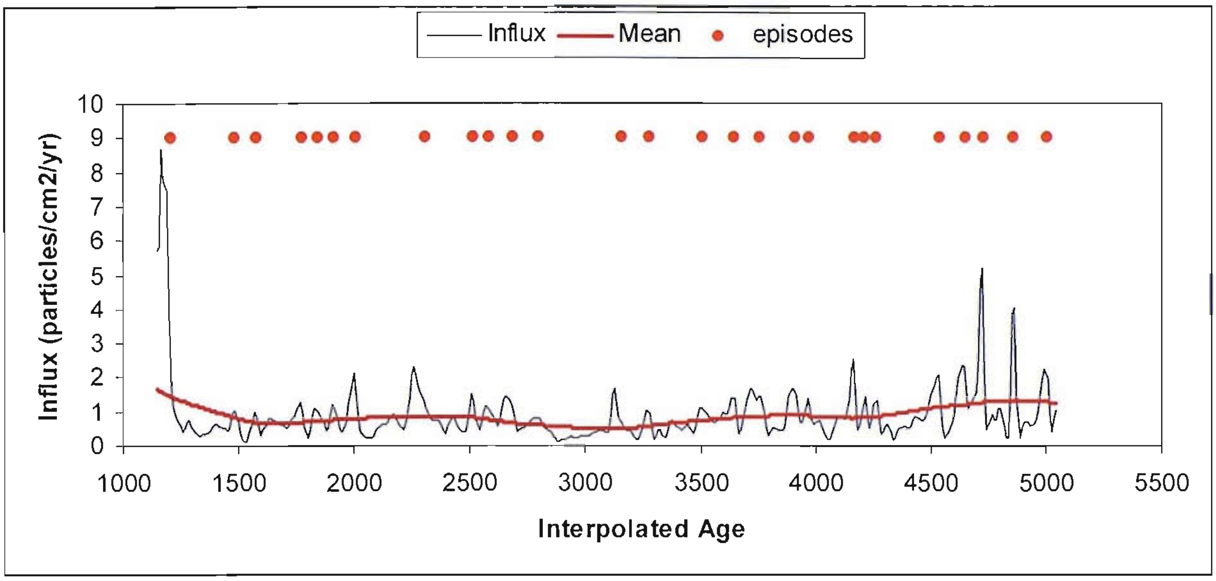


Figure 7.18a: Dragonfly Lake charcoal influx and fire episodes for the stomata zone using linear interpolation for the age-depth model. Red line marks the mean, and the point at which a peak exceeded the threshold by the orange dots.

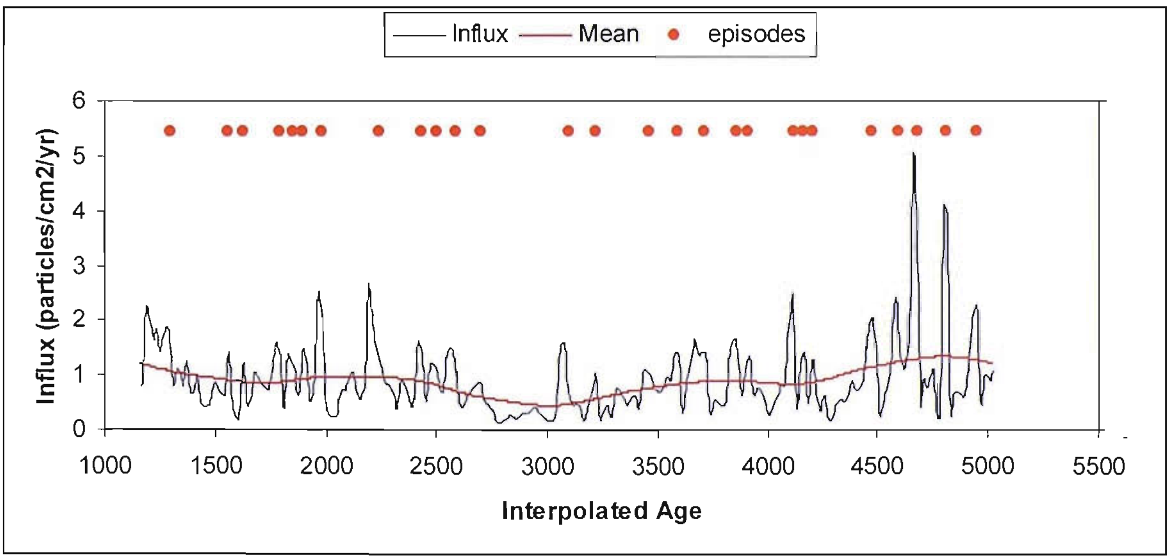


Figure 7.18b: Dragonfly Lake charcoal influx and fire episodes for the stomata zone using spline interpolation for the age-depth model. The red line marks the mean, and orange dots signify a peak exceeding the threshold.

The charcoal influx values are slightly lower than those of the spruce zone, averaging 0.92 particles/cm^2/yr (Table 7.4). The peak magnitude increases slightly at the transition between the spruce and stomata zones; however it is not sustained through the stomata zone, with the exception of one large magnitude peak at the transition with the pine zone.

The episode frequency is slightly lower than that of the spruce zone at 7/1000 years, the pattern of peaks above the threshold is relatively sporadic, with numerous quiescent periods and clusters of peaks. Quiescent periods are associated with much lower than average influx values. The large peak at the end of the zone demarcates the transition to the pine zone, and is associated with rising pine pollen abundances and the WRA tephra. This peak is slightly off-set when using spline interpolation and is included in the pine zone diagram.

The charcoal data for the pine zone, which is demarcated by the abrupt increase in pine pollen abundances in the record (Figure 6.12), are represented in Figures 7.19a and b below. The pine rise occurs relatively late at this site (~1100 cal yrs BP), after the WRA deposition. The surface sediments at Dragonfly Lake were unconsolidated, and as such the top is considered unreliable for the reconstruction of the charcoal record (the top ~20cm demonstrate reduced pollen influx values, Figure 6.13). The charcoal influx diagram (Figure 7.18a) for the pine zone displays the complete zone to the core surface. However, the record should be truncated by ~200 years due to the uncertainty in the sediment integrity, and the statistics represented in Table 7.3 reflect the truncated record.

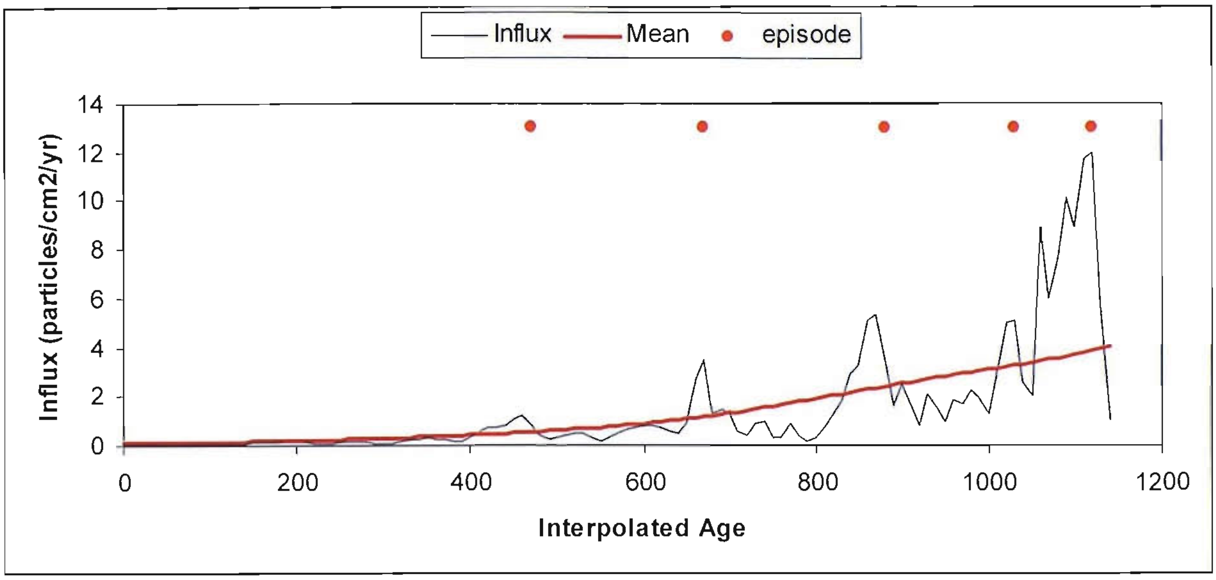


Figure 7.19a: Dragonfly Lake charcoal influx and fire episodes for the pine zone using linear interpolation. The mean is demarcated by the red line, and fire episodes by the orange dots.

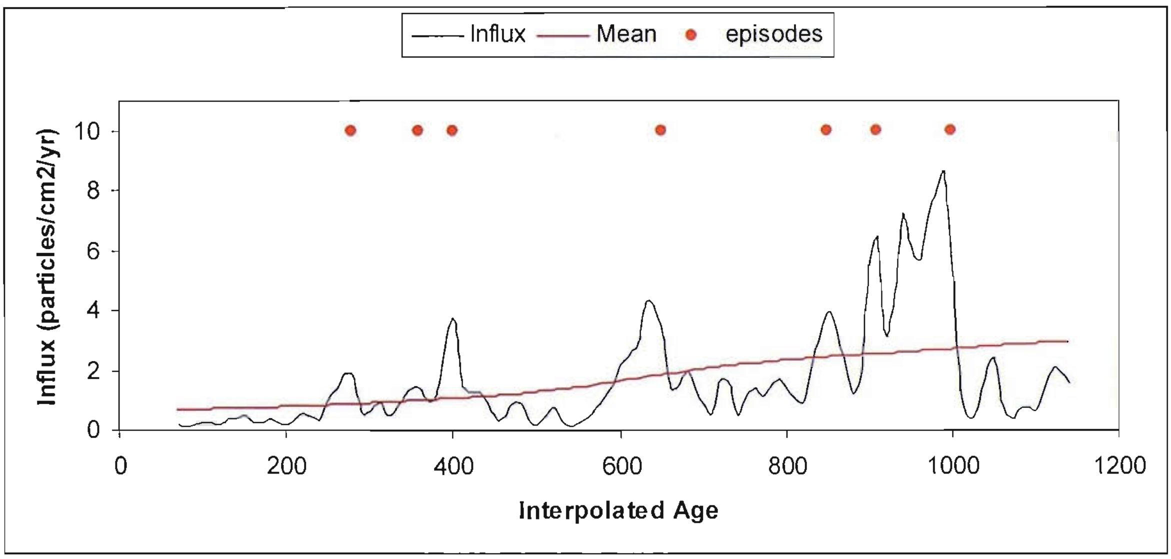


Figure 7.19b: Dragonfly Lake charcoal influx and fire episodes for the pine zone using spline interpolation. The mean is demarcated by the red line, and the point at which a peak exceeds the threshold by the orange dots.

The CHAPS output when using the spline interpolated age-model produces an altered charcoal influx diagram (Figure 7.18b), with slightly lowered peak magnitudes and influx values (Table 7.12), and a peak record that continues to ~200 cal yrs BP. The abrupt increase in charcoal influx is still present in the record when using spline interpolation, although the date at which it occurs is ~100 years younger than when using linear interpolation. This alteration in timing is explicable within the normal errors associated with the radiocarbon determinations, and the uncertainty associated with the use of polynomials. It is not considered critical for the interpretation of the influx data, although it will increase the errors underlying the timing of the event.

Due to the shortness of this record (~700 years once truncated), relatively little inference can be made on the fire regime at Dragonfly Lake during the period of pine dominance. There is a general decreasing trend in the charcoal influx values. The magnitude of the peaks are relatively high at the start of the record, and there is a distinct phase, between ~1150 and 1050 cal yrs BP, where the influx and peak magnitude are elevated compared with the end of the stomata zone and the remainder of the pine zone. This is caused by an inflection in the age-depth model due to a radiocarbon determination very close to the WRA deposit producing a nearly identical age (see Chapter 5), which coincides with the expansion of the pine population in the catchment. Influx values decrease from this point to 400 cal yrs BP where values are near zero. The average influx value for the zone as a whole is 2.2 particles/cm^2/yr, which is significantly higher than that of the preceding stomata zone; however the mean TBP is not significantly different.

The record shows that the fire episode frequency alters between the different vegetation zones (Figure 7.20), with an initial increase in fire episode frequency from the deciduous to spruce zone, and a general decreasing trend through the stomata and pine zones.

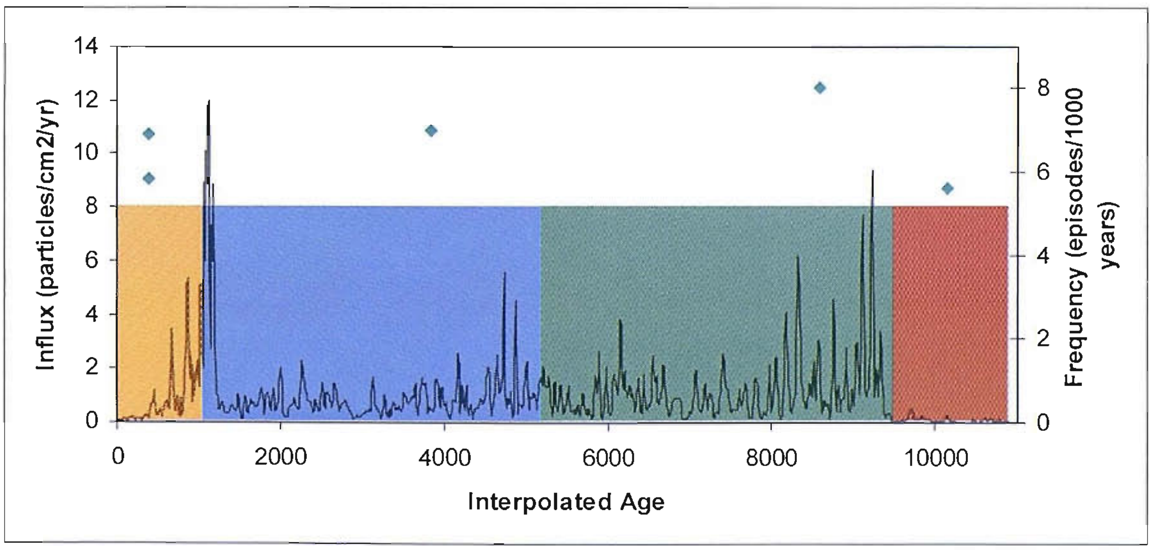
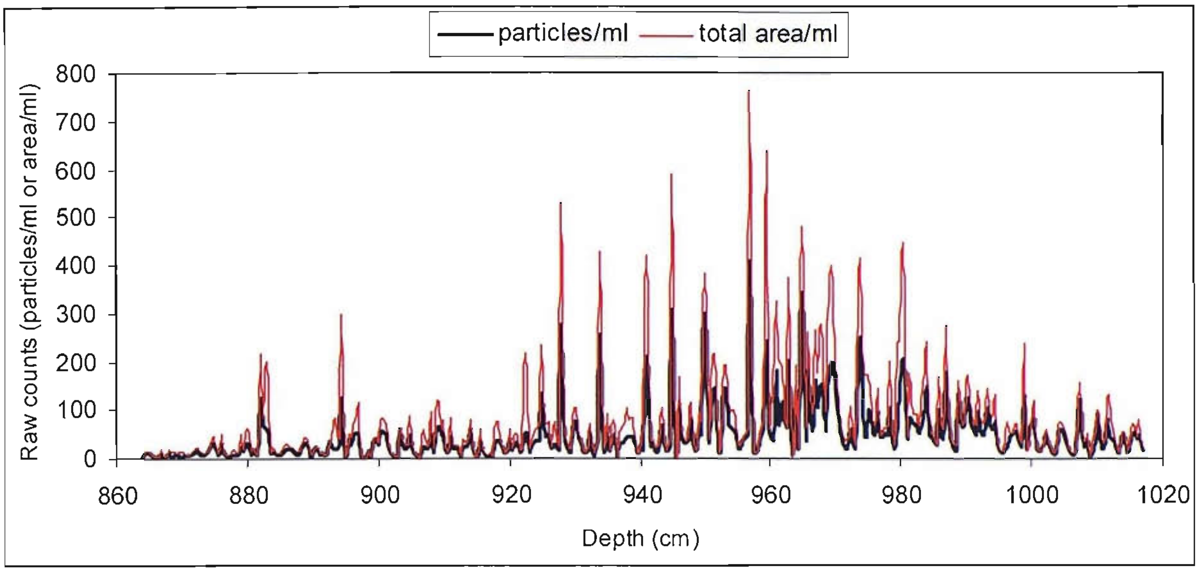


Figure 7.20: Dragonfly Lake charcoal influx and mean fire episode frequency for each vegetation zone (blue dots): shading - brown (deciduous), green (spruce), blue (stomata) and orange (pine). The two different age-models used to calculate the charcoal influx for the pine zone are represented by the two blue dots at the surface.

7.5.3 Haircut Lake

The charcoal record at Haircut Lake is restricted to the last ~8000 cal yrs BP (see Table 5.5, Chapter 5), therefore only the mid-spruce through to pine zones are represented. The raw charcoal counts and influx record for the whole core is represented in Figure 7.21. It is apparent that there are distinct phases within the influx data. The record begins with relatively low influx values and low magnitude peaks. At ~5000 cal yrs BP there is a significant gap in the peak occurrence, after which the mean influx appears to decrease, in addition to a coincident increase in the magnitude of the peaks. The record alters from ~2500 cal yrs BP, with a decrease in influx values and peak magnitude. There are two distinct large peaks within the last ~1200 years. This record is of particular interest as it displays two distinct types of peak morphology. Large magnitude acute peaks occur when the influx values exceed ~1 particle/cm2/yr and deposition times are low. When influx becomes very low, the peaks are broader, and of much reduced magnitude partly due to increased deposition times. The surface samples appear to be affected by dilution effects as the influx values decrease to near zero, with a distinct absence of peaks at the top of the core.



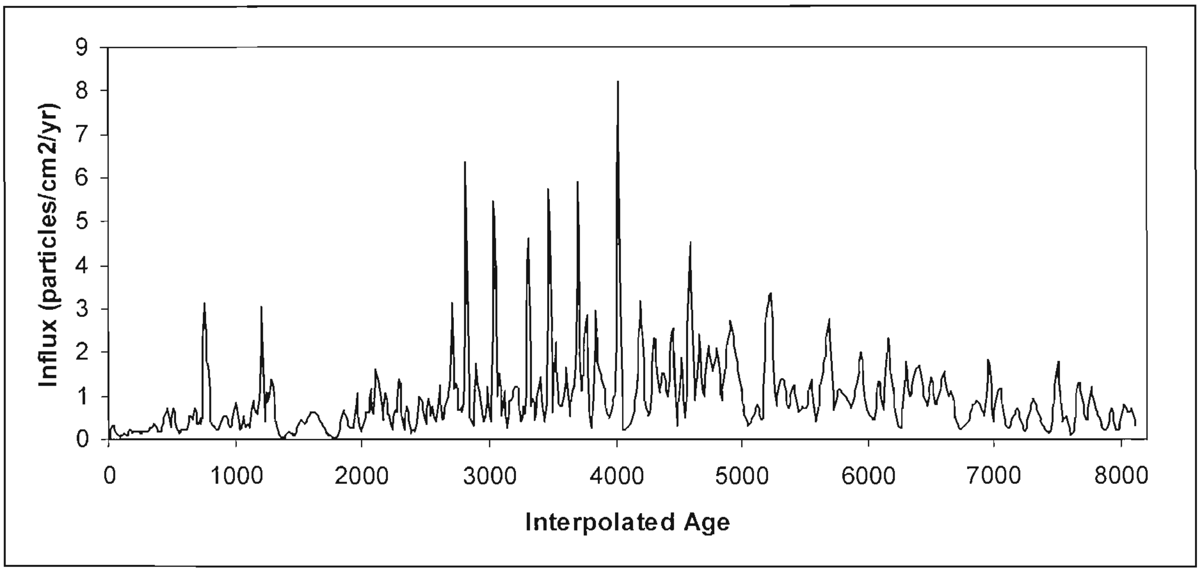


Figure 7.21: Haircut Lake, core A, showing above: raw charcoal counts against depth, and below: charcoal influx.

The record was split into vegetation zones as described on the pollen diagram (Figure 6.15). The rise in spruce pollen is absent, and the characteristic pollen assemblage of the spruce zone is present at the start of the record. It is probable that lake and the catchment vegetation were fully established prior to the start of sedimentation. The spruce zone therefore begins at ~8000 cal yrs BP in this record, and extends to ~6400 cal yrs BP (Figure 7.22).

The charcoal record for the spruce zone shows distinct peaks above the background, with the average influx at 0.68 particles/cm^2/yr (Table 7.13). The mean influx demonstrates an overall increasing trend through the zone.

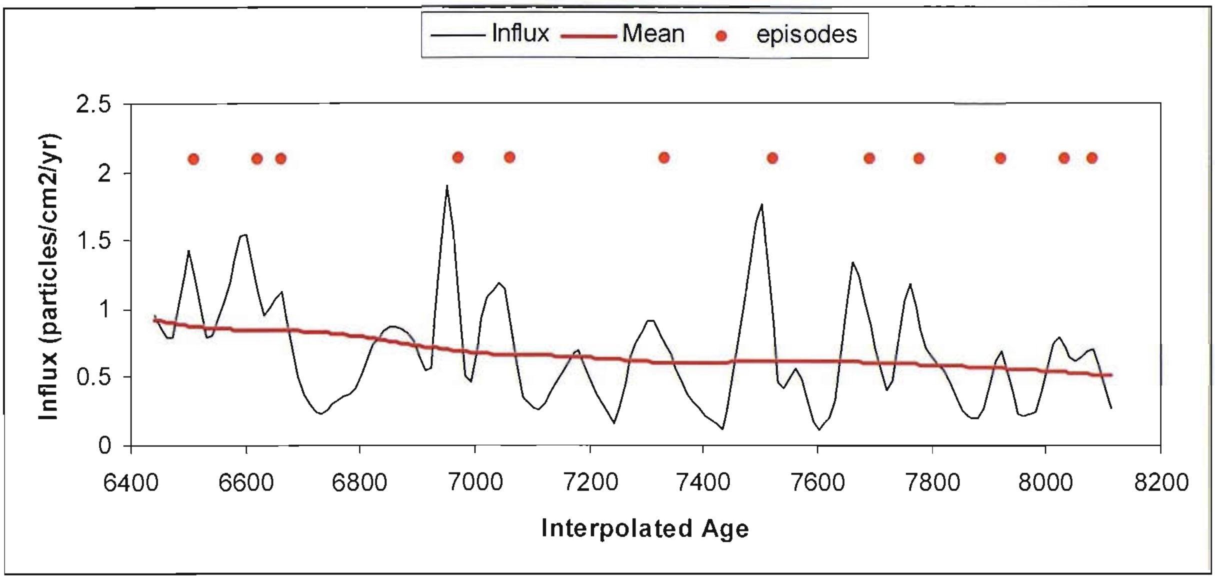


Figure 7.22: Haircut Lake, core A, showing charcoal influx for the spruce zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

There are well defined peaks in this zone, interspersed with areas of comparatively little charcoal influx. These periods cover significant time spans in places, and appear as quiescent periods in the episode frequency record. However, with this in mind, the mean TBP is still calculated as ~180 years (Table 7.13), which is comparable to contemporary fire return frequencies for spruce dominated communities (Viereck, 1973; Van Cleve *et al.*, 1986).

Vegetation	Threshold	Mean Time Between Peaks	Fire Frequency (#/1000 yrs)	Fire Episodes in zone	Average Influx (particles/cm2/yr)
Spruce	1.2	181.3	7.1	12	0.68
Stomata	1.1	157.6	8.0	25	1.41
Pine	1.1	131.2	9.6	30	0.66

Table 7.5: Haircut Lake fire regime statistics for each vegetation zone

The transition to the stomata zone at ~ 6400 cal yrs BP, shows little variation from the end of the spruce zone. The influx values remain at ~1 particle/cm^2/yr, with numerous peaks identified (Figure 7.23). From ~5900 cal yrs BP to ~5000 cal yrs BP there is a period of instability in the record. There are periods with very low influx, creating quiescent periods in the peak record, interrupted by two relatively high magnitude peaks. After ~5000 cal yrs BP, there is a significant increase in the frequency of peaks identified, associated with a slight increase in the mean influx, and an increase in peak magnitude.

This is due to an increase in temporal resolution linked to a change in the sedimentation rate.

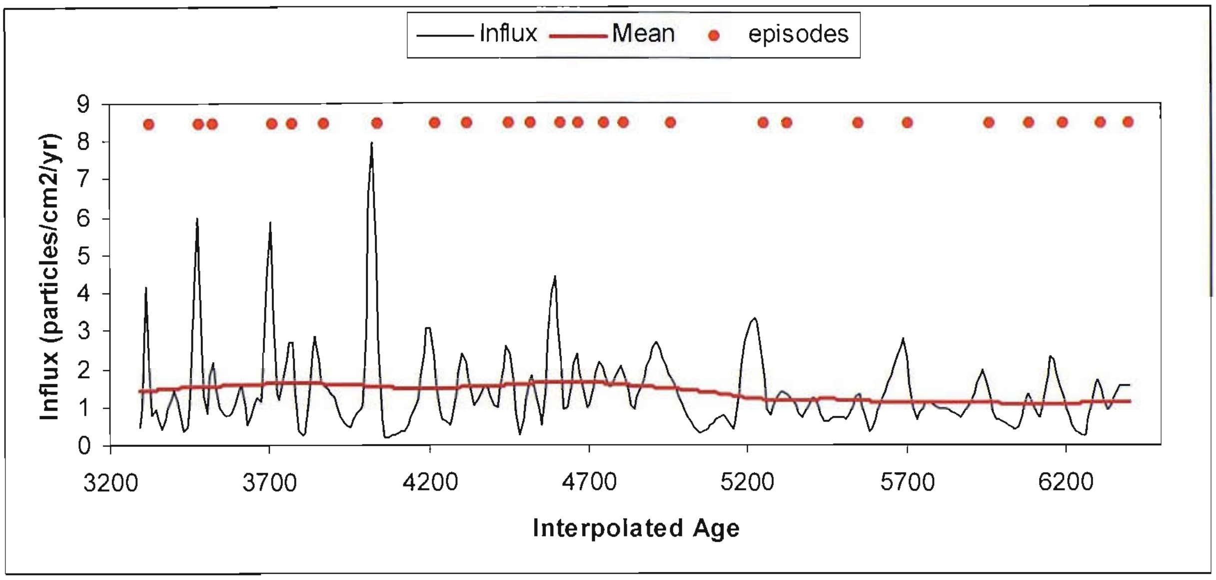


Figure 7.23: Haircut Lake, core A, showing charcoal influx for the stomata zone. The mean is demarcated by the red line, and fire episodes by the orange dots.

The average influx is greater in the stomata zone than the spruce zone (1.41 particles/cm^2/yr [Table 7.5]), due primarily to the large magnitude peaks in the second half of the zone. However in general the mean influx remains relatively stable throughout the zone (red line in Figure 7.23). The mean TBP is slightly less than the spruce zone at ~157 years between episodes.

The frequency of large magnitude peaks and an influx of >1particle/cm^2/yr continues across the transition to the pine zone (Figure 7.24). At the beginning of the zone, until ~2000 cal yrs BP, the peak frequency remains relatively high, although the peak magnitude decreases significantly over this period. The mean influx gradually decreases through the pine zone, with an average influx of 0.66 particles/cm^2/yr (Table 7.5). Between ~2000 and 1400 cal yrs BP there is a distinct lowering of influx values, and only 2 peaks identified above the threshold level.

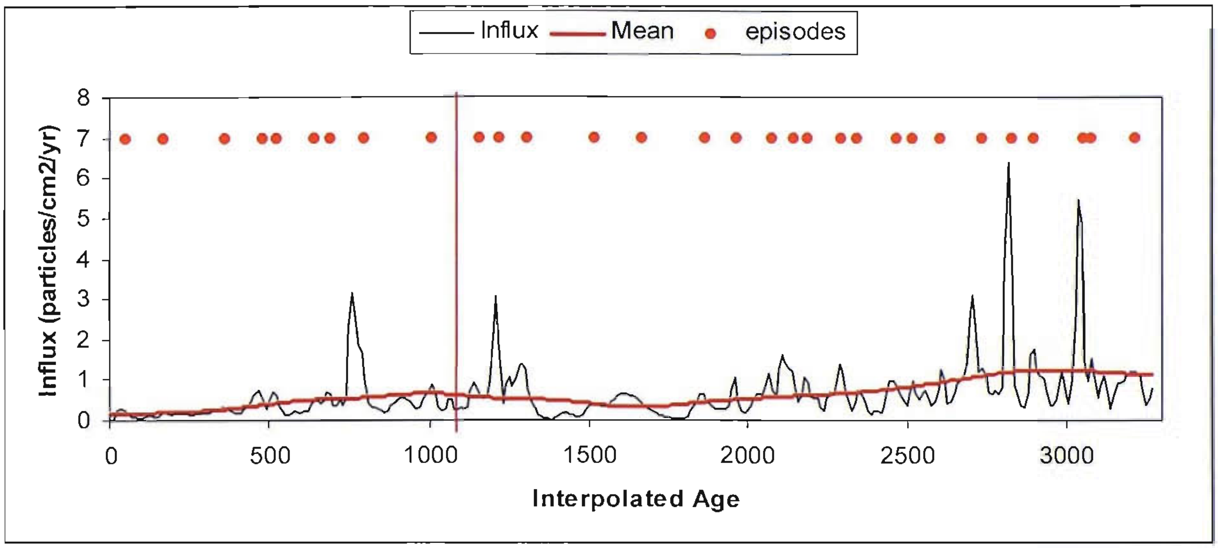


Figure 7.24: Haircut Lake, core A, showing charcoal influx for the pine zone. The mean is demarcated by the red line, and fire episodes by the orange dots. The WRA is marked by the red vertical line at ~1150 interpolated yrs BP.

The mean influx increases slightly after ~1400 cal yrs BP, coincident with an increase in peak magnitude and peak frequency, until ~700 cal yrs BP, where the mean influx begins to decrease towards the present. The very low influx levels at the core surface are due to dilution effects caused by unconsolidated sediments.

The mean TBP in the pine zone was calculated at ~131 years, which is less than either of the two preceding zones. Overall the fire episode frequency displays a general increasing trend from the spruce through to the pine zones, reflected by the individual zone mean TBP values (Figure 7.25).

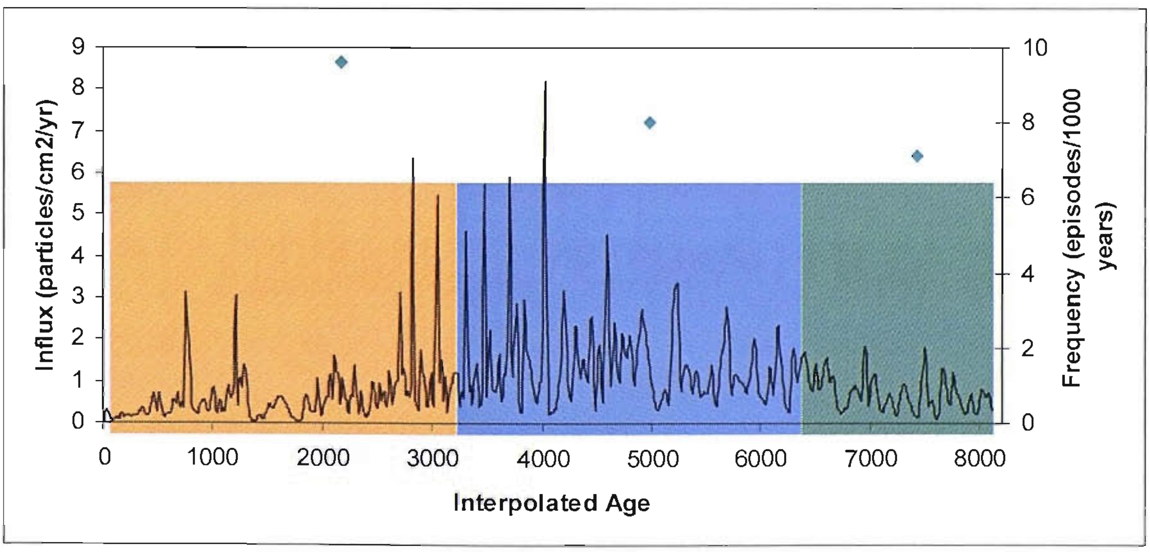


Figure 7.25: Haircut Lake, core A, showing charcoal influx for the whole record, with mean fire frequency plotted for each vegetation zone (blue dots). Shading – green (spruce), blue (pine), and orange (pine).

7.4.6 Marcella Lake

The charcoal record for the amalgamated core A-Z at Marcella Lake is shown in Figures 7.26a,b and c. There is a distinct discrepancy between the CHAPS output of charcoal influx between the two age-depth models used at the top of the core. The main body of the record remains broadly similar regardless of the age-model used.

In the age-depth relationship (Chapter 5) there is a discrepancy between the age of the WRA tephra and the radiocarbon determinations adjacent to it, creating an abrupt shift in sedimentation at this point (see Chapter 5). There is limited, if any, evidence to support this shift within the core lithostratigraphy, except for an increased production of carbonates. This inflection in the age-depth curve has the effect of producing the pronounced increase in influx values at ~1200 cal yrs BP in Figure 7.26b. In Figure 7.26c a spline interpolated model has been used to calculate CHAR, which illustrates that the influx event is a function of linear interpolation rather than the charcoal data. There is a slight increasing trend in the influx values over this period, but a distinct absence of abrupt increases in influx and extremely high peaks. For this reason the record will be interpreted using the spline-interpolated model.

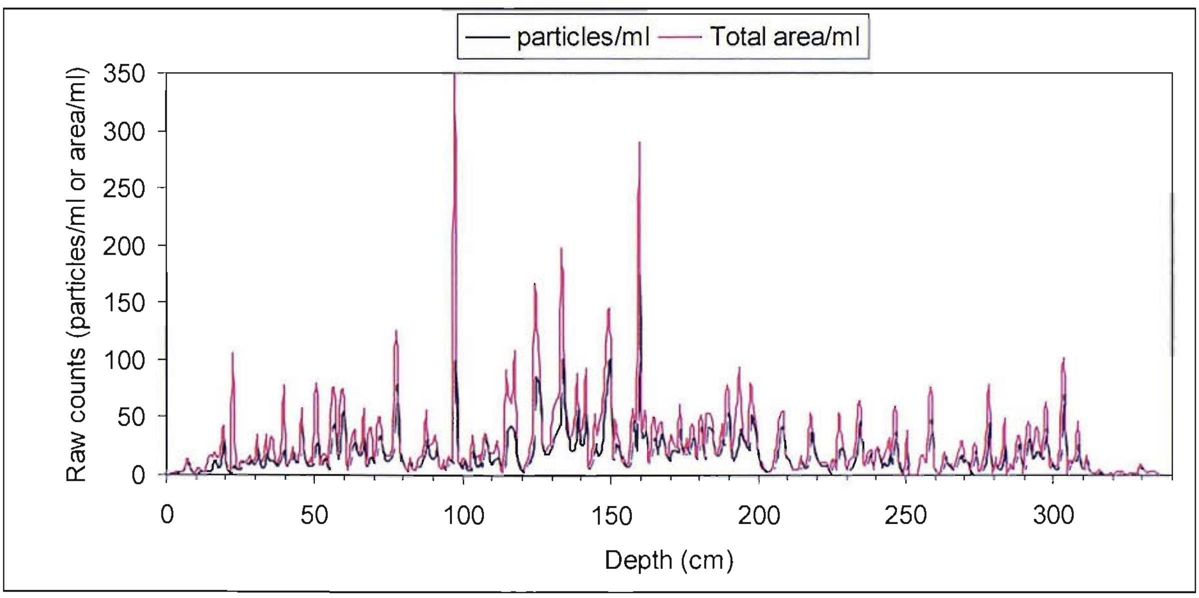


Figure 7.26a: Raw counts of charcoal particle and area per ml for Marcella Lake core A-Z

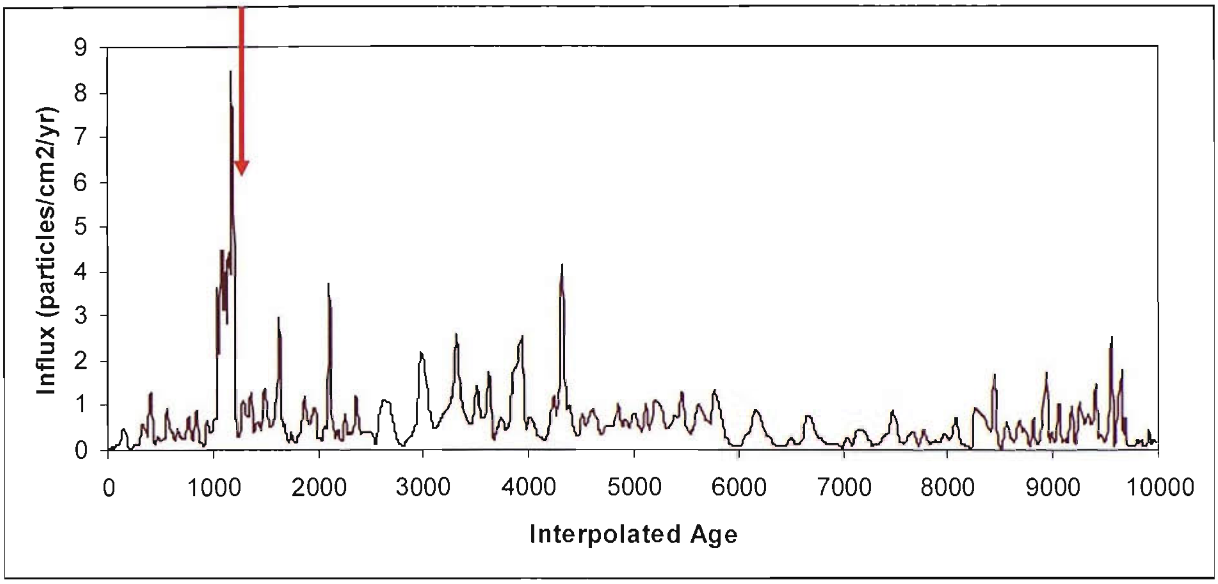


Figure 7.26b: Marcella Lake, core A-Z, charcoal influx diagram using linear interpolated age-depth model

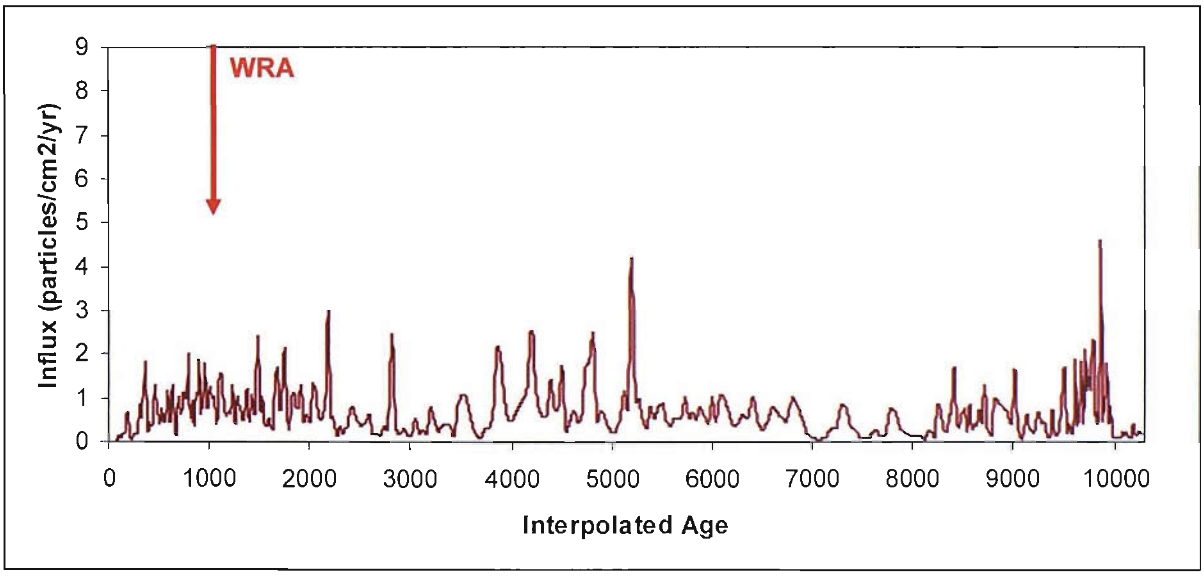


Figure 7.26c: Marcella Lake, core A-Z charcoal influx using spline interpolated age-depth model. Note: y-axis has been scaled to match the y-axis in Figure 7.26a.

In this record a short period of low influx values is observed prior to ~10000 cal yrs BP. From this point onwards a marked step increase in influx values is coincident with an increase in peak magnitude. From ~8100 cal yrs BP, the record becomes more complacent, with lowered mean influx, and broader peaks. A second step increase in influx is observed at ~6900 cal yrs BP, although the peaks remain relatively broad, with lowered magnitudes until ~5200 cal yrs BP.

Between ~5200 and ~2500 cal yrs BP, the influx values gradually decrease. However, the peaks retain their magnitude. After 2500 cal yrs BP the record displays an overall increase in influx values associated with small magnitude peaks.

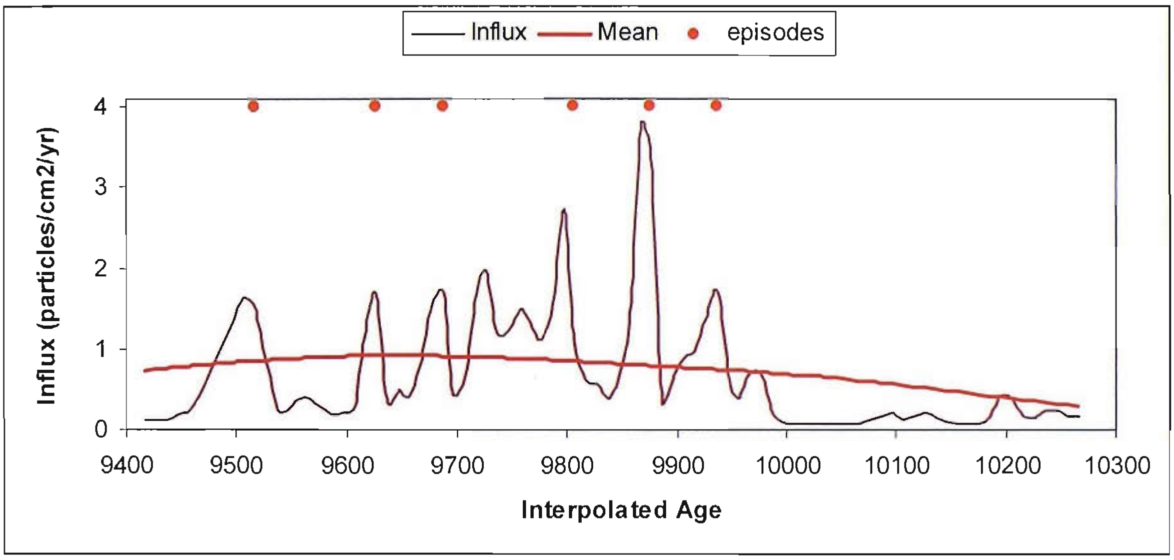


Figure 7.27: Marcella Lake, core A-Z, charcoal influx diagram for the deciduous zone. Fire episodes identified above the 1.1x threshold. The mean is demarcated by the red line, and fire episodes by the orange dots.

The deciduous zone is very short in this record (Figure 7.27), lasting less than 1000 years. At the start of the record the influx values are extremely low, with little variation. As discussed for other sites, the influx values are very low, with zero values in the raw counts (Figure 7.26a). As it is unlikely for there to be no background component, some of the peaks must be a function of noise within the data, and probably not an accurate identification of fire. From ~9900 cal yrs BP the charcoal influx increases, and distinct peaks can be identified. Over the whole zone the running mean displays a slight increasing trend, but the mean influx is low at 0.68 particles/cm^2/yr (Table 7.16). The increasing mean is likely influenced by the initial near zero values pre-10ka, and the shortness of the zone.

The mean TBP was calculated at ~190 years (Table 7.16). However, due to the shortness of this record, it appears impossible to assess whether this is an accurate estimate of fire episode frequency.

Vegetation	Threshold	Mean Time Between Peaks	Fire Frequency (#/1000 yrs)	Fire Episodes in zone	Influx (particles/cm2/yr)
Deciduous	1.1	190.9	7.63	6	0.68
Spruce	1.2	283.0	5.3	19	0.48
Stomata	1.1	256.1	4.7	14	0.78
Pine	1.1	115.7	10.9	28	0.74

Table 7.6: Marcella Lake core A-Z, fire regime statistics for each vegetation zone.

From ~9500 cal yrs BP the mean influx appears to decrease slightly towards the transition to the spruce zone (Figure 7.28). From ~9400 cal yrs BP there is a relatively high frequency of peaks with a slightly increased mean influx. After ~8300 cal yrs BP the mean influx decreases the peak frequency become less frequent and the peaks broader. The temporal resolution over this period decreases gradually and is probably influencing the peak characteristics. From ~6800 cal yrs BP the peak frequency appears to gradually increase to the end of the zone at ~5800 cal yrs BP.

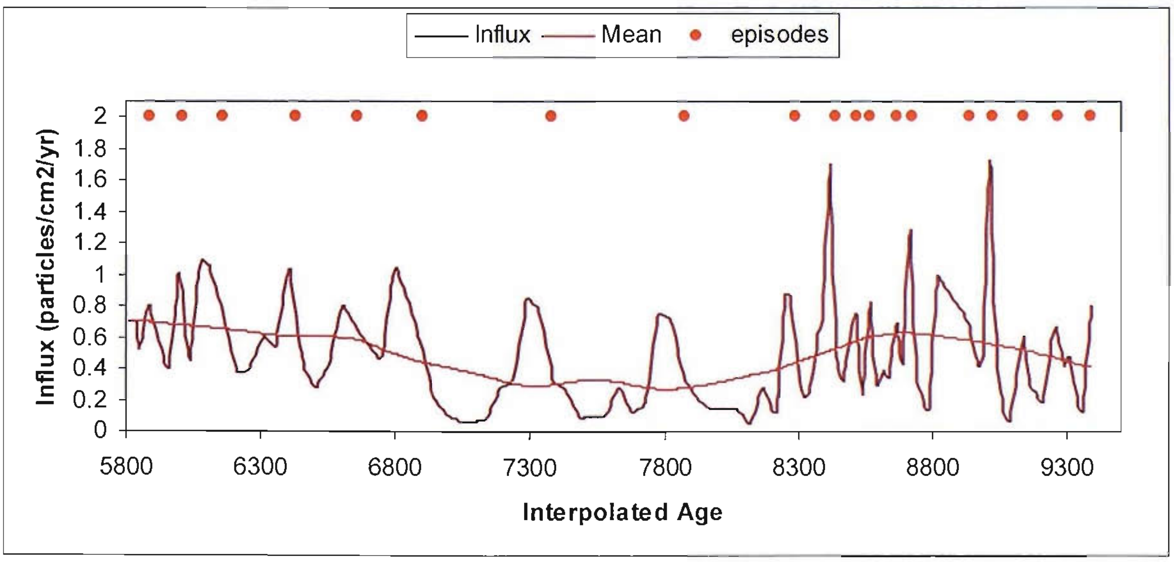


Figure 7.28: Marcella Lake, core A-Z, charcoal influx diagram for the spruce zone. Fire episodes identified above the 1.1x threshold. The mean is demarcated by the red line, and the point at which a peak exceeded the threshold by orange dots.

The average influx is lower than that of the deciduous zone at 0.48 particles/cm^2/yr (Table 7.6), with the mean TBP as ~283 years. This is much longer than contemporary observations of spruce dominated forests ~100-130 years (Viereck, 1973, 1983; Van Cleve *et al.*, 1986; Johnson, 1992).

The transition to the stomata zone (Figure 7.29) sees a slight increase in the mean influx values. Thepeak frequency appears to remain relatively steady, but with a short cluster of peaks at ~5250 cal yrs BP. From ~4300 cal yrs BP, the mean influx values start a decreasing trend to the end of the zone. From ~300 cal yrs BP, the record displays a decrease in influx values, and an eventual tailing off in values to the surface.

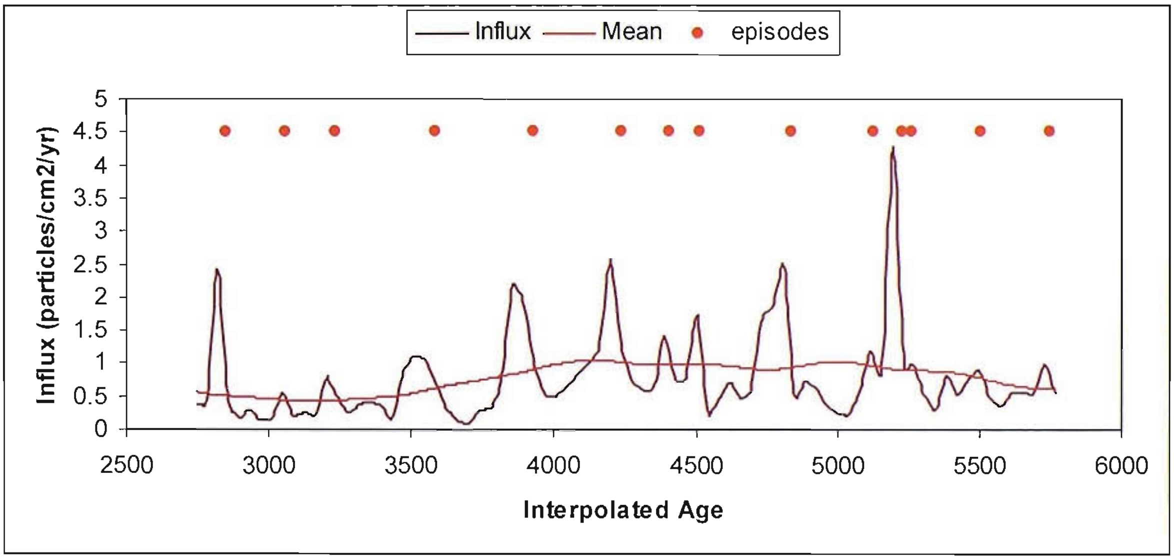


Figure 7.29: Marcella Lake, core A-Z, charcoal influx diagram for the stomata zone. Fire episodes identified above the 1.1x threshold. The mean is demarcated by the red line, and the start of a peak exceeding the threshold by the orange dots.

Although the mean influx is greater in this zone compared to the spruce zone (Table 7.6), the mean TBP remains relatively unchanged at ~256 years.

The transition to the pine zone is characterised by a gradual increase in mean influx and a subsequent increase in peak frequency (Figure 7.30). From ~2400 cal yrs BP, the record displays relatively large magnitude peaks, of increasing frequency.

The fire regime statistics represented in Table 7.6 indicate that the mean TBP is much lower than the rest of the record at ~115 years, which is more comparable to contemporary estimates of fire return frequency in pine/spruce forests (Viereck, 1973, 1983; Van Cleve *et al.*, 1986)

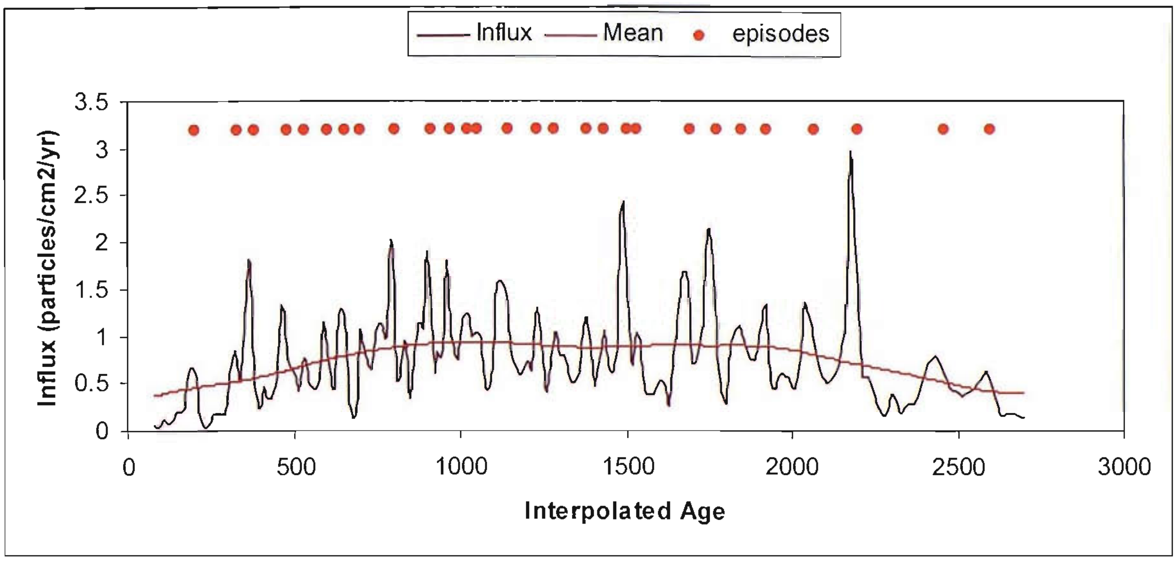


Figure 7.30: Marcella Lake, core A-Z, charcoal influx diagram for the pine zone. Fire episodes identified above the 1.1x threshold. The mean is demarcated by the red line, and fire episodes by the orange dots.

The record for Marcella Lake core A-Z shows a decreasing trend in fire episode frequency from the deciduous to the stomata zones (Figure 7.31). It is possible that the deciduous zone fire frequency estimate is not a true reflection of reality due to the shortness of the record within this zone, and the extremely low influx values. The stomata zone has the lowest peak frequency due to a low temporal resolution. The peak frequency then increases into the pine zone to over 10 fire episodes per 1000 years.

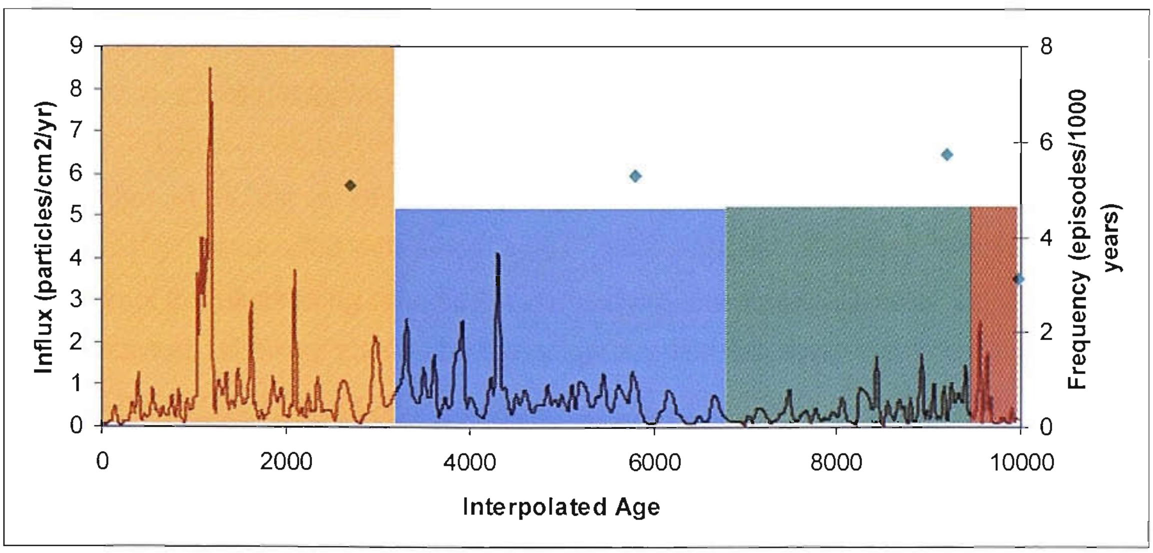


Figure 7.31: Marcella Lake, core A-Z charcoal influx and fire episode frequency for the whole record. Shading represents the deciduous (brown), spruce (green), stomata (blue) and pine (orange) zones.

7.5.3 *Jellybean Lake*

The charcoal record for Jellybean Lake is represented in Figure 7.32. Due to the absence of pollen data for this site, the record was analysed as one complete zone. The record demonstrates two distinct phases. From the start of the record at ~7000 cal yrs BP to ~4100 cal yrs BP it is dominated by large magnitude peaks in conjunction with high fire episode frequency. The influx varies over this period, but in general remains <2 particles/cm^2/yr, with a short-lived depression in values between ~5400 and 5500 cal yrs BP.

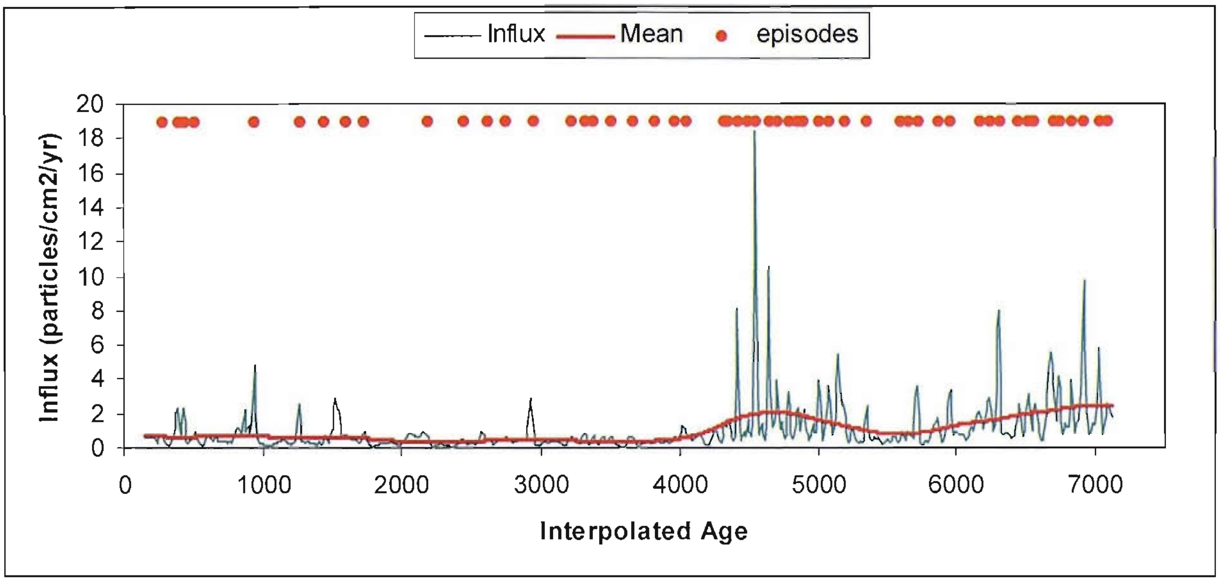


Figure 7.32: Jellybean Lake, showing charcoal influx, running mean and identified peaks above a 1.1x threshold for the whole record.

After ~4100 cal yrs BP the mean influx decreases significantly to <1 particle/cm^2/yr, associated with a brief hiatus in fire episodes identified. This is most likely a function of the decreasing running mean masking any potential peaks. Once the running mean becomes relatively stable, the fire episodes identified increase in frequency, although not to the level preceding 4100 cal yrs BP. The peaks identified are of lower magnitude, with infrequent relatively high magnitude peaks occurring mostly between ~1800 – 900 cal yrs BP.

The record ends before 0 cal yrs BP, as to the top 15cm of the core being unsuitable for charcoal analysis, therefore any potential surface related issues such as dilution effects are not considered to affect this record.

7.6 Summary

In this Chapter the charcoal results for each site have been explored. There are difficulties associated with interpreting records with low influx, as is the case in the deciduous zone of all the records; however it is felt that the threshold-plateau method has provided an objective way of interpreting the peaks.

The deciduous zone of all the sites produced low influx values, and relatively few peaks that were confidently interpretable as fire episodes. The Yukon sites displayed rapid increases in charcoal influx in the spruce zone, associated with increased peak frequency. The Alaskan sites showed a slight decrease in the charcoal influx values from deciduous to spruce zones, with significant alterations in the peak frequency, although not as apparent as the Yukon sites. The transitions from the spruce to the stomata and then pine zones in the Yukon lakes showed relatively little reaction in the peak frequency distribution to potentially altered vegetation type, and inter-site comparisons are necessary to draw conclusions on the trends within the records. The records will be fully compared, discussed and interpreted in Chapter 8 in order to construct a regional synthesis.

Chapter 8. Discussion

8.1 Introduction

The main aim of this project was to test hypotheses about how climate, fire, and vegetation interact to bring about large-scale change in the boreal forest ecosystem of Alaska-Yukon using palaeodata. The main hypotheses to test were outlined in Chapter 1, and are as follows:

- *H1. Climate change drives fire regime, which alters vegetation type*

This will be tested against the transition to spruce from deciduous vegetation at all sites (except Jellybean, as there is no pollen record at present, Jellybean and Haircut do not extend beyond ~8 ka BP). In order to accept this hypothesis the climatic shift (equivalent to a 20-35% increase in precipitation; Barber and Finney, 2000) at ~10 ka BP should have suppressed regional burning leading to a switch in the boreal forest composition from deciduous taxa to spruce. Therefore, a reduction of the fire regime should be observable across the transition.

- *H2. Climate driven vegetation change alters the fire regime.*

This can be tested against both vegetation transitions. The dominant vegetation types have differing flammabilities (deciduous < spruce < pine), and therefore against the deciduous-spruce transition, a shift in vegetation type in response to altered climate would precede a change in the fire regime. Thus, the deciduous to spruce transitions should result in an increased fire return frequency.

The spruce-pine transition is not associated with a clear climatic shift, but rather with late Holocene climatic variability; it nevertheless represents a major change in forest composition. A shift to a more flammable dominant species should result in an alteration in the fire regime. To accept this hypothesis an increase of fire across the vegetation change should be observed, based on contemporary flammability observations.

Some of the interactions between climate, vegetation and fire are established (solid arrows in Figure 8.1). Climate exerts a direct control over vegetation through physiological and reproductive geographic range limitations. Climate impacts the fire regime through fire season weather patterns, as shown in recent inter-annual fire variability studies (Amiro *et al.*, 2003, and Chapter 2), in addition to having direct influence over fuel dryness and ignition sources (Amiro *et al.*, 2003, and see Chaper 2).

It has been established through modern observation of different forest types that vegetation structure influences the availability of fuel and the potential flammability of a landscape (e.g. Gedalof *et al.*, 2005, and Chapter 2). Deciduous woodland and spruce and pine forest are associated, as far as it is possible to determine, with different mean fire return intervals. Differences are partly a function of the varying flammability of the different species: *Pinus* > *Picea mariana* > *P. glauca* > deciduous taxa (Johnson, 1992). However, deciduous taxa appear to be associated with the most drought- and fire-prone sites. The deciduous species that occur in the boreal forest environment were categorised as invaders, pioneer species that perpetuate well on disturbed landscapes, and endurers due to the ability to vegetatively reproduce, which would partly explain their abundances on these sites. Paper birch, and trembling aspen are typically associated with south-facing slopes as they require warmer soil temperatures, and higher nutrient regimes, and in the absence of disturbance, are typically replaced by white spruce. Therefore, although the deciduous species may be less flammable than the coniferous species, in this environment today they are associated with higher fire return frequencies which retains the landscape in an early successional stage. Vegetation can also influence the climate system via effect on albedo for example (Bonan *et al.*, 1992; Kutzbach *et al.*, 1996).

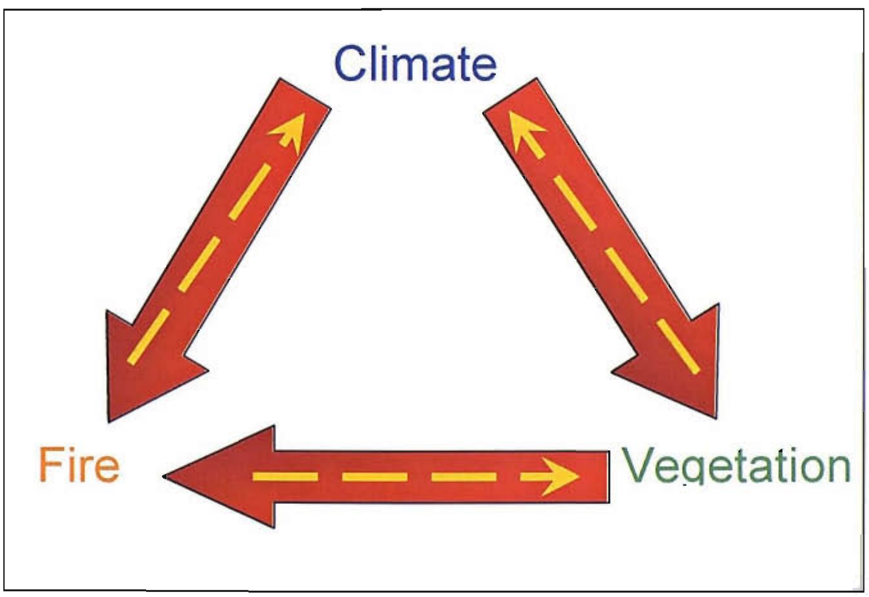


Figure 8.1: Conceptual model of the interactions between climate, fire and vegetation. Solid arrows represent established direct controls, dotted arrows represent indirect/unknown controls. (Adapted from (Lloyd et al., 2006))

Fire, especially on the scale regularly occurring in the boreal forest, has an obvious influence on climate due to the release of vast quantities of carbon, and other gaseous products of combustion to the atmosphere (Kasischke, 2000; Chambers & Chapin, 2002), altering atmospheric chemistry and resulting in climatic feedbacks. It is possible that in response to altered climate via feedbacks, the fire return frequency of a landscape can have a direct influence on vegetation through successional forcing (Kasischke *et al.*, 2000; Chapin *et al.*, 2004a; Chapin *et al.*, 2004b), and/or habitat removal.

As interactions occur on decadal to century time scales, modern observations of forest dynamics lack temporal depth to show what happens during a transition, and 20th Century fire-climate records do not encompass large-magnitude climate variability and therefore are an inadequate basis upon which to estimate future changes. This study provides an opportunity for long-term reconstructions to further inform the model (arrows in Figure 8.1) for future responses to climate change at high latitudes. In this chapter the charcoal and vegetation results are discussed in conjunction with established palaeohydrologic and palaeoclimatic data to explore the interaction of fire, vegetation and climate over the two main vegetation transitions. Inter-site comparisons are made to produce a regional reconstruction. Timings of the shifts in vegetation and fire regime in relation to lake-level records are examined to provide an indication of the hierarchy of controls governing the interaction of the three main variables.

8.2 Deciduous to Spruce Transition

This section will compare the charcoal results from the individual sites with each other to form regional reconstructions, against a climatic and a vegetation shift. The author presents palaeohydrological data (Figures 8.2 and 8.4), with permission, of lake levels during the Holocene for Marcella (Anderson *et al.*, 2005) and Birch Lake (Abbott *et al.*, 2000; Anderson *et al.*, 2005b). From comparison with other sites for example Windmill Lake (Bigelow, 1997; Bigelow & Edwards, 2001), Dune, Jan and Sands of Time Lakes (Edwards *et al.*, 2001) and off-shore sediment core transects (Abbott, 1996) these lakes have been shown to provide a regional signal of moisture balance. Lake levels increased abruptly from ~10ka linked to a climatic shift to increased effective moisture (Abbott *et al.*, 2000; Barber & Finney, 2000; Anderson *et al.*, 2005b) coincident with the regional expansion of spruce.

8.2.1 Yukon Territory Reconstruction

Marcella Lake represents the lake-level record for the Southern Yukon Territory (Figure 8.2). At Marcella Lake, a rise in lake level begins at ~10250 cal yrs BP, almost reaching modern levels, when allowances are made for the smaller sediment package present, by ~9750 cal yrs BP. The period of time over which the lake-level increases is interpreted as a period when the Southern Yukon experienced a significant increase in effective moisture (Rampton, 1971; Schweger, 1997; Anderson *et al.*, 2002; Anderson, 2004; Anderson *et al.*, 2005b). This is further supported by lake-levels records from two other southern Yukon lakes (north of Whitehorse) (B. Finney, M. Abbott, M.E. Edwards and L. Anderson, unpublished data). The Marcella lake chronology of moisture change is well established and has been used for the interpretation of other sites (Lynch *et al.*, 2002; Shuman *et al.*, 2002; Lynch *et al.*, 2004; Hallett & Hills, 2006)

The charcoal records for the sites that contain the ~10ka spruce rise/climate shift in the Yukon Territory (Marcella, Salmo and Dragonfly Lakes) display similar responses (Figure 8.3). In all three sites the rise in moisture inferred from the lake-level increase begins prior to the expansion of spruce in the catchment. The appearance of spruce in all the pollen records (see Chapter 6) falls within a ~300 year period, and can be generally referred to a synchronous expansion, especially in consideration of the errors associated with radiocarbon dating.

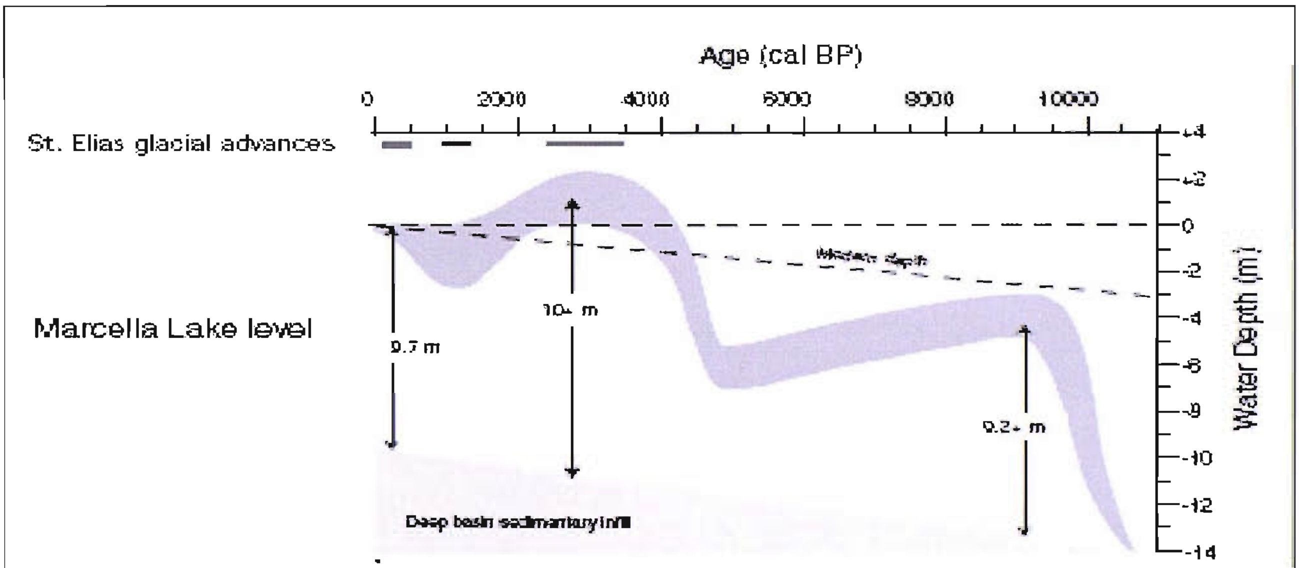


Figure 8.2: Correlation on a calibrated radiocarbon timescale of Marcella Lake water levels (Anderson et al., 2005b) with glacier activity in the St. Elias Mountains (Denton & Karlen, 1977). Reproduced with permission from L. Anderson (Anderson, 2004).

All of the lakes produced very low charcoal influx values within the deciduous zone. For all of the lakes the records likely reflect largely an extra-local or regional fire signal due to the lack of clear separation between the background and peak components (See Chapter 7). Relatively few of the peaks identified can be attributed with confidence to local fire episodes. At all of the sites, primarily due to this effect, the estimated mean fire return time is long. The charcoal records show a distinct increase in influx values shortly after or at the spruce rise. At Marcella Lake, the increase in peak frequency is synchronous with the increase in influx values in the record shortly before the appearance of spruce in the record. In the case of Salmo Lake, there is a gradual increase in peak frequency after the spruce rise. At Dragonfly Lake there is an almost immediate increase in peak frequency at the spruce rise (Figure 8.3)

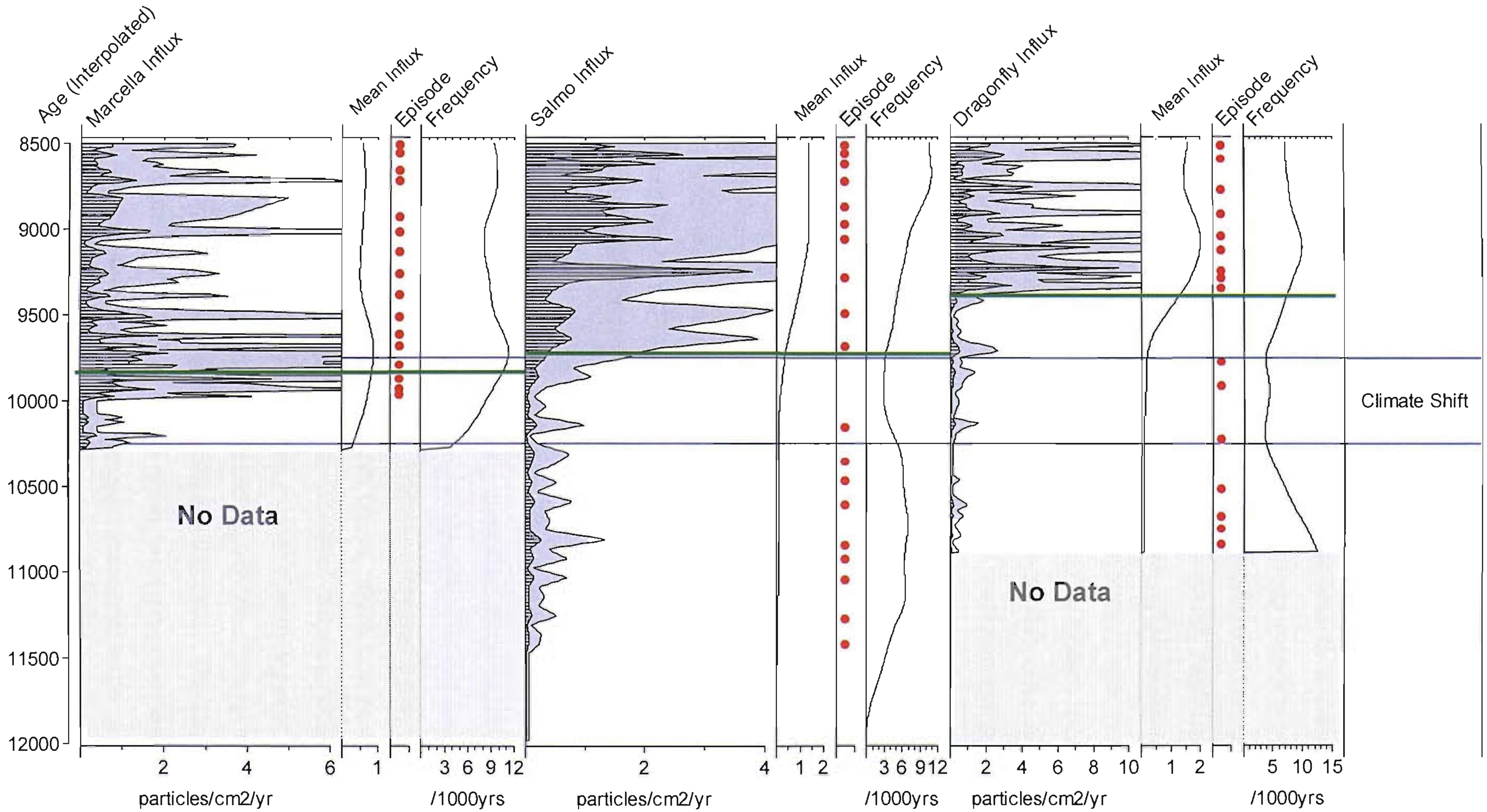


Figure 8.3: Charcoal influx data across the spruce rise for the Yukon sites (Marcella, Salmo and Dragonfly). The blue band denotes the lake level rise (Figure 8.1) The green lines represent the rise of spruce in the pollen record above 15% (see Chapter 6).

The deciduous zone pollen records at Marcella and Dragonfly Lake (Figures 6.12 and 6.20) display pollen abundances typical of other regional sites during the deciduous period (Rampton, 1971; Cwynar, 1988; Schweger, 2001). When taking into consideration site topography, geology (Chapter 4) and pollen influx values (Chapter 6) it would appear that the vegetation was relatively well established in the catchments at these two sites and was most likely a birch scrubland consisting of open woody vegetation and relatively high proportions of grasses and herbs. The deciduous zone record at Salmo Lake displays characteristics of an establishing lake system within a recently deglaciated catchment dominated by herbaceous taxa. The rapidly rising pollen influx values and physical sediment property changes over this period lead to uncertainty over the interpretation of the low charcoal influx values at this site. Such an unstable system is unlikely to have an established fire regime characteristic of the regional deciduous zone, and the charcoal record is likely strongly influenced by the low temporal resolution within this zone.

If we use the most conservative method for the interpretation of the deciduous zone charcoal record overall, based on modern fire ecology theory for fires in deciduous taxa dominated environments, we must apply a relatively high threshold, which identifies very few charcoal peaks in this zone. In this situation it is assumed that a background component is still present, which must be removed from the record, and a threshold applied which reduces the amount of noise included in the peak component, which produces a very long mean TBP estimate.

Alternatively, the low charcoal influx values could be the result of low amounts of available biomass to burn compared with a coniferous-forested landscape, and the comparatively low flammability potential of the catchments at this time that would have resulted in the predominance of low intensity ground fires. It has been shown that fire detection using sedimentary charcoal is strongly dependent on fire severity (Higuera *et al.*, 2005), and that fine-scale spatial patterns of lower-severity burns play an important role in determining the charcoal signal of these events (Agee, 1993). Low severity fire detection is further hampered by the relatively short transportation distances of charcoal particles due to weaker winds, and less buoyant smoke plumes (Sandberg *et al.*, 2002). The patchy spatial patterns of low severity fires could explain the charcoal influx values and peak frequency observed in the deciduous zones of this study. The temporal resolution of the records has a significant influence on the ability to decompose these low influx values with confidence. A higher temporal resolution is achievable at some of the sites (Alaskan and Dragonfly Lakes), but due to the sedimentation rate (>50 years/cm) at this level in the cores, it is suspected that the increased resolution of ~20 yrs/cm would still be too coarse to remove the uncertainty completely.

If the low severity fire, low charcoal influx approach is a possibility, it would be more prudent to apply a lower threshold to try and identify the much smaller signal expected to be produced by the low intensity ground fires. This approach includes much of the low-level variation in the deciduous zone record, and produces more identified charcoal peaks, however the mean TBP still remains relatively long, and in either case the mean TBP is longer than that of the spruce zone.

A Kolmogorov-Smirnov two-sample test was conducted using the amalgamated deciduous zone data to increase the length of the data set, and the robustness of the record. The comparative peak frequency estimates for the deciduous zone interpreted with the higher threshold, more conservative estimate approach, was significantly different from that of the spruce zone ($p>0.05$). Reducing the threshold to identify more peaks, under the assumption that low intensity ground fires are more difficult to detect in the sediment record, there was still significant difference between the mean TBP and fire episode frequency of the deciduous and spruce zones ($p<0.05$), with the spruce zone having a more frequent charcoal peak return time. Therefore, the data suggests that the spruce zone must have burnt more frequently in the early Holocene than the deciduous zone.

8.2.2 Alaskan Reconstruction

The Birch Lake water level reconstruction of Abbott *et al., (2000)* (Figure 8.4) was used for the interpretation of the charcoal records in Interior Alaska. The record is longer than the record of effective moisture changes at Marcella Lake, and the ~10ka moisture shift is easily identifiable. It should be noted that although Birch Lake was a relatively large and established lake at this point with already deep water levels, it is considered to have been a good closed-system amplifier of precipitation in the early Holocene (Barber & Finney, 2000), as the lake has a relatively large drainage basin compared with the lake area. As such the response to the climate shift occurs over a shorter period than at Marcella Lake(~10,000 - ~9650 cal yrs BP) (Figure 8.2). The dates of the lake-level rises at both lakes suggest this is a synchronous event (Abbott *et al.*, 2000; Anderson *et al.*, 2001; Anderson *et al.*, 2002).

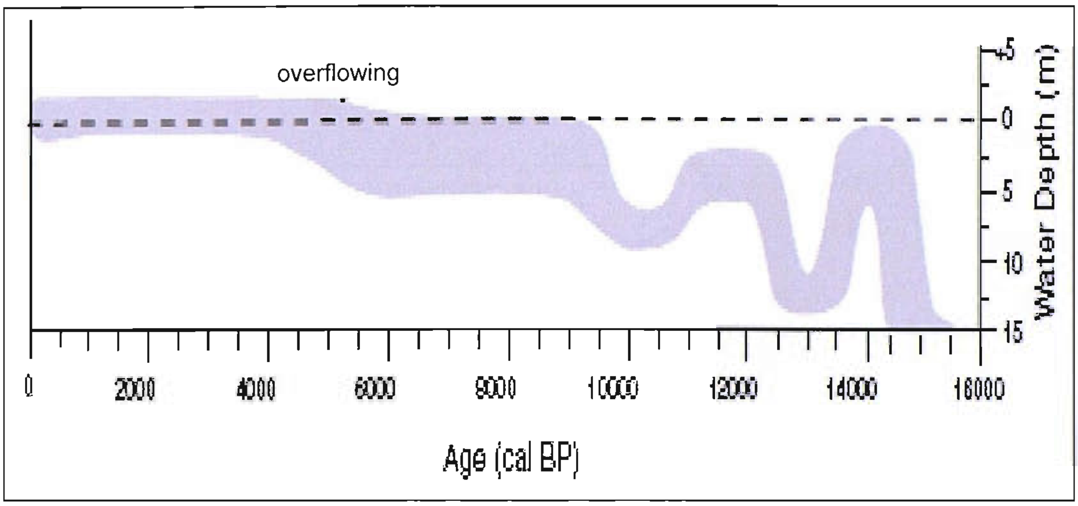


Figure 8.4: Birch Lake water levels (Abbott et al., 2000) on a calibrated radiocarbon timescale. Reproduced with permission from L. Anderson (Anderson et al., 2005b).

The charcoal records for Little Harding and Birch Lakes are shown in Figure 8.5 with the climate shift and vegetation transition marked for each site. The influx values for both sites are very low (and it should be noted again that the charcoal influx curve for Little Harding Lake is based on a second order polynomial as a temporary age-depth model, see Chapter 7).

In the Little Harding Lake record spruce first appears at ~10,100 cal yrs BP (Figure 6.3), and increases gradually over time, although the pollen abundances remain relatively low for the whole of the section analysed (<20%). The transition period retains a relatively stable fire episode frequency, increasing towards the end of the analytical window.

At Birch Lake the spruce rise is relatively rapid over a short vertical core distance, with pollen abundances increasing from ~5 – 30% at ~9300 cal yrs BP. However, the abundance is not sustained and very quickly decreases to ~20%. This effect is mirrored in other records from Birch Lake (Ager, 1975; Bigelow, 1997), and at other sites in the Interior and North of Alaska (for example: Brubaker *et al.*, 1983; Edwards *et al.*, 1985). In the charcoal record there is a step-like decrease in charcoal influx after ~9200 interpolated yrs BP resulting in a quiescent period in the record. The fire episode frequency displays an overall decreasing trend over the whole record.

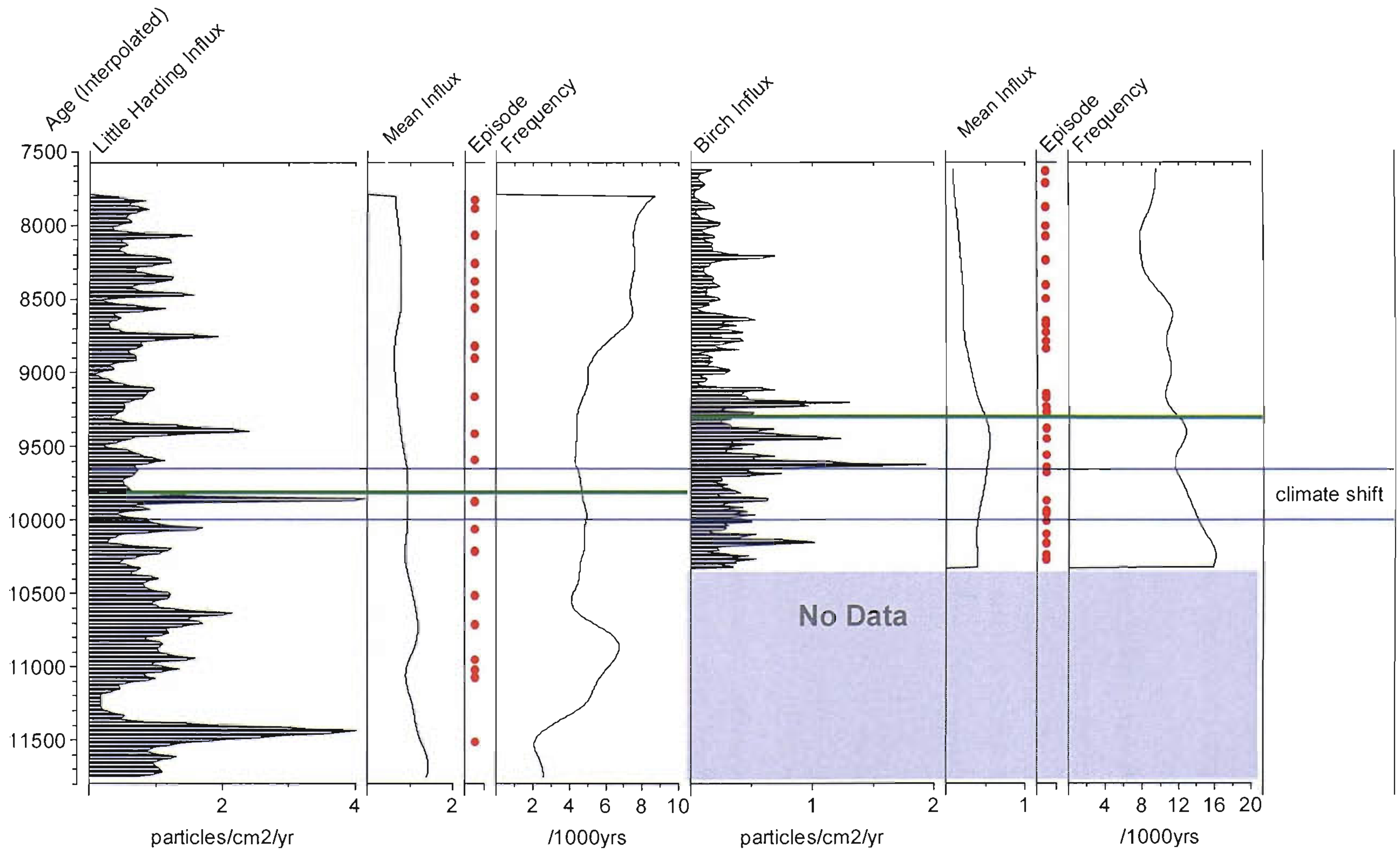


Figure 8.5: Charcoal influx data across the spruce rise for the Alaskan sites (Birch an (Little Harding Lakes). The blue band denotes the lake level rise (Figure 8.3). The green lines represent the rise of spruce in the pollen record above 15% (see Chapter 6).

It is probable that the relatively large basin sizes of these two lakes lead to the dilution of both the pollen and charcoal concentrations, resulting in the apparent low influxes. There is no significant increase in charcoal influx values across the spruce rise as can be seen in the Yukon Territory sites, and if anything Birch Lake displays a reduction in charcoal influx shortly after the spruce rise. There is no obvious alteration in peak frequency after the rise in spruce.

A Kolmogorov-Smirnov two-sample test was used to test whether the mean TBP distribution differed between the individual vegetation zones. The deciduous and spruce zones were not significantly different from each other ($p>0.05$) at Little Harding Lake, using the temporary age-depth model. At Birch Lake there was no significant difference between the two zones, but there is nevertheless a slightly higher episode peak frequency in the deciduous zone.

8.2.3 Role of Fire in the Early Holocene

It appears that regionally fire was a less frequent and/or less severe event during the deciduous zone (pre ~10,000 cal yrs BP) in the Yukon Territory. During this period a birch scrubland community was present on the landscape. Although it was not confirmed at these sites, it is generally accepted that shrub birch (*Betula glandulosa* or *nana*) as opposed to tree birch (*Betula papyrifera*,) was predominant at this time (Edwards *et al.*, 1991). Despite the low charcoal influx values, complete absence of fires is unlikely, as significant raw counts of charcoal particles were present, and other studies have also reported low influx values in this vegetation type (Brunelle-Daines, 2002). It is probable that small, low intensity infrequent fires characterised this period.

In Interior Alaska, fires appear to have been a more significant phenomenon than in the Yukon Territory. Fire episode frequency is relatively high in the deciduous zone, possibly indicating the influence of relatively high *Populus* abundances towards the end of the zone. Regionally the *Populus* peak is well recognised (Anderson *et al.*, 1988; Bartlein *et al.*, 1995; Mann *et al.*, 2002; Edwards *et al.*, 2005), and *Populus* forest/woodlands and shrub *Betula-Salix* tundra were common, although the interpretation of the forest extent is varied, ranging from relatively restricted populations along south-facing slopes or floodplains to extensive dense forests covering lower elevations(Ager, 1983; Brubaker *et al.*, 1983; Hu *et al.*, 1993; Anderson *et al.*, 1994). The increase in *Populus* is considered to be associated with a period of increased temperature, and is coincident with range expansions and increasing abundance of various other thermophillic minor taxa in northern Alaska (Edwards *et al.*, 1985; Anderson *et al.*, 1988; Lamb & Edwards, 1988; Edwards & Barker, 1994). The increase in *Populus* abundance would increase the available biomass contributing to the charcoal record in comparison with the Yukon sites. In addition, although direct evidence from these cores is not present, macrofossils of tree birch have been found at this time near Fairbanks (Hopkins *et al.*, 1981) and at nearby Wein Lake (Hu *et al.*, 1993). If tree-birch were present, it would further increase the available burnable biomass on the landscape, and this might contribute to the relatively stable charcoal influx over the transition compared with the Yukon sites.

Warmer/drier than present conditions have been inferred for the pre~10,000 cal yrs BP period from Birch (Abbott *et al.*, 2000), Marcella (Anderson, 2004; Anderson *et al.*, 2005a; Anderson *et al.*, 2005b) and Farewell Lakes (Hu *et al.*, 1998). In the contemporary boreal forest of Canada/Alaska warm/dry periods have been correlated with periods of increased fire occurrence and area burned due to the drying of fuels and increased ignitions and fire spread (Flannigan & Van Wagner, 1991; Amiro *et al.*, 2003; Flannigan *et al.*, 2003). Therefore, past conditions that were warmer/drier than present would be expected to favour a higher fire episode frequency. Some researchers hypothesise that the deciduous zone burned frequently enough to prevent the establishment of coniferous species on the landscape (Chapin *et al.*, 2004a). The deciduous species present are early successional species which can re-sprout from roots or below-ground stems to regenerate quickly and predictably after a disturbance event, reinforcing repeated cycles of the same successional sequences which can aid in the resilience of vegetation composition. Higher early Holocene fire frequencies have been reported in eastern Canada (Carcaillet & Richard, 2000; Carcaillet *et al.*, 2001a) but are associated with spruce-pine vegetation. Lower fire frequencies are reported when hardwood-hemlock species establish on the landscape. There are no other data available for fire frequencies in the early Holocene deciduous zone in the study region. One study has produce charcoal data for a site in Alaska spanning the last ~9.5 ka, which display very low CHAR from ~9.5-6 ka BP, but analyses of the fire regime were restricted to a mid-Holocene sequence (Lynch *et al.*, 2002).

The lower fire frequencies associated with a warmer/drier climate in this region more likely indicate that fire was not the excluding factor for spruce on the landscape. As the climate was significantly drier than present (25-45% less precipitation than modern) (Barber & Finney, 2000), it seems more likely that coniferous species were not able to establish due to habitat unsuitability (Mann *et al.*, 2002), and therefore deciduous taxa dominated. In addition to the relatively low flammability of deciduous scrubland in comparison with forested landscapes, the lower available biomass to produce charcoal due to low primary productivity in a moisture-limited environment, and the probability that the fires that did occur were of low intensity (Higuera *et al.*, 2005), it seems reasonable that the observed early Holocene fire episode frequency was lower than that estimated to occur on the contemporary landscape.

The cause of the rapid increase in effective moisture preceding the spruce rise is unknown at present. It may have been partly a result of postglacial sea-level rise flooding the Bering land bridge (Mann *et al.*, 2002). Alternatively the collapse of the Laurentide ice sheet has been interpreted as a cause of step-like climatic changes (Shuman *et al.*, 2002). The regional synoptic pattern that would deliver this shift in climate regime would require ridging over Alaska in summer, with a trough further westward (Edwards *et al.*, 2001). A stronger than present Pacific subtropical high is simulated, a slight northward migration of which would bring moisture into Interior Alaska (Edwards *et al.*, 2001). This configuration of systems would provide the conditions for storms to enter Alaska from the East Siberian Sea.

White spruce, the first spruce species to expand in the region after the climate shift, is rapidly growing and currently moisture limited in the Alaskan interior (Barber *et al.*, 2000). It is most abundant on better-drained and warmer soils (Yarie & Van Cleve, 1983). The fact that the vegetation shift occurs after the increase in effective moisture, the widespread onset of peat deposition (Mann *et al.*, 2002), and minor re-advances in the Brooks Range glaciers (Hamilton, 1986), appears to support the hypothesis that spruce was excluded from the landscape due to an unsuitable moisture regime. Further, it seems unlikely that a rapid fire regime leading to deciduous taxa dominance prevented spruce establishment as in this, and other records, the fire frequency is lower in the deciduous zone than the spruce zone (see also Lynch *et al.*, 2004).

After the transition to spruce-dominated forest the Yukon sites display an increase in the fire episode peak frequency. The increase in peak frequency is probably a result of increased available biomass (due to conifer presence, and increased primary productivity) and accumulating fuel load, and the increased flammability potential of the catchment from the increasing presence of coniferous trees.

It might also be likely that the number of ignition events increases over this period, and even on the contemporary landscape, lightning accounts for over 90% of the area burned in Alaska (Chapin *et al.*, 2003). Lightning occurrence and location is controlled by both synoptic processes related to El Nino, and by factors such as topography and presence of forest vegetation. Therefore, the introduction of storms into the region, and the expansion of coniferous forest would lead to a cycle of increased probability of ignition, and spread of fires in the landscape. It may be the case that the climate change itself may have added to the shift in the fire regime via increased ignition events.

All sites display a similar sequence of events whereby the climatic shift is followed by the transition to coniferous-dominated vegetation, indicating that climate is the primary driver of the vegetation shift during this period. The fire regime records a non-uniform response to the vegetation change. The Alaskan sites display a fundamentally unchanged fire episode peak frequency after the climate shift, whereas the Yukon sites display increased peak frequency. This can be more readily explained by the more established deciduous forest at the Alaskan sites, and possibly the reduced importance of spruce in the early spruce period, compared with the Yukon. It should also be borne in mind that the basin characteristics of the Alaskan lakes are quite different from the Yukon sites, so the comparison of charcoal records is subject to some uncertainty inherent to these differences. It is clear that whether the response is dramatic, or not, the alteration in the fire regime is subsequent to the vegetation transition, indicating that vegetation has more of an influence on the fire regime than does climate.

8.3 Spruce to Pine Transition

The timing of the arrival of pine at a site was traditionally assigned to the point in the pollen abundance curve where pine exceeded a threshold limit (e.g. 15% MacDonald & Cwynar, 1985; MacDonald & Ritchie, 1986; Cwynar & MacDonald, 1987). However, as discussed in Chapter 2, it is possible that pollen is not the best indicator of local pine populations within a lake catchment due abundant pollen production, especially, when the presence of pine macrofossils (stomata) indicate local presence when pollen abundances are ~1% or less. Pine is regionally absent from Alaska, and therefore the discussion presented here focuses on the Yukon sites only.

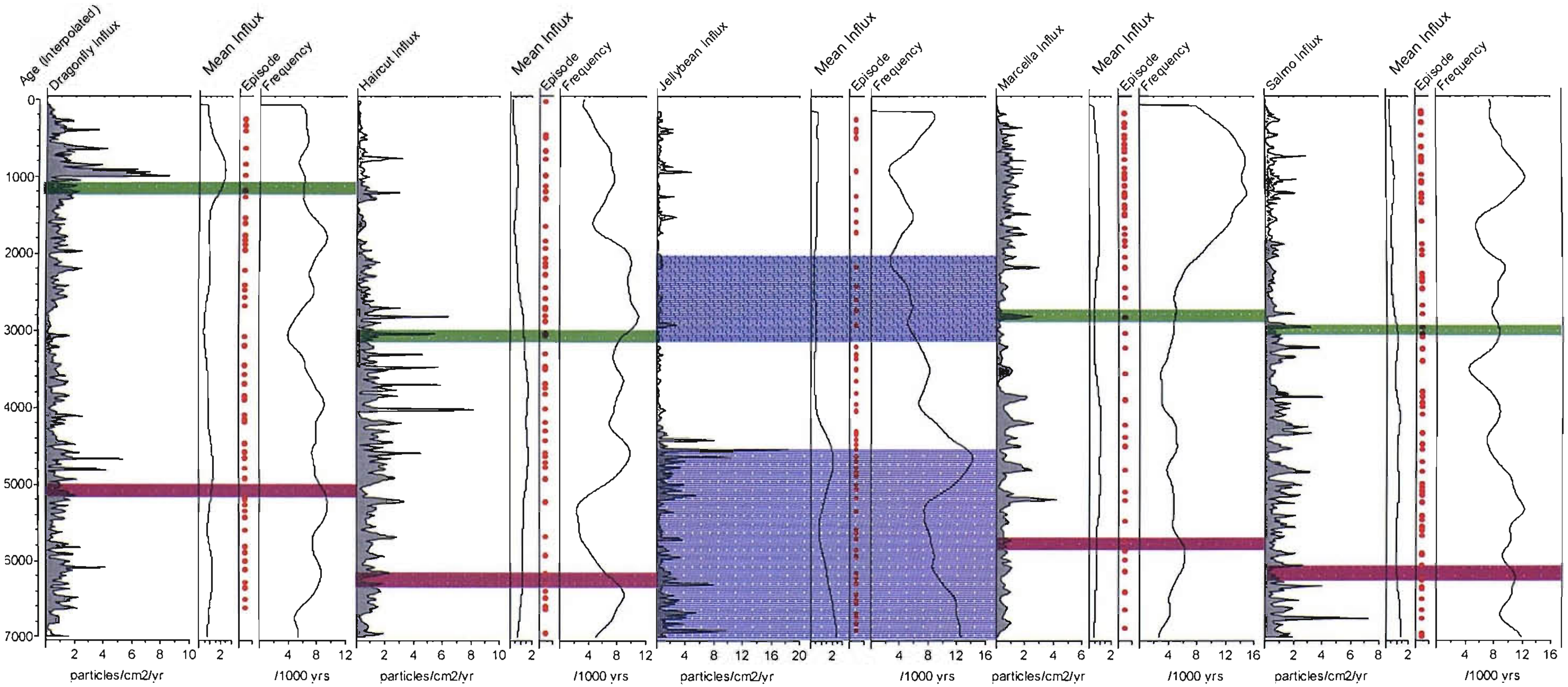


Figure 8.6: Regional comparison (Dragonfly, Haircut, Jellybean, Marcella and Salmo Lakes) of fire regime with vegetation transition from spruce-to spruce/pine or pine-dominated forest. The first appearance of pine stomata in the pollen records is represented by the purple line, the increase in pine pollen abundances over 15% by the green line. Oxygen-isotope inferred moisture increases are for Jellybean Lake (Anderson et al., 2005a)are represented by the blue boxes. ***Note:*** *Jellybean lake at present does not have a pollen record, hence the absence of stomata and pine rise lines.*

8.3.1 Yukon Lakes

The transition from spruce- to spruce/pine-dominated vegetation is not associated with a dramatic climatic shift compared with the spruce rise. However, three of the lakes display near-synchronous appearances of pine stomata and the later increases in pine pollen abundances, which are discussed further below. A key finding is that the peak frequency data for the spruce/stomata/pine zones are not significantly different from each other ($p>0.05$) at any of the sites, using a Kolmogorov-Smirnov two-sample-test. This suggests that introducing pine into an already conifer-dominated landscape does not significantly alter the fire regime. Comparisons of the sites that contain pine within the lake catchment today (Marcella and Haircut Lakes), to those where pine is regionally present, but is absent directly in the catchment (Salmo, Dragonfly and Jellybean), illustrate that there is little difference in the fire regime. Marcella Lake does display an increased charcoal peak frequency after the main pine expansion, although this is probably linked to more site-specific variables such as topography (Chapter 4), and moisture regime, as the most pine-dominated site (Haircut Lake), displayed no observable alteration in charcoal peak frequency.

It is useful to make inter-site comparisons in order to assess if the synchronicity of pine appearance is climatically forced. A comparison of the charcoal records for the five Yukon Lakes across the transition from spruce- to spruce/pine-dominated vegetation is presented in Figure 8.6, in association with the oxygen-isotope inferred moisture regime for Jellybean Lake. The oxygen isotope record reflects the composition of mean annual precipitation in the Interior Yukon Territory, which corresponds to changes in the North Pacific Index (NPI) (Anderson *et al.*, 2005a). The NPI is a measure of the intensity and position of the Aleutian Low (AL) pressure system (the semi-permanent low pressure located over the Gulf of Alaska). The results suggest a predominately weaker and/or westward AL between ~7500 and 4500 cal yrs BP and ~3000 to 2000 cal yrs BP. A more westward position and/or weaker pressure system has been associated with increased autumn and winter moisture delivery to the area. The record suggests that the AL shifted eastward and/or intensified between ~4500 and 3500 cal yrs BP and between ~1200 and 300 cal yrs BP. The eastward position or intensification of the system suggests a flow of moisture from the Gulf of Alaska directly into the St. Elias and Coast Mountain ranges, which in turn creates increased rainout on the coastal side and decreased precipitation in the interior Yukon Territory (Anderson *et al.*, 2005a).

Pine Tail/Stomata Zone

At Dragonfly Lake the appearance of stomata occurs during the period of highest charcoal peak frequency. The record at Haircut Lake is highly variable, although across the stomata rise there is a temporary increase in charcoal peak frequency. There is a pattern of peak clusters in the Salmo record, and the stomata rise occurs within the highest peak frequency period. The stomata rise at Marcella Lake is within the time period of lowest fire episode peak frequency, characterised by long quiescent periods. It is not likely that the appearance of pine in low densities would affect the fire regime significantly, even with the coincident shift to increased black spruce on the landscape it seems improbable. The fact that in three of the sites the stomata appear during a phase of increased peak frequency suggests that it is possibly more likely that the fire patterns affected the pine invasion, by creating openings in the forest whereby pine seedlings could either establish, or build up a seed bank. This suggests that pine invades a region in an opportunistic fashion, establishing a holding position, and expanding subsequently when conditions are optimal.

The charcoal record in the Yukon sites displays relatively little alteration in fire regime and there was found to be no significant difference between the fire episode peak frequency distribution of the pine and stomata zones (ANOVA, $p>0.05$). Data were compared for each of the Lakes individually, with all zones compared against each other. Therefore although the stomata data may give more highly resolved information on the timing of the arrival of the first populations of pine reaching a lake catchment, the inclusion of pine in an already conifer-dominated environment has relatively little impact on the fire regime as recorded in the charcoal record.

It can be seen that at the sites with pollen records (Dragonfly, Haircut, Marcella and Salmo Lakes) the first appearance of pine stomata occurs during or after the 7500-4500 cal yrs BP period of increased effective moisture. At Jellybean Lake the charcoal data within the 7500-4500 cal yrs BP higher moisture period reflects the highest influx values within the record. There is a step-like decrease in moisture after ~4500 cal yrs BP, and subsequent decrease in charcoal peak frequency.

Main pine rise

It appears that the expansion of pine on the landscape was near-synchronous at all sites except Dragonfly (the northernmost site). This could suggest that climate is exerting some direct influence over the vegetation transition. The main rise in pine pollen abundances occurs at all sites (except for Dragonfly) within the second period of increased moisture levels between ~3000 and 2000 cal yrs BP. At Haircut Lake, the vegetation shift is almost synchronous with the moisture shift, although due to the inherent errors associated with the radiocarbon determinations mean it is not possible to resolve the exact timing of the two events further at present It suggests that the pine expansion was a regional event, probably responding to more favourable climatic and ecological conditions in an opportunistic fashion.

The charcoal record, in contrast, does not display such uniform characteristics. At Dragonfly Lake the increase in pine pollen is associated with an abrupt increase in charcoal influx values and the WRA deposit. The fire episode peak frequency is relatively unchanged over this transition. This insensitivity of the charcoal record to increased pine is observed at Salmo Lake also. At Haircut Lake there is a general decrease in influx values after the pine rise, and the fire episode peak frequency also decreases gradually over time. In contrast, Marcella Lake displays an increase in fire episode peak frequency after the increase in pine abundances.

If we take the view that the vegetation change to pine-dominance may affect the fire regime, it appears that the introduction of pine into an already conifer dominated landscape produces little effect on the charcoal production and distribution within a catchment. During the mid- to late-Holocene in which the pine transition occurred, the climate was relatively stable (Bartlein *et al.*, 1998), with some evidence instead for smaller scale climatic variability (Anderson *et al.*, 2005a). Therefore, there is no evidence to suggest that the climate is driving any change in the fire regime, but it may be that climate is varying coincident with the vegetation changes. Thus, any alteration in the fire regime would be more likely be attributed to changes in the vegetation composition. This would be attributed to the relative increase in flammability of the catchment as a result of the introduction of another coniferous species likely to produce more intense crown fires, and a possible increase in the area burned. There is no significant alteration of the frequency of charcoal peaks at most of the lakes between the pine/stomata zones, and therefore the vegetation shift appears to have little/no impact on the fire regime. Marcella Lake displays an increase in charcoal peak frequency after the main pine expansion, and this is possibly due to site-specific factors. Marcella Lake contains large areas of dry steppe communities, interspersed with patches of black spruce and pine in the catchment. It may have been the relative dryness of the site that contributed to an increase in charcoal peak frequency in the late Holocene.

Contemporary observations of the different vegetation types suggest specific fire frequencies are necessary to ensure continued dominance (Johnstone *et al.*, 2004; Johnstone, 2005). It is possible that the fire regime response to the altered vegetation composition is masked by the analytical noise in the data, through alterations in sedimentation rates, and the respective age-depth curves. This is unlikely, however, as all sites display relatively unchanged charcoal peak patterns since ~6000 cal yrs BP. It could be that an alteration in fuel type altered the base charcoal contribution to the lake system, and therefore small changes in peak frequency are not detected. It is more likely that as far as the charcoal record is concerned, the transition from spruce to spruce-pine does not constitute a significant catchment scale alteration in terms of fire, as it is not as dramatic an alteration in the sense of flammability and biomass as the deciduous-spruce transition. Each of the lake catchments today contain very different amounts of pine, without any clear trend or differences in fire histories. It appears that catchments almost completely dominated by pine have a similar late-Holocene fire history to catchments with little/no pine present.

8.3.3 Stomata vs. Pollen

The migration and range of lodgepole pine (LPP) following the onset of deglaciation of the North American Ice Sheets has been extensively studied using fossil pollen analyses (MacDonald & Cwynar, 1985; Cwynar & MacDonald, 1987; Delcourt & Delcourt, 1987). Invasion of an area by a species consists of an initial arrival, followed by establishment on the landscape and subsequent population increase to the ecologically and environmentally determined carrying capacity (McLeod and MacDonald, 1997). The timing of this invasion and establishment is of particular importance in the transition from spruce- to spruce/pine-dominated forest due to the increased flammability potential associated with pine *versus* spruce forests.

Palynology is a well established and standard technique used to reconstruct palaeoenvironments (Birks & Birks, 1980; Faegri & Iversen, 1989). But whilst inference from fossil pollen can distinguish large-scale vegetation zones and assemblages, precise vegetation boundaries and local presence of species are often difficult to confirm from the pollen record alone (Clayden et al, 1997, Pisaric et al, 2001). Therefore, species existing in low densities over extensive areas in the past are not accurately mappable using fossil pollen data (McLachlan & Clark, 2004), and thus, low initial pollen occurrence and detection of species arrival requires the application of additional techniques (Sweeney, 2004).

Stomata are generally thought not to be dispersed far from source (Clayden *et al.*, 1996) and their presence in sediments is thus an indicator of local presence (Hansen, 1995; Carlson & Finney, 2004; Leitner & Gajewski, 2004; Sweeney, 2004). Pine stomata presence/absence has been used in this study to pinpoint more accurately the arrival of pine in the local area. It has been shown that stomata are limited to a source area of approximately 0.1ha (Parshall, 1999), which is a significant improvement for local reconstructions when compared to the 1-3ha representation of pollen (Sugita, 1994;

Parshall, 1999). This technique has the potential to become a standard palaeo-proxy, and is increasingly used to supplement palynology to improve resolution and determination of vegetation boundaries (Gervais *et al.*, 2002), and pinpoint the local arrival of species (Hansen, 1995; Clayden *et al.*, 1996; Hansen *et al.*, 1996; Clayden *et al.*, 1997; Carlson & Finney, 2004; Froyd, 2005). At the very least it has been recommended as an adjunct to pollen studies in areas where conifers dominate, or are an important component of the vegetation and vegetation history (Hansen, 1995).

Lodgepole pine is adapted to both rapid migration and explosive population increases (Critchfield, 1985), and was ranked by Rowe and Scotter (1973) as the most adapted for successional regeneration after fire of all the northern conifers by virtue of its seed retention on the tree, earliness of seed production, and seedling frost hardiness and growth rate. Several taxonomically recognised geographic races of LPP apparently survived the last glaciations without drastic modification (Critchfield, 1985), the northern races either persisting in far-northern refugia (Hulten, 1937) or migrating from the south. Near the end of the Pleistocene LPP is known to have occurred near the ice front in western Washington (Critchfield, 1985) and probably in the Rocky Mountains, and macrofossil evidence places it as part of a coniferous forest in northwest Montana near the Canadian border by ~11,000 cal yrs BP (Carrara & Wilcox, 1984). Populations in the Yukon were considered sufficiently different from other northern populations in terms of the number of needles per fascicle, and the timing of terminal bud formation for example, to instigate a reinvestigation of Hulten's (1937) hypothesis of a Yukon glacial refugium. In 1982, Wheeler and Guries published genetic evidence suggesting that northern interior Yukon populations originated from a refugium in the unglaciated part of west-central Yukon. However, it can be argued that pine migrating the relatively short distance from a refugium in the Yukon would have spread throughout the region early, and this is not reflected in the pollen record. Palaeobotanists have all but abandoned the idea that lodgepole pine had a Yukon refugium due to a lack of fossil pollen and macrofossil evidence (Hopkins, 1979), and argue instead that it reoccupied its northern range from refugia between the ice sheets that covered most of western Canada in the late Pleistocene (Anderson, 1970), or from south of the ice (Cwynar & MacDonald, 1987; MacDonald & Cwynar, 1991).

It is probable that the appearance of pine stomata at the very start of the pine "tails" in the records of this study represent small isolated populations establishing on the landscape ahead of the main migrating population (Johnstone & Chapin, 2003). The pollen abundances in the pine "tails" remain below 5% until the main pine rise at each lake. The abundances of pine pollen after the main population expansion exceeds 50% rapidly at every location and are maintained until the present, regardless of the abundance of pine immediately within the catchment today. The abundances are therefore probably a reflection of a swamping effect from a regional signal due to the overproduction of pine pollen rather than reflecting the proportion of pine within the catchment itself.

Regionally, then, it can be shown that pine populations established at ~6000 cal yrs BP, and subsequently expanded rapidly at Salmo, Marcella and Haircut Lakes south of Whitehorse at ~3000 cal yrs BP over *ca* a 100 year period, taking into consideration the errors associated with the radiocarbon determinations. At Dragonfly Lake, the northernmost site, the arrival is timed at ~5000 cal yrs BP, with the expansion at ~1100 cal yrs BP, coincident with the fall of the WRA. The Holocene migration pattern of LPP established by Cwynar and MacDonald (1987; see also MacDonald & Cwynar, 1991) suggests that LPP established as small, peripheral populations in excess of 70km from the mature main body of the population, via long distance dispersal events. From this initial establishment they propose that it takes ~1000-4000 years to achieve full dominance in an area, causing the pine "tail" during the delay prior to the main expansion observable in the pollen record. The evidence presented in this study regarding the presence of stomata in the pine "tail" suggests that this theory is correct, however, pine migration rates have probably been underestimated if they are based solely on pollen records. If we assume that when analysed for stomata, the locations cited in MacDonald and Cwynar (1991) display the same trend in stomata appearance at the start of the pine "tail" as the sites in this study, the initial appearance of pine on the landscape is much earlier than first thought, and average migration rates increase from ~130 m/yr^{-1} to over 160 m/yr^{-1} (more in line with Huntley's (1991) estimation of migration rates).

It has been thought that if a species migration occurs as a diffuse spread of low density populations, the advancing species range limit may not be possible to detect palynologically (Giesecke, 2005), and the abrupt increase in pollen values may be representative of the expansion of scattered, small populations (Bennett, 1988; MacDonald & Cwynar, 1991).

The relationship between the pollen composition in lakes and bogs and the surrounding vegetation led to the development of models to describe the pattern of dispersal and deposition of pollen (e.g. Prentice, 1985; Sugita, 1994; Giesecke, 2005) to simulate pollen spectra from hypothetical landscapes in order to explore the implication of possible scenarios. Simulations of a moving front-type of species migration show that the pollen detected at a site should increase sigmoidally (Giesecke, 2005), with steeply increasing pollen abundances (such as those observed in the pine curves of Chapter 6) remaining until the front has passed the sampling site. Here the author proposes an explanation for the method of lodgepole pine migration into the study area.

Lodgepole pine has been observed to establish satellite populations up to 70km from the main population via its abundant far-spreading, wind dispersed seed (MacDonald & Cwynar, 1991). Therefore it is probable that the stomatal evidence of pine trees at the lake sites at ~6 ka BP, far in advance of the timing of the main pollen rise (~3 ka BP) is a result of the satellite population migration regime. The continued low levels of pine pollen through the pine "tail" are possibly explained via marginal or sub-optimal habitat conditions persisting in the region. The vegetation composition displays evidence of gradual alteration in general through the period of pine migration in response to climatic variability for example. Analyses of spruce pollen indicate that black spruce became increasingly more abundant from ~6ka (Cwynar & Spear, 1995; Gallagher, 2006), which, when taken into consideration with the coincident expansion of alder (Cwynar & Spear, 1995), have been interpreted to imply cooler and wetter growing seasons, coinciding with the first period of increased precipitation inferred from the Jellybean Lake record (Anderson *et al.*, In Press). Peatland analyses in the region have also produced evidence of an increase in soil moisture at this time (Wang & Geurts, 1991). Lodgepole pine generally occurs as a minor seral species in areas with cold/wet soils, and obtains dominance in areas of well-drained and warmer soils. Therefore, it is possible that lodgepole pine initially appeared in the region on sites that had become too cool and/or wet for white spruce to retain dominance, coincident with the black spruce expansion. Once present in these locations, lodgepole pine would remain a minor component of the vegetation due to being out-competed by black spruce which is much more dominant on wetter or nutrient poor substrates, probably restricted to more marginal sites. Lodgepole pine may then spread into the traditionally white spruce dominated areas as and when disturbance events (e.g. fires) occur, out-competing white spruce seedlings, due to more rapid germination and growth rates, and continuing to expand the local population.

8.4 Interactions among Fire, Climate and Vegetation – What has been learned?

The vegetation transition from deciduous- to coniferous-dominated forest at ~10,000 cal yrs BP displays a clear sequence in which the climate shift precedes the alteration in vegetation composition, after which alterations in the fire regime are observed (Figures 8.2 and 8.4).

The transition from spruce- to spruce/pine-dominated forest is time transgressive between sites. There is no clear response to the arrival of pine in the charcoal record in the Yukon (Figure 8.5), nor is there a response of the pine to patterns in the charcoal record, suggesting that there is no strong interaction between vegetation and fire during this transition. The pine transitions do occur during moist periods, and therefore it seems likely that climatic variability is influencing the vegetation shift over this transition to some extent.

In consideration of the hypotheses outlined at the start of the project, the weight of evidence suggests that climate is the overall primary control on vegetation composition, with the fire regime more directly controlled by the vegetation over the spruce rise. The fire regime appears to be unresponsive, or decoupled from shifts in conifer dominance (the pine rise). Therefore H_1 is rejected, and H_2 is accepted as the more likely explanation for the interaction of fire, vegetation and climate in the boreal forest of Alaska-Yukon during the Holocene.

8.5 Data Significance and contribution to the field.

This study provides a long-term perspective on the dynamics of fire and ecological invasion in the boreal forest of northwest North America. It adds weight to the assumption that a primary impact of predicted global climate change will be on the vegetation composition of the far north, supported by the regional scale agreement between these records of the vegetation response to past climatic alterations. The specific pathway of the vegetation change will depend on the exact alterations in climate. Increasing temperature only will likely lead to a decline of spruce (Barber *et al.*, 2000), whereas a coincident increase in moisture could see the northward migration of the tree-line, and a conversion of the southern boundary to include more deciduous species. This study has shown that vegetation is the primary driver of the fire regime over the long term, at the very least for the shift between deciduous trees and evergreen conifers. This finding is consistent with the findings of Clark *et al.* (1996), where coniferous and hardwood-dominated vegetation recorded very different fire frequencies in the early Holocene (Chapter 2). Alterations in vegetation composition caused by global warming will probably lead to alterations in the fire regime of the boreal forest, causing alteration of land-atmosphere energy feedbacks, and impacting on the global carbon budget. It has been estimated that the long-term average fire emission of C from the boreal forest is much larger than has been accounted for in global C cycle models (Harden *et al.*, 2000), and although these areas may act as slight net sinks for C from the atmosphere to land over decadal to century times, periods of drought and severe fire activity may result in a switch to net sources of C. This was almost certainly the situation over the spruce rise as suggested by the increase in mean influx of charcoal and peak frequency at the sites studied, and it is possible that this situation may occur again in the future.

In Alaska where the initial response to global warming is likely to be the more rapid spread of lodgepole pine north and west into the region (Johnstone & Chapin, 2003), the fire regime may remain relatively unchanged, as in the case of the Yukon lakes during the late-Holocene pine invasion, although the intensity of fires and possibly the area burned will probably increase judging by modern observations. In this scenario, it may be possible that the relative importance of climate as a regulating force on the determination of the regional fire regime may become more apparent, as in the studies of Carcaillet *et al (2001a; 2001b)*. Regionally there is likely to be a reversion to more deciduous forest types (aspen-parkland for instance, Hogg & Hurdle, 1995; Johnstone *et al.*, 2004; Legare *et al.*, 2005) particularly in the Yukon Territory, which is geographically closer to the southern border of the boreal forest biome, and there are already observations of this situation today. This will probably be the case, especially if warming is not offset by equivalent increases in moisture as expected (Maxwell, 1992), as the coniferous species are more susceptible to drought stress. This alteration in vegetation composition will likely have secondary impacts on albedo via decreased surface roughness in winter, on permafrost dynamics, due to increased ground insulation, and alterations in the fire regime, in addition to probable alterations of faunal composition.

If there is a coincident increase in moisture, GCMs results indicate a significant increase in penetrating convection, implying a change in the intensity and frequency of large thunderstorms (Overpeck *et al.*, 1990), which will again favour the increased abundance of early successional species if there is a significant increase in fire frequency. If fire frequency remains relatively stable, spruce could retain dominance on the landscape. It has been conjectured that the time-lag between vegetation responses to rapid climate change to warmer and drier, and the trend of development of more open/deciduous vegetation could delay the positive feedback of vegetation change to climatic warming (Chapin & Starfield, 1997), through controls over the fire regime.

In the short term local climate conditions are likely to drive the fire regime seasonally, and it is likely that this transient condition could feedback to further alter the long term patterns of climate and vegetation. In an average year in Alaska, the fire season lasts only three weeks at present. However, over 50% of the total area burned between 1961 and 2000 was burnt within just 6 years, coincident with fire seasons that lasted far longer than normal (Whelan, 2002; Chapin *et al.*, 2003). Unusually severe fire events/seasons, although infrequent, may trigger vegetation changes under certain conditions (i.e. when species are already existing at their geographic and/or physiographic range limits).via the removal of seed banks, extreme pre-fire weather patterns and alterations of successional pathways (Kittel *et al.*, 2000). It is possible that such events participated in the rapid and abrupt pine expansion in the Yukon Territory. The presence of unusually large magnitude peaks within the record, and periods of large peaks around the vegetation transitions may be part of the signal for such events within the sedimentary charcoal record.

This study can provide data to further inform ecological models simulating land-cover changes as a response to global warming. Comparison of the response of dynamic global vegetation models to transient climate (Cramer *et al.*, 2001) have shown that models were most consistent at high latitudes, possibly due to temperature being of vital importance in any ecosystem response (Kittel *et al.*, 2000). All models projected increases in net primary productivity and vegetation biomass, and included a northward shift of vegetation zones under future climate change scenarios. However, none of the models include a fire module that can realistically simulate the interactions between fire, climate change, and vegetation. Fire is just beginning to be included in such models (Bachelet *et al.*, 2005), and, this fire information is mostly based on modern fire theory and observations. An example being the SPITFIRE module (Thonicke & Cramer, 2006) which has been incorporated into the LPJ Dynamic Global Vegetation Model, which simulates fire patterns that correspond well with observed average fire conditions, although this leads to underestimation of extreme fire years, and an overestimation of low fire years. The simulations showed that the vegetation composition influenced fire spread conditions, and had little impact on fire ignition potentials, except when the vegetation consisted of only broad-leaved deciduous forests, when ignitions increased. Long-term fire records such as this one highlight the disparity between modern observations and the charcoal evidence in the sedimentary record for drastically altered climate regimes, thereby their inclusion for informing future models can go some way to reduce the uncertainty associated with the output. Using contemporary fire ecology theory, deciduous forest is considered to burn more frequently in this area than coniferous dominated landscapes. The data presented here show that under a drier and warmer climate regime than present, the deciduous forest may in fact burn less, coincident with other fire records in this region (Hu *et al.*, 1993; Hu *et al.*, 1996; Lynch *et al.*, 2002; Lynch *et al.*, 2004). Modellers need to incorporate realistic fire modules into their systems, which can be informed via studies like this, in order to produce effective simulations of future ecosystem dynamics.

The project can also contribute to the planning for fire and resource management in the future boreal forest. Over the long term vegetation is the primary driver of fire regime, therefore forest management has to incorporate either the contingency for altered fire dynamics, or to manage landscapes to ensure overall vegetation composition remains relatively stable. Ultimately the recent efforts of large-scale fire suppression have proved to be fruitless, economically unviable and ecologically unsustainable. Fire is an important ecological process, and resource management needs to adapt accordingly.

Chapter 9. Conclusion

9.1 *Summary of Findings*

The very nature of palaeoecological research results in a significant amount of uncertainty associated with the conception of a project. This leads to a reliance on a certain amount of favour from "Lady Luck" when proposing to address specific research questions, regardless of the amount of care undertaken by the researcher, both in the field and in the planning stages. In addition the results of palaeoecological studies, and scientific investigations as a whole, often give rise to new hypotheses and research questions not identified by the original project objectives, or not identifiable within the knowledge of current research. This project is no exception, and has produced new hypotheses to be answered in the future, but in regard to the initially proposed questions, it has achieved a relatively high amount of success. At the onset of the research, the main focus was on the proposal of clearly-defined hypotheses concerning the interaction of fire, vegetation and climate in Alaska-Yukon during the Holocene (see pages 4 and 209). Difficulties associated with obtaining long enough records, producing enough high resolution analyses to inform the study, and the ever present chronological constraints mean that the findings discussed in Chapter 8 are by no means definitive; they are though a clear and proper first step towards the resolution of a complex set of palaeoecological and intellectual challenges. This chapter provides a summary of the key research findings as related to the specific aims and objectives outlined in Chapter 1 and discussed in Chapter 8, and an appraisal of the methodologies used in this study especially in relation to possible directions for future work.

This study has shown that the vegetation transition from deciduous- to coniferous-dominated forest at ~10,000 cal yrs BP displays a clear sequence, where the climate shift precedes the alteration in vegetation composition, after which alterations in the fire regime are observed (Figures 8.2 and 8.4). This is the first time that the hierarchy of control over the fire regime has been reconstructed across a major climate and vegetation shift, and in this geographic area, and suggests that climate is the primary control on vegetation composition, to which the fire regime responds.

The timing of the presence of LPP within the catchment of a lake is likely underestimated by the pollen record, with stomatal evidence of local presence preceding the main pollen rise by ~3000 cal yrs BP. This initial appearance is a synchronous event in three out of the four sites studied. The transition from spruce- to spruce/pine-dominated forest is time transgressive between sites, but is generally associated with periods of increased moisture during the late Holocene. This evidence challenges the conventional view on the migration history of lodgepole pine through the Yukon Territory, and provides further evidence of a moving front style migration pattern.

There is relatively little response in the charcoal record to this alteration in vegetation composition (Figure 8.5), contrary to the modern theory of species flammability, and therefore it would seem probable that a transition between coniferous types is not identifiable in the charcoal record alone, or that the alterations are small in scale, and that the records are not of sufficient temporal resolution.

9.2 Methodological Appraisal and Recommendations for Future Research

The data produced for this study are in general of high enough resolution where necessary to answer the objectives outlined in Chapter 1. The multi-proxy approach used, and the comparison of the vegetation and palaeoclimate data produced vegetation histories with a high degree of similarity to other reconstructions within the study region. This allows for relative confidence in the interpretation of regional burning under the various vegetative regimes. Studies that have produced charcoal data in relatively near geographical range to the study site in this project are also available for general comparisons. However, there is a need to improve the chronologies of the majority of the sites used in this study, to provide more robust age-depth models used to inform the calculation of CHAR, and distinguish between the effects of site-specific alterations in sedimentation and regional burning characteristics.

There are some key limitations associated with the datasets, and they primarily concern 1) the chronological constraint, and the resulting difference in time-equivalent sampling resolution, 2) the low charcoal influx values, especially in the deciduous zones and 3) the uncertainty associated with uncalibrated fire records.

Chronological constraints

Chronological error is the primary limitation of all non-annually resolved palaeo-datasets. It can provide a distinct obstacle in the comparison of records, even those within this study, due to the differences in chronological resolution and the error associated with each radiocarbon determination. There is a distinct possibility that some of the seemingly synchronous events discussed are in fact significantly time-transgressive, but also that the more time-transgressive transitions are more closely matched than allowed for in the interpretation. The presence of the WRA tephra in the Yukon cores provided a pinning point to aid comparisons in the late Holocene part of the records. It would have been advantageous to conduct more detailed tephra analysis on the Alaskan lakes, and on the early-Holocene portion of the Yukon records. There was unfortunately not the time for this within the current study, however, it will be considered in the future work using the datasets produced here.

Low charcoal influx values

The low influx values and fire episode peaks during the deciduous zone at the Yukon and the generally low influx values at the Alaskan sites have been interpreted conservatively, in order not to assign trends in data associated with what effectively might be noise within the data being identified solely as part of the peak component. However, these values are consistent with qualitative (presence/absence data) sediment charcoal records at Dune (Lynch *et al.*, 2002), Farewell (Hu *et al.*, 1996) and Wien Lake (Hu *et al.*, 1993). It would appear that low influx values are not uncommon in charcoal records, and they usually are derived from high or adequate original data or raw counts (Millspaugh *et al.*, 2000; Whitlock, 2001). The Alaskan sites used in this study provided high charcoal raw counts, and it is the high sediment accumulation rates that result in low influx values. In areas of slower accumulation increased sampling resolution could allow for more rigorous analysis, and therefore confidence in the model output. The Yukon sites suffered from both the short amount of time between local deglaciation and the onset of the spruce rise and the low charcoal influx values. This disallowed a thorough interpretation of fire changes across the spruce rise. However, it should be noted that the original charcoal counts were especially low, indicating that there was generally very little burning in the landscape at this time. This is most likely the result of low fuel availability, low flammability status and/or lack of ignition sources, although the relative importance of each is not specifically interpretable from these sites. The issue over whether the fires of high frequency, patchiness and consisting of low biomass that they are not detectable in these records is more important to the overall question, and it would be advantageous to reconstruct the deciduous zones from sites within the Yukon Territory that were not glaciated (i.e. further north or west), and therefore able to provide longer deciduous zone records. The aspen parkland zone in western Canada is considered to be a dry-climate analogue for the future boreal forest (Hogg, 1994). In this ecosystem conifer species are largely absent, and aspen is restricted to patches of stunted trees interspersed with grassland. Further south the parkland grades into true grassland/prairie, where trees are absent (Hogg & Hurdle, 1995). These areas are well-suited to persist in drier climates, due to their clonal nature, and the ability to resprout vegetatively following fire or severe drought. In this environment conifers became gradually eliminated as they burn before reaching seed-bearing age, which is delayed due to drought stress, or the soil conditions become too dry for successful seedling establishment (Hogg, 1994). It would be possible to use such sites to construct charcoal records to further inform the interpretation of the pre-10ka deciduous zone.

Calibration

The calibration of surface samples with tree ring or fire scar data and known historical fires could provide the opportunity to calibrate the parameters used in CHAPS and therefore reduce some of the assumptions and error associated with the model interpretation. It would therefore be possible to identify a site-specific CHAR level that is known to provide an accurate reconstruction of the catchment fire regime, and also to provide information of the amount of time over which a single fire event provides charcoal to each basin. This could only be conducted at sites with consolidated surface sediments, and/or with long enough historical fire data to allow for accurate identification of charcoal peaks. The catchment at Dragonfly Lake is known to have burned almost completely in 1958, and with a ^{210}Pb chronology of recent sedimentation rates, it might be possible to identify the charcoal signal that this event produced. Due to the relatively young ages of the stands, and reasonably regular and frequent fire events the use of fire scars and dendrochronology would be restricted to the most recent time. Historical records would extend this record slightly further, and although the length of the calibration record would be restricted, it would still provide more site-specific insights into CHAR and aid in the interpretation of the longer record.

Arguably the resolution of the charcoal record is not as high as originally anticipated. High temporal resolution was recognised as being imperative during the planning stages of the project to accurately reconstruct CHAR, and provide the opportunity to analyse the timing of the vegetation and fire shifts in conjunction with the known climate history. Due to the geographic location of the cores *versus* that of the author for the majority of the project, the acquisition of samples became increasingly time-consuming It was necessary to decide on a sampling strategy relatively early on, with the view that samples could be amalgamated if necessary to achieve a time-equivalent sampling resolution. The author sampled at the highest resolution possible basing decisions on either a known low-resolution chronology (i.e. based on range-finder dates), or by being guided by the identification of vegetation zones prior to charcoal sampling. However, it became apparent as the chronologies became more established that in some cases the sampling intervals were too low to provide a time equivalent sampling resolution at the level ideally required for charcoal analysis (~20 years/cm^2 or less). The deposition times in all cores range between ~10 - ~50 years/cm^2 with the current chronologies. The charcoal records at Haircut, Salmo and Birch Lakes were sampled at 0.5cm resolution, and realistically this is the highest sampling resolution within current methodological limitations. Marcella, Little Harding, Dragonfly and Jellybean Lakes could potentially undergo higher resolution charcoal analysis in areas of increased sedimentation in order to further resolve the records as they were sampled at 1cm intervals. The NERC radiocarbon steering committee application procedure meant it was not possible to secure dates prior to considerable data collection, and even once dates were awarded there was significant delay in the processing of samples within the NERC laboratory (the author is still waiting on some submitted in April 2006). The strategy of the committee is understandable considering the cost of a radiocarbon determination, and the numbers of requests it receives; however, its current operating procedure inhibits the ability of the investigator to provide an optimal dataset.

To address these issues in future the author recognises the need to obtain more sediment in the first instance, either by taking multiple cores, or cores of larger diameter to ensure that there is; a) the ability to separate large samples immediately for radiocarbon determinations, b) enough material to yield sufficient carbon to produce a radiocarbon determination, and c) opportunity for the sampling resolution for charcoal analysis to be further adjusted by re-sampling at higher resolution when the chronology dictates it necessary. Ultimately a reasonably secure chronology is essential for charcoal analysis and ideally this should be established prior to any analyses being undertaken.

9.3 Future Research

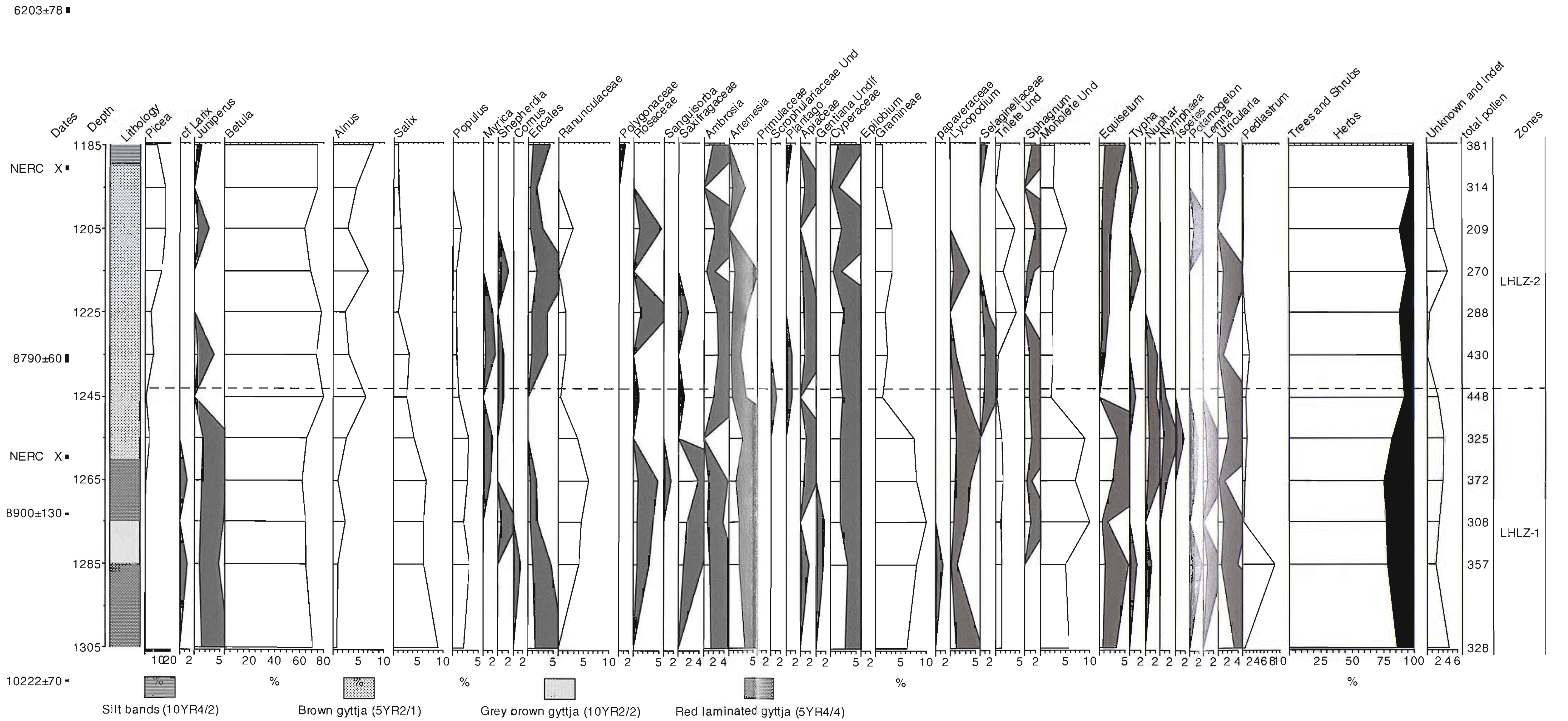
There are naturally further analyses and interpretations that can be made on the dataset, and these will be conducted during the compilation of journal articles for publication. The primary aim of this investigation was to address the overall hierarchy of controls on the fire regime in terms of climate and vegetation. Within the time constraints of the project, these issues have been investigated and conclusions drawn; however, the author would like to highlight some analyses that will further inform the interpretation and final publication of the results:

1. High-resolution pollen and possible increased charcoal resolution across the transitions, to further inform the interpretation of these unstable periods within the record, especially across the deciduous- to spruce-transition in the Yukon sites.
2. The interpretation of the isotopic data available for Dragonfly Lake to calibrate the lake level interpretation used from Marcella Lake
3. The completion of pollen and stomata analysis at Jellybean Lake
4. The inclusion of model analyses using the Poll-and-Cal system (Sugita, 1994) to estimate the respective impacts of the LPP migration and spruce establishment within the catchments on the charcoal record.

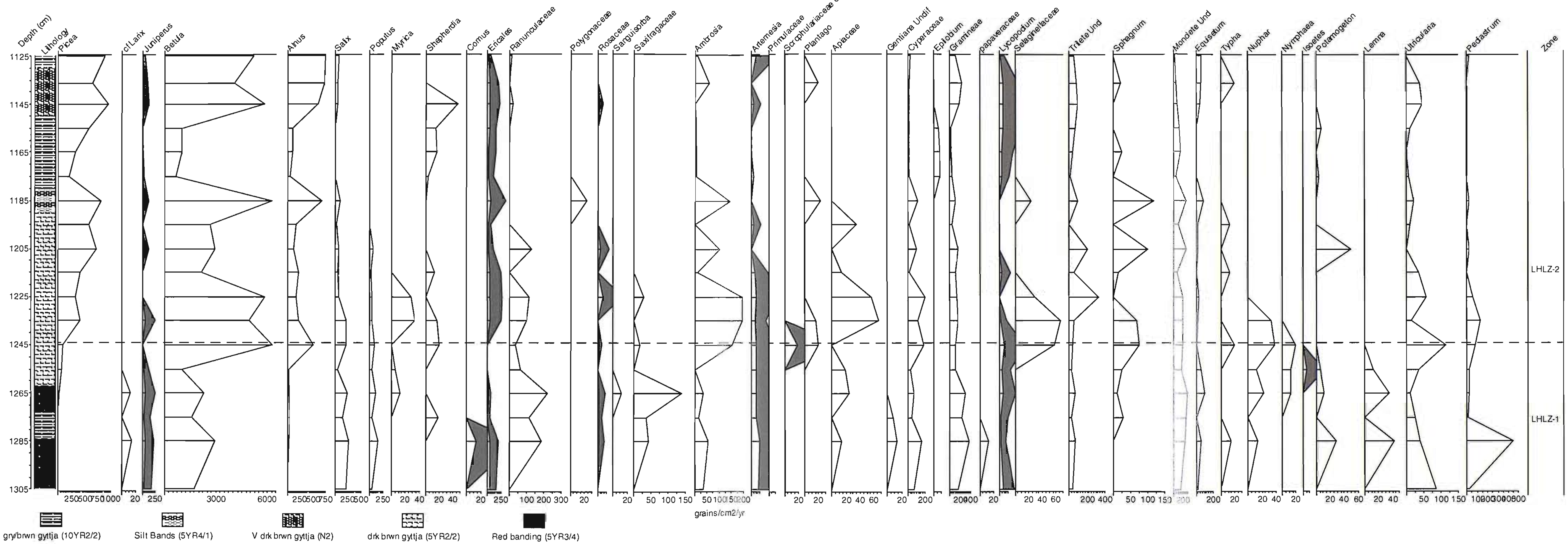
Appendices

Appendix 1: Pollen Diagrams

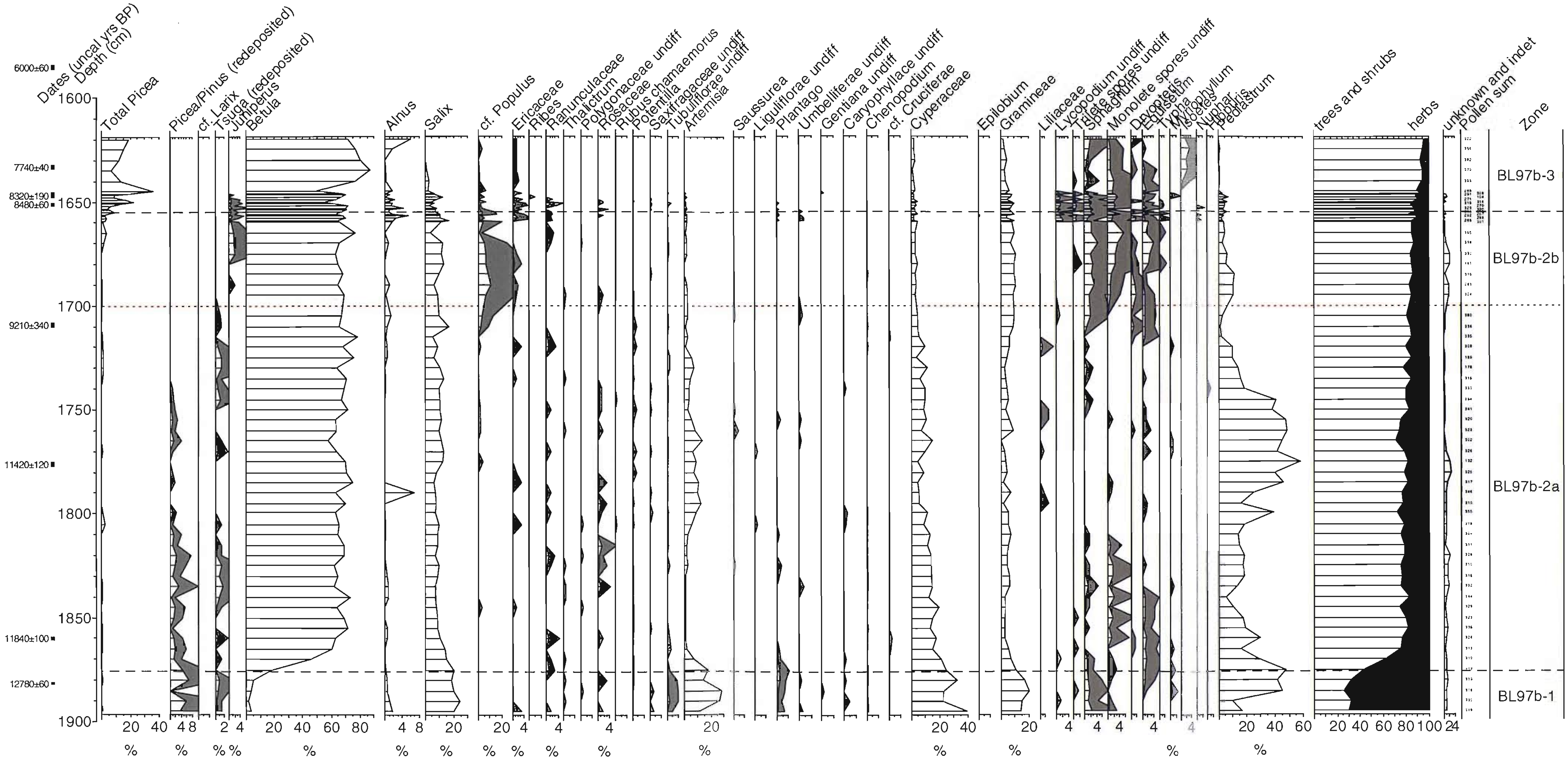
Little Harding Lake, Pollen percentage Diagram for Analytical Window
Full Taxa. Grey shading represents a 5x exaggeration factor applied
Analyst: LMFS



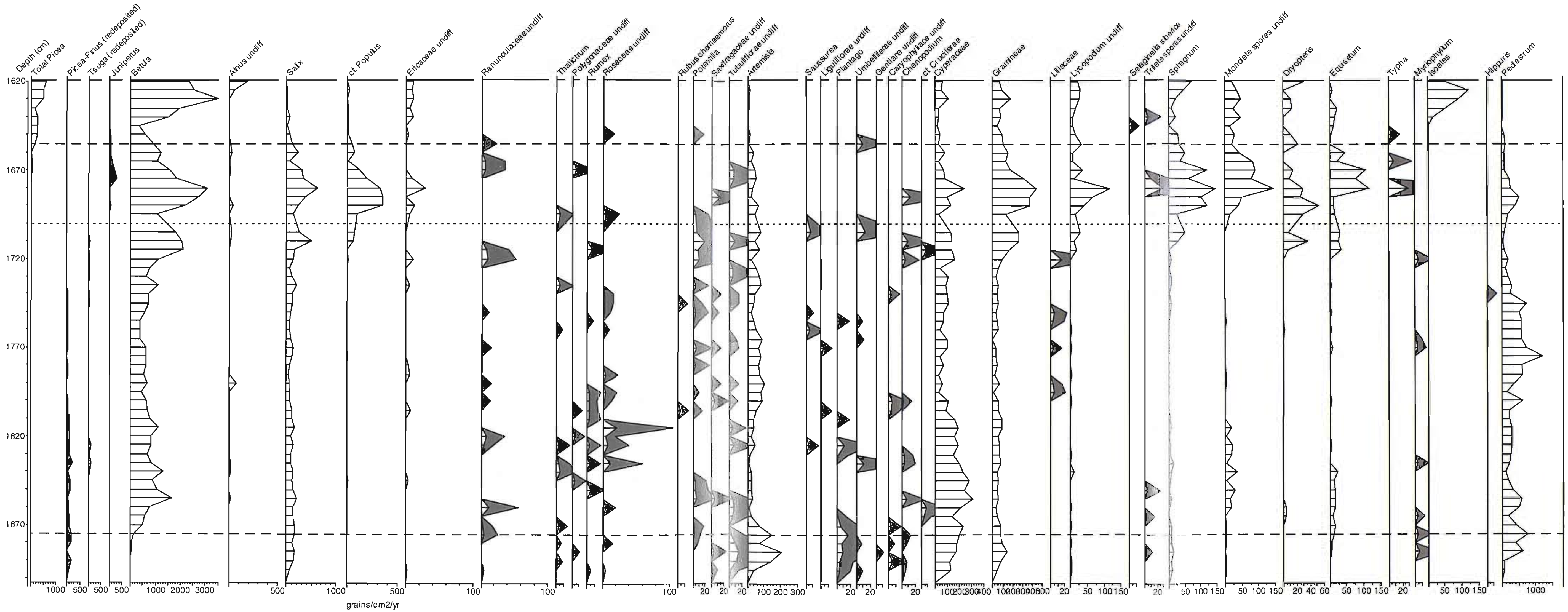
Little Harding Lake, Polle Influx diagram for the Analytical Window
Full taxa. Grey Shading indicates a 5x esaggeration foactor applied
Analyst: LMFS



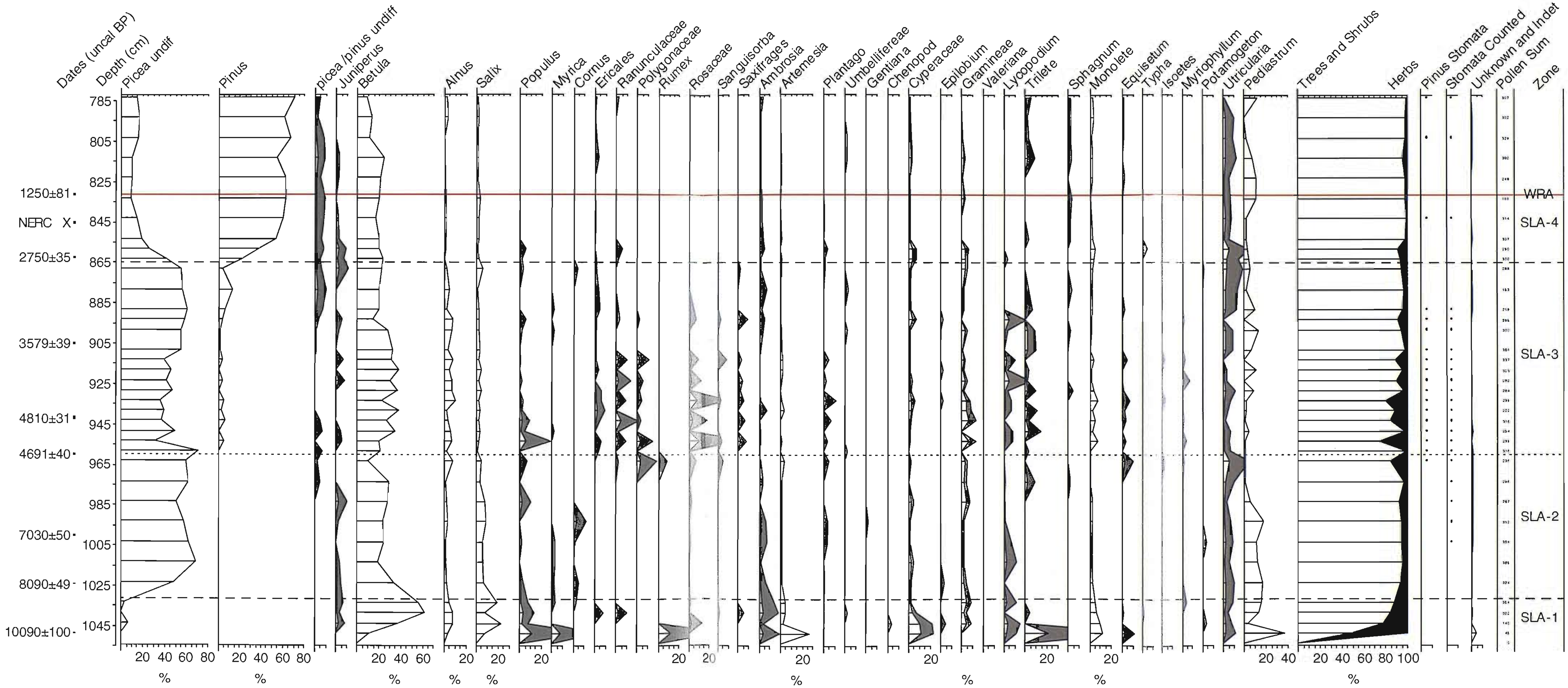
Birch Lake, Pollen Percentage Diagram
Full Taxa. Grey shading represents a 5x exaggeration factor applied
Analyst: LMFS and NHB



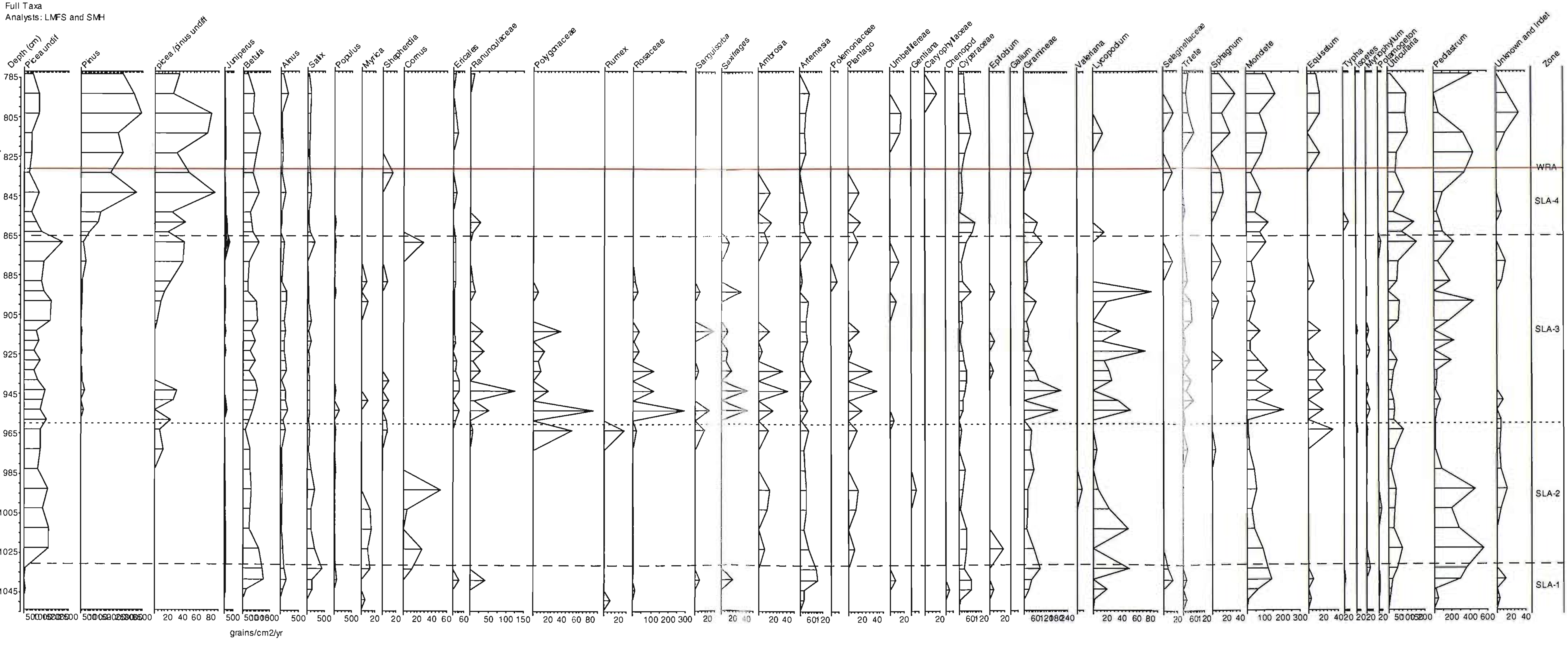
Birch Lake Pollen Influx Diagram
Full Taxa. Grey shading represents a 5x exaggeration factor applied
Analysts: LMFS and NHB



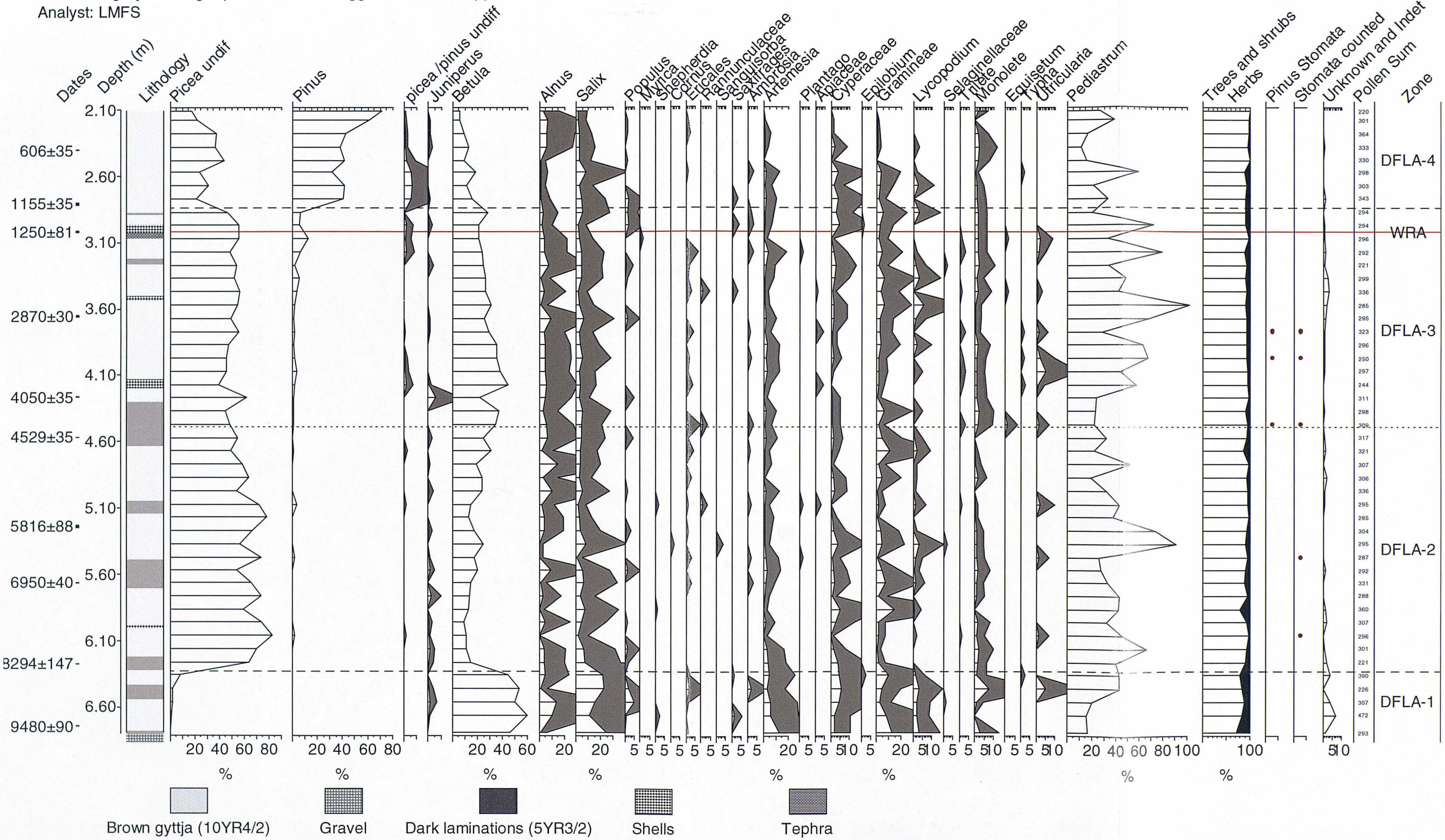
Salmo Lake Pollen Percentage Diagram
Full Taxa. Grey Shading represents a 5x exaggeration factor applied
Analysts: LMFS and SMH



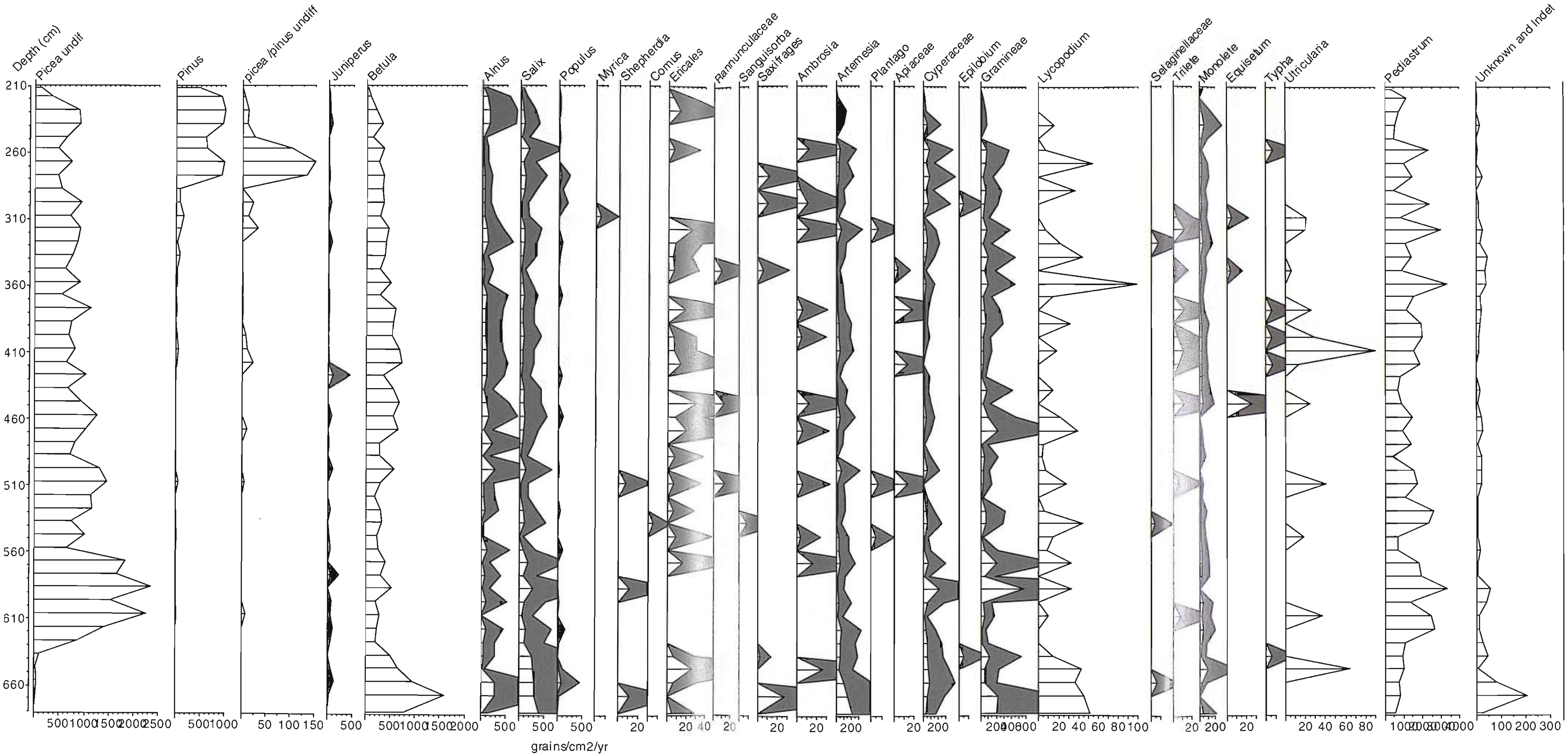
Salmo Lake Pollen Influx Diagram
Full Taxa
Analysts: LMFS and SMH



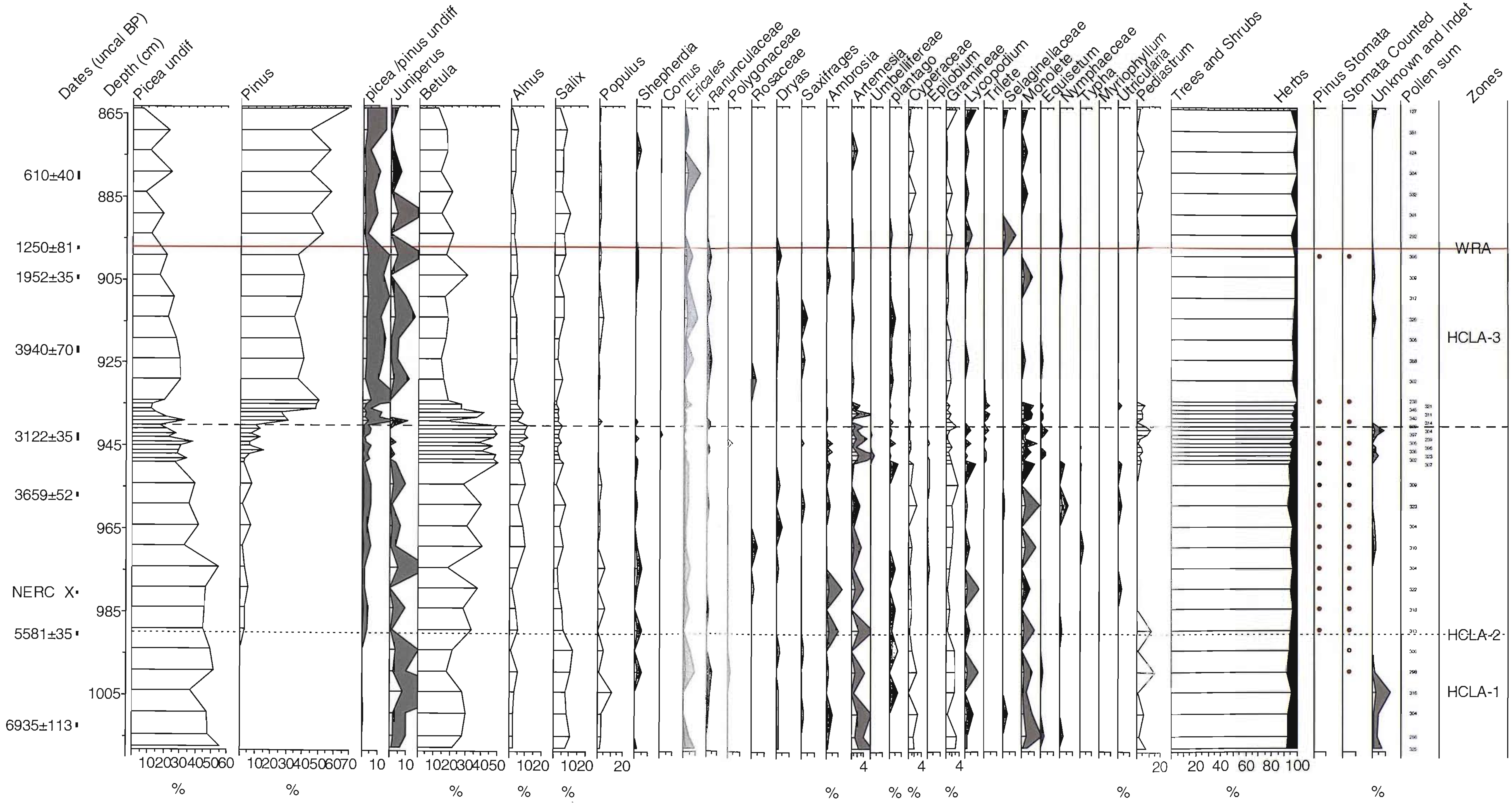
Dragonfly Lake Pollen Percentage Diagram
Full Taxa, grey shading represents a 5x exaggeration factor applied
Analyst: LMFS



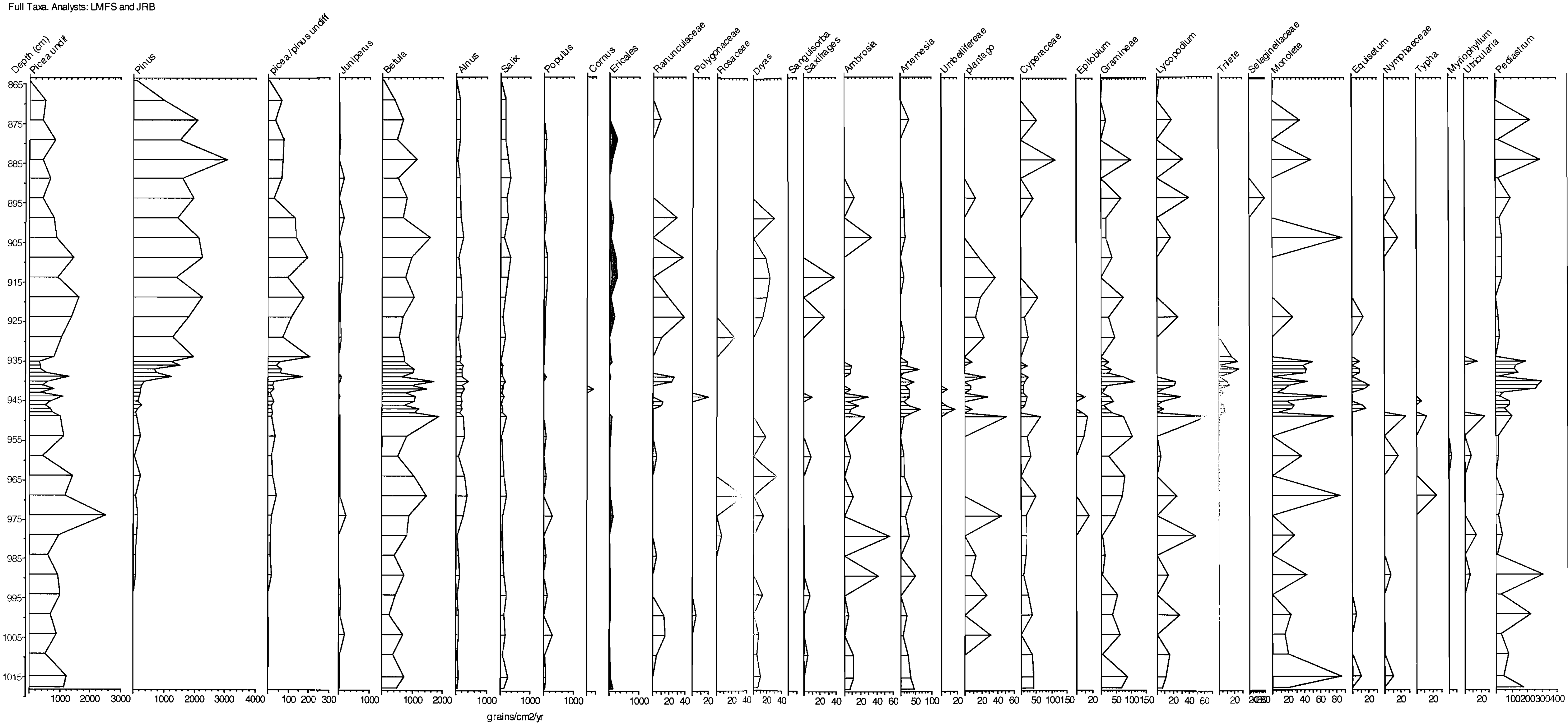
Dragonfly Lake Pollen Influx Diagram
Full taxa. Grey shading represents a 5x exaggeration factor applied
Analyst: LMFS



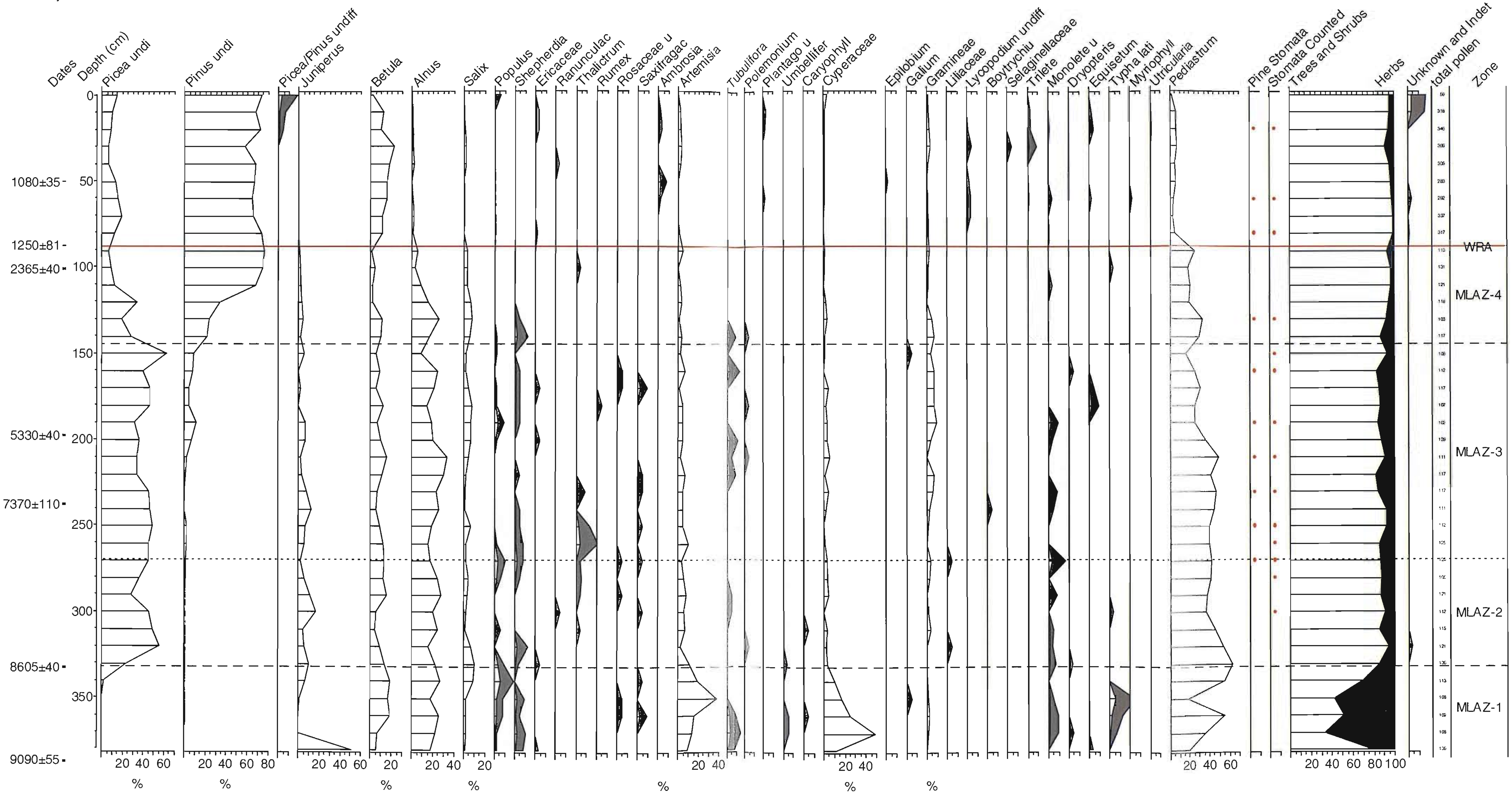
Haircut Lake Pollen Percentage Diagram
Full Taxa. Grey shading represents a 5x exaggeration factor applied
Analysts: LMFS and JRB

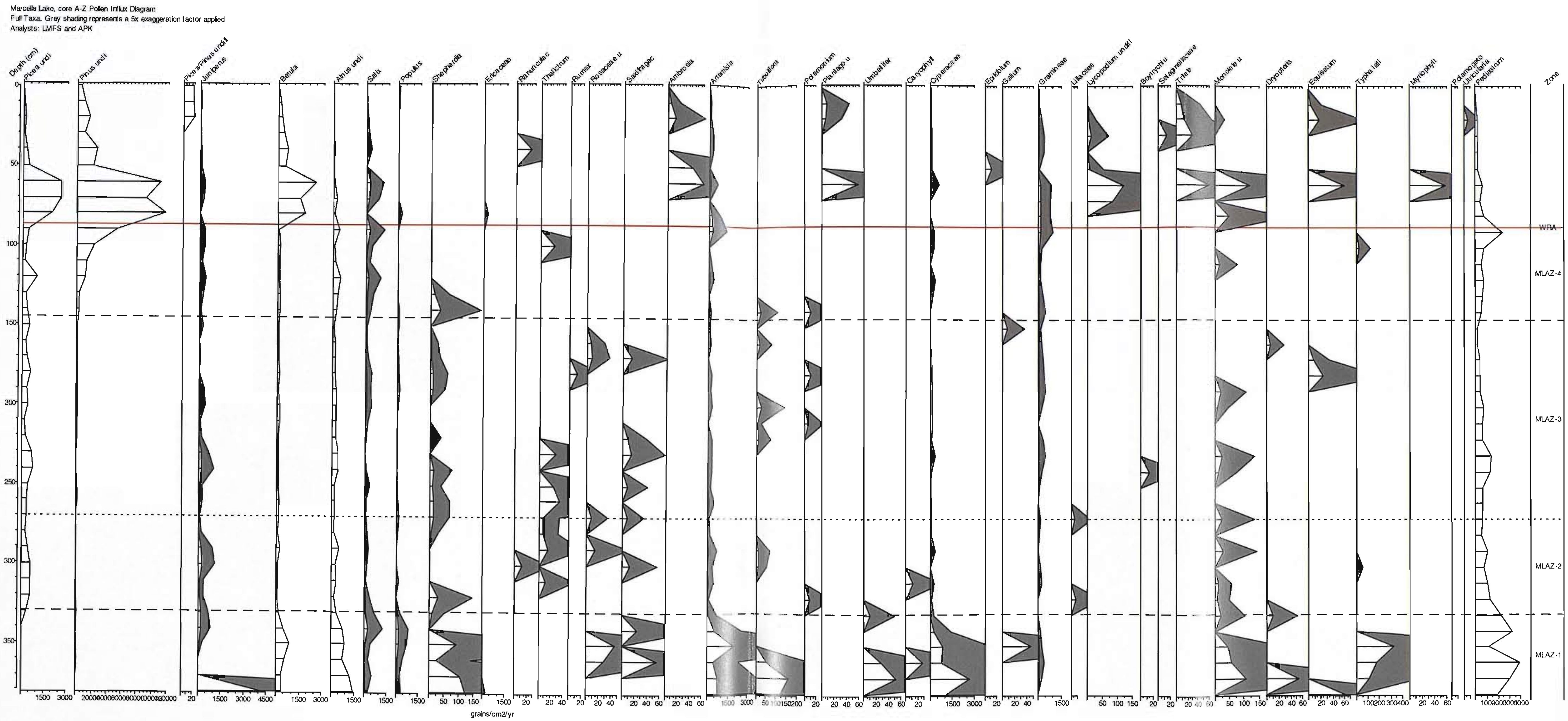


Haircut Lake Pollen Influx Diagram
Full Taxa. Analysts: LMFS and JRB



Marcella Lake Core A-Z, Pollen Percentage Diagram
Full Taxa. Grey shading represents a 5x exaggeration factor applied
Analysts: LMFS and APK





Marcella Lake, core A-Z Pollen Influx Diagram
Full Taxa. Grey shading represents a 5x exaggeration factor applied
Analysts: LMFS and APK
Depth (cm)
Zone
WRA
MLAZ-4
MLAZ-3
MLAZ-2
MLAZ-1
grains/cm2/yr

Appendix 2: Core Correlations

Marcella Lake

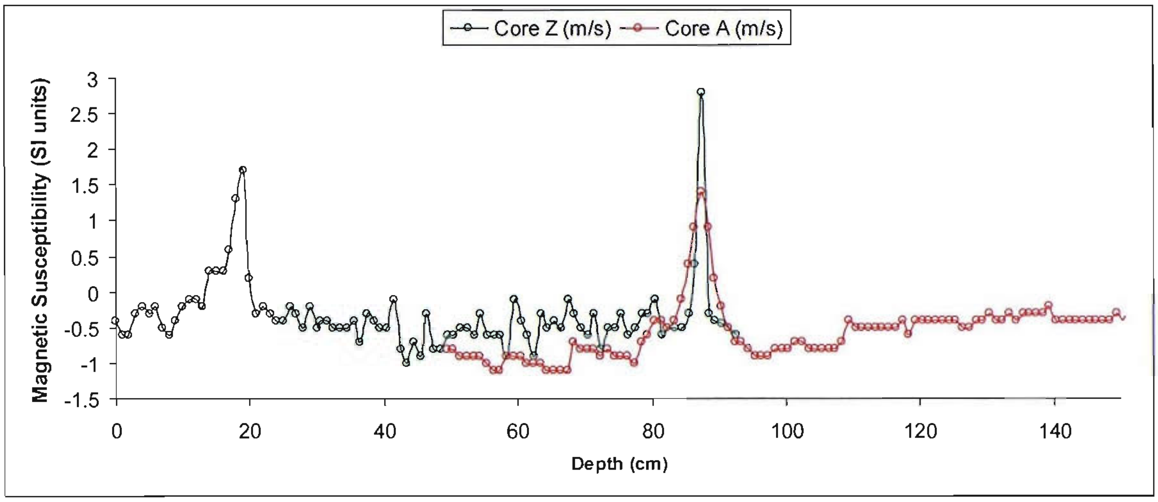


Figure A.1: Core correlation using magnetic susceptibility readings. Red line indicates results for core A adjusted to the depth of core Z (green line). The adjusted depths were used to inform all subsequent interpretations and age-depth relationships as core A-Z.

Birch Lake

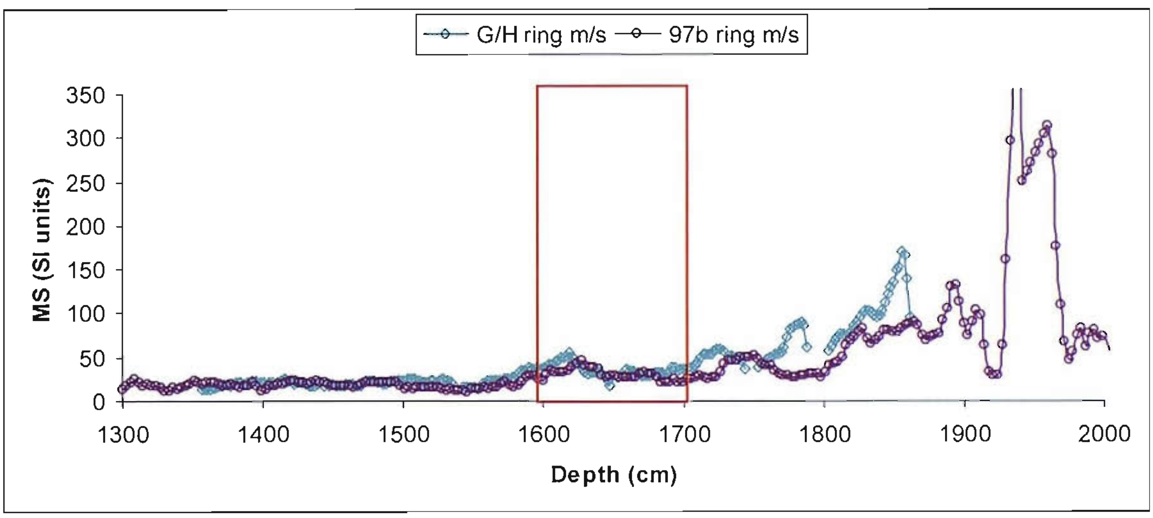


Figure A.2: Core correlation using magnetic susceptibility readings. The purple line indicates results for core 97b adjusted to the depth of core G/H (light blue line). Adjusted depths were used to identify radiocarbon determinations from core G/H to create a complete chronology for the analytical window (red box).

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