HOLOCENE FIRE HISTORY OF THE SOUTHEAST YUKON TERRITORY

by

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ABSTRACT

The global climate is dynamic and has undergone many significant changes throughout Earth's history. Since the industrial revolution, anthropogenic activities have released significant amounts of CO_2 and other greenhouse gasses (GHG) into the atmosphere, which is believed to be impacting global climate. A change to a warmer climate as is predicted by many scientists will have a profound effect on various ecosystems, such as those in northern regions. To better understand the impacts of climate change and disturbances such as forest fires on vegetation, this study examined proxy data to determine how these factors have interacted during the Holocene. The objectives of the thesis were to determine if fire frequency has changed throughout the Holocene, and if so, what is the relation between fire frequency, climate and vegetation. Three sediment cores from small, closed-basin lakes in the southeast Yukon Territory were analyzed for their macroscopic charcoal content and the resulting data was examined using the Charcoal Analysis Programs (CHAPS). CHAPS statistically distinguishes background levels of charcoal from charcoal peaks that are associated with fire events. The charcoal record was also analyzed against pollen percentage diagrams from the surrounding areas, to study the interactions between fire activity, changes in climate and vegetation. The results from the charcoal records indicate that the fire regime of the southeast Yukon was dynamic and responded to changes in climate throughout the Holocene. Fire frequency increased during the Early Holocene, about ~10,000 to 7000 yrs BP, when the climate was warmer and drier than present. Pollen diagrams from the surrounding areas indicate the widespread establishment of *Picea glauca* and *Juniperus spp.*, which is indicative of a warmer and drier climate. Fire

frequency decreased during the mid-Holocene, about ~7000 to 4000 yrs BP when the climate became more cool and moist. *Alnus crispa* and *Picea mariana*, which favor such an environment became more abundant at this time. The charcoal record also provides evidence that periods of increased fire frequency were associated with extensive stands of fire-prone species such as *Picea mariana* and *Pinus contorta*, despite the onset of a more cool and moist climate at ~4000 yrs BP to present-day. This study also examined the effect of elevation on fire frequency, which was found to be highest in the lower elevation closed-canopy forests and lowest in the high elevation alpine tundra.

Further analysis of charcoal and pollen from the southeast Yukon Territory is necessary to further develop the fire frequency record of the area and limit the impacts of site specific conditions and/or topographic parameters which may alter the charcoal signal in a sediment core.

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CHAPTER ONE

1.0 INTRODUCTION

The earth's climate is constantly changing, due to a range of natural factors which operate over a variety of timescales. Although some historical fluctuations in climate have resulted from natural causes such as changes in earth's orbital parameters and solar insolation, the patterns of change that have emerged in recent decades indicate that human activities may now be a significant factor in global temperature patterns (Mann et al., 1998). Anthropogenic activities such as the burning of fossil fuels release greenhouse gasses (GHG) into the atmosphere, which trap the earth's radiative heat and increase global temperatures. Evidence shows that levels of carbon dioxide (CO_2) and temperature have simultaneously increased during the past century (IPCC, 2001). Predictions based on a doubling of CO_2, indicate a further increase in global average temperatures of 1.5 - 5.8°C over the course of this century (IPCC, 2001).

Temperature increases are expected to be particularly enhanced in high latitudes, where the impacts upon boreal and polar regions are expected to be greatest (IPCC, 2001). Mean annual temperature in parts of the arctic has already increased 2 - 3°C from 1954 - 2000, with winter temperatures expected to increase an additional 4 - 7°C by the end of the century (Hassol, 2004). Such dramatic temperature changes will have a number of possible impacts on northern ecosystems, such as an increase in the frequency and intensity of fire events. In northwestern Canada, more intense lightning storms coupled with milder winters and warmer summers will produce an environment ideal for fire occurrence (Natural Resources Canada, 2000). In Yukon Territory, these warmer conditions may double the average annual fire occurrence and area burned by 2069

(McCoy and Burn, 2005). The fire season length in Canada on average may increase by 22% or 30 days in a $2\times CO_2$ climate (Wotton and Flannigan, 1993). Thus, forest fires are expected to increase in both severity and frequency, resulting in a 50% increase in area burned (Stocks et al., 1998). However, the predicted increase in fire activity will still be strongly dependent on other weather variables as well, including precipitation, wind speed and relative humidity (Flannigan et al., 2001).

Significant changes in arctic temperatures may also cause major shifts in ecological boundaries or perhaps a rearrangement of the species found in the present ecosystems (Rizzo and Wiken, 1992). The vegetation type, amount and structure in an ecological community have a direct influence on the characteristics of a fire, or the fire regime. Thus any changes in vegetation type due to changes in climate would have a feedback on the fire regime (Flannigan et al., 2000). It is expected that with a warmer climate, the distribution of some species could shift northward, and change the present-day boundaries of ecological regions. Subarctic tundra may shrink in total area, possibly by 58%, due to the northward encroachment of the boreal forest (Timoney and Wein, 1991). But as the boreal forest shifts northward, its southern boundary will also shift further north, as warmer temperatures increase evaporation and limit soil water availability. As this happens, drought-tolerant grasses from the prairies of southern Canada may shift northward, especially in the continental interior. The total area of the boreal forest may in fact decrease as gains at the northern edge of the ecotone may be exceeded by the area lost at the southern grassland/forest boundary (Lindsey, 2002). This change in vegetation cover may drastically alter the frequency of forest fires in northern

Canada, as boreal forests and grasslands are both more prone to fire activity than the tundra, where fires are infrequent (Timoney and Wein, 1991).

To increase our understanding of how climate change may affect northern fire regimes in the future, we can examine how such relationships operated in the past by studying paleoecological records. Charcoal and pollen records from lake sediment cores are ideal proxies to examine past fire occurrence. Records of past fires can then be evaluated against paleo-reconstructions of climate to develop an understanding of the relations between fire and climate across long timescales. By examining charcoal records from periods in the past when climatic conditions were warmer than today, we can gain insights into the possible impacts of global warming on fire regimes under a warmer $2\times CO_2$ scenario. Additionally, by examining the occurrence of fire prior to anthropogenic activity, we can determine baseline conditions and the natural fire frequency for an area. Our current understanding of boreal fire regimes suggest that forest fires have a frequency or return interval of approximately 100 - 400 years (MacDonald, 2003). Determining the natural cyclicity associated with fire return intervals in this area will be of importance to forest fire managers.

1.1 Research Objectives

This study has a number of objectives related to climate, vegetation and disturbance in southeastern Yukon Territory. A hypothesis related to these objectives will be tested using the data collected for this study as well as previously published data. Because the occurrence of fire is largely determined by complex interactions between abiotic and biotic components of an ecosystem, the hypotheses I have developed are closely linked to one another. The objectives and hypotheses are as follows:

1. To determine how fire frequency in southeastern Yukon Territory has changed throughout the Holocene. It is hypothesized that the fire regime in southeast Yukon is dynamic and has fluctuated throughout the Holocene. Charcoal records from southeast Yukon Territory will be examined to verify this hypothesis.

2. To determine the relation between fire frequency and climate throughout the Holocene. Today, the occurrence and severity of fire is governed by a complex set of abiotic and biotic factors, including fuel type, moisture and availability, weather conditions, vegetation type, and long term patterns of temperature and precipitation (climate) (Bond and van Wilgen, 1996). It is hypothesized that fire frequency in southeast Yukon increased during the early Holocene when the climate was warmer than present, and then decreased during the mid-Holocene with the onset of cooler conditions about ~5000 yrs BP. This second objective will draw upon our understanding of Holocene climate in southeastern Yukon Territory based on previous studies using a variety of paleoclimatic evidence.

3. The relation between fire frequency and vegetation will also be investigated using fossil pollen and charcoal. The third objective will be to examine how fire frequency is influenced by changes in vegetation. A number of palynological studies have been carried out in southern Yukon Territory. The published pollen records from Hail Lake (Cwynar and Spear, 1995), Kettlehole Pond (Cwynar, 1988) and Snowshoe Lake (MacDonald, 1987) will be compared to my charcoal records to determine if certain vegetation types are associated with increased fire activity. Thus, my final hypothesis is that changes in fire frequency were partly driven by changes in the composition of forests in southeast Yukon Territory throughout the Holocene.

CHAPTER TWO

2.0 LITERATURE REVIEW

2.1 The Fire Regime

Fire is an integral part of the environment, and is considered one of the most important stochastic physical forces that influence northern vegetation (Timoney and Wein, 1991). It helps shape the landscape mosaic and influence biogeochemical cycles such as the carbon cycle. Wildfires vary in terms of how often they occur (frequency), when they occur (seasonality) and how intensely they burn (intensity). These three elements combined are termed the fire regime and many plant species are specialized in their response and adaptations to the local fire regime. Fire also varies in type: ground fires burn underground in organic layers of the soil; surface fires burn just above the soil surface; and crown fires burn in the canopy layers of forests (Bond and van Wilgen, 1996).

The fire regime in any given ecosystem is the result of complex interactions between climate, weather conditions, ignition sources, type and amount of fuel, the continuity of fuels on the landscape, and topography (Bond and van Wilgen, 1996; Flannigan et al., 2000). Weather and climate conditions are crucial to the occurrence and growth of fires. The climate regime is important in shaping the weather patterns in a given ecosystem and hence indirectly influences fire activity. The weather prior to ignition is important in determining the fuel moisture, which in turn, will determine if ignition will occur and if the fire will spread (Flannigan et al., 2000). The weather conditions that influence fuel moisture include temperature, precipitation and relative humidity (Bond and van Wilgen, 1996; Flannigan et al., 2000). Fuel flammability will increase during periods of dry, hot weather (MacDonald, 2003). Wind speed affects the spread and intensity of a fire directly, and is the key factor in the growth of a fire, if fuels are available and dry (Flannigan et al., 2000). Winds drive the fire forward and bring the flame front down into contact with flammable ground fuels (MacDonald, 2003). Weather is also important in terms of ignition agents, where lightning is the key ignition agent for naturally caused fires (Flannigan et al., 2000).

Fires will only occur if an ignition source is present. Lightning is the main cause of fires in the environment and on average strikes the land surface of the earth about 100,000 times each day (Goudie, 1997). Lightning is the result of an electrical discharge from a thundercloud, which itself is a result of the appropriate meteorological conditions, namely, atmospheric instability, moisture and a lifting agent (Flannigan et al., 2000). A lightning strike will not always ignite a fire; conditions need to be conducive to creating a fire, specifically, prolonged dry conditions, high air temperatures, low humidity and high winds (MacDonald, 2003). Other natural ignition sources include spontaneous combustion, volcanic activity, meteorite fall or sparks due to rock falls (Cope and Chaloner, 1980). Humans are now becoming a significant cause of contemporary fires in inhabited areas (MacDonald, 2003). In Canada, approximately 65% of all forest fires are ignited by anthropogenic activities (MacDonald, 2003). With increasing populations and encroachment into natural areas, the relative importance of humans as a source of ignition will continue to rise. However, despite the increasing frequency of human-ignited fires, lightning-caused fires still account for 90% of the area burned in the boreal forest (MacDonald, 2003).

The occurrence of fire also depends on the fuel properties of plant material (fuel type, moisture, structure and spatial continuity) and the dynamics of litterfall in the community. The most important factor in determining fire-susceptibility of an environment is the production or retention of dead material. When the accumulation of litter fall is greater than the rate of decay, a higher fuel load will develop and increase the likelihood of a fire (Bond and van Wilgen, 1996). The buildup of sufficient biomass to support a fire requires three conditions: relatively high levels of primary productivity; low levels of decomposition; and low levels of herbivory. Thus, one reason why rainforests rarely burn despite their high rate of productivity, is that they experience high litter turnover and thus large accumulations of dead material do not occur. Coniferous forests are more prone to fires because of their high accumulation of dead material due to relatively low levels of decomposition (Bond and van Wilgen, 1996).

Additionally, the moisture content of dead material is usually low, between 5 - 15% in dry weather and it is the dead fuel rather than the living tissues that initially carry fires (Bond and van Wilgen, 1996). Combustion of this dead material drives out the moisture from living plant parts. These initially act as a heat sink, but once they have dried to a point where they begin to burn, they contribute a large amount of energy to the fire. Living leaves will burn more readily if their moisture content is low, as less water needs to be expelled from the leaves before they ignite. Living plant tissues typically have moisture contents of 50 to more than 250% (expressed as a % of dry mass) (Bond and van Wilgen, 1996). Fires burn readily in the canopies of North American conifer forests, where foliar moisture contents range between 70 - 130% in the dominant trees. The relatively high moisture content (140 - 200%) of the leaves of broad-leaved species such as aspen (*Populus* spp.), birch (*Betula* spp.) and maple (*Acer* spp.) trees on the other hand, effectively prevent the onset of crown fires (Van Wagner, 1977).

Besides the amount and type of fuel that a plant produces, the shape, size and arrangement of plant parts has a large influence in determining susceptibility to fire. Surface area to volume ratio is important because it influences the rate of moisture loss, as small-diameter plant parts lose moisture more quickly than larger ones (Bond and van Wilgen, 1996). Examples of fire-prone species include pines, grasses and heaths with their finely divided leaves and large surface area. Some trees and shrubs maintain dead lower branches that allow flames to climb from the ground to the canopy. Many species of pine (*Pinus* spp.) posses dead lower limbs or "fire ladders", which allow flames to move up to the canopy layer and increase the intensity, spread and areal extent of the fire (MacDonald, 2003). In addition, the conical shape of conifers also promotes fires and their rapid spread (MacDonald, 2003).

The content of oils, fats, waxes and terpenes in plant parts also increases the susceptibility to fire (Bond and van Wilgen, 1996). Terpenes are volatile and make plants more flammable, because they ignite much easier and then burn as gases, increasing the intensity of the fire. Species of pine, such as jack and lodgepole pines (*Pinus banksiana* and *Pinus contorta*) have been found to be much more flammable than associated deciduous trees such as trembling aspen (*Populus tremuloides*) and paper birch (*Betula papyrifera*) (MacDonald, 2001).

The spatial arrangement of plants also appears to be important in determining whether vegetation will burn or not. Fire-prone communities like the coniferous forests experience catastrophic stand clearing fires due to the densely-packed nature of the tree crowns (Bond and van Wilgen, 1996). In contrast, deserts and arctic tundra ecosystems have discontinous fuel beds and thus rarely experience any fire activity (Bond and van Wilgen, 1996).

Geographical generalizations can be made regarding the role of fire as an agent of disturbance. There is a large difference between the frequency of fires in tundra regions and seasonally dry areas found in western North America. The arctic tundra has many lakes and rivers that serve as topographic fire breaks; the tundra is also cold, moist, has few thunderstorms and supports sparse, slow-growing vegetation. The southern tundra areas burn about once every 1000 years, and fires are generally small (MacDonald, 2003). Most of the northern arctic almost never experiences fires (MacDonald, 2003). In contrast, seasonally dry areas in western North America dominated by stands of ponderosa pine *(Pinus ponderosa)* and Douglas fir (*Pseudotsuga menziesii*) will burn almost annually (MacDonald, 2003). Fires are more common in this ecosystem since it is made up of highly flammable tree species, with low moisture contents. In addition, these fire-susceptible communities accumulate well-aerated fuel beds with relatively large amounts of fine dead material, due to low levels of decomposition (Bond and van Wilgen, 1996).

2.2 Influence of Topography on Fire Activity

Spatial and temporal variation in the biophysical environment, specifically vegetative structure, terrain, and weather lead to variations in fire behavior (Ryan, 2002). The topography can influence the spread of a fire through natural fire breaks such as lakes, rivers and ridges (Flannigan et al., 2000). In the central boreal forest, areas with lakes and sharp topography may not burn for 100 - 400 years (MacDonald, 2003). By contrast, dry areas with sandy soils and little topographic relief can have average fire-return intervals as short as 28 years (MacDonald, 2003).

However, topography has a more indirect role in influencing the fire regime, by primarily determining the type and structure of plant communities in a given region. Properties such as slope, aspect, elevation and soil type affect site energy and water budgets, which in turn influence the growth of specific plant communities. These terrain features also have a major influence on the moisture and flammability of fuels (Ryan, 2002). As previously mentioned, vegetation structure and composition affect the type and mass of fuel available for burning and therefore the fire potential. Consequently, lower valley regions dominated by boreal forest may experience a higher fire frequency than upper regions dominated by alpine tundra.

Evidence from northwestern United States has shown that fire frequency varies with elevation, where elevation and aspect interact to create gradients in snow-cover duration and also where steep talus interrupts fuel continuity (Heyerdahl et al., 2001). Fire frequency at the local scale also varied with different parameters of topography (aspect, elevation and slope) in watersheds with steep terrain, but not in watersheds with gentle terrain (Heyerdahl et al., 2001).

2.3 Charcoal as a Proxy of Fire Occurrence

The analysis of charcoal in lake sediments provides a useful method for determining long-term variations in fire occurrence. Charcoal studies can complement and extend short-term reconstructions provided by dendrochronological and historical records, thus providing a more comprehensive and complete fire history for a region. Since charcoal records can span over several millennia, they allow for the examination of fire regimes against other long-term trends such as major vegetational shifts and changes in climate.

Charcoal is an inorganic carbon compound which is produced when a fire incompletely combusts organic matter at about 280 - 500°C (Whitlock and Larsen, 2001). Charcoal particles can be identified by their dark, opaque, angular and usually planar characteristics. There are a number of factors that can affect the production of charcoal, including the nature of the fire and the fuel conditions that affect combustion efficiency. Fires of low intensity are known to produce large amounts of particulate matter due to their low combustion efficiency (Whitlock and Larsen, 2001). High intensity fires (with long flame length) produce proportionately larger charcoal particles than low intensity fires, as turbulent winds move them beyond the combustion zone (Whitlock and Larsen, 2001). Fire regimes that are characterized by frequent and efficient ground fires do not produce much charcoal, as they are generally small in size and charred particles are not carried aloft (Whitlock and Larsen, 2001). In contrast, prairie fires which are typically cool and fast, produce significant amounts of charred herbaceous material (Whitlock and Larsen, 2001).

Once created, charcoal particles encounter a number of taphonomic processes, which transport the particles from their source to the depositional basins in which they are preserved. The rate at which charcoal accumulates in lake sediments depends on a number of factors, most importantly on the processes that transport and deliver charcoal to the lake (Whitlock and Larsen, 2001). *Primary* charcoal refers to the material that is introduced during or shortly after a fire event and is a direct measure of biomass burning. *Secondary* charcoal is introduced during non-fire years, as a result of secondary processes such as surface run-off and lake-sediment mixing (Whitlock and Larsen, 2001). Secondary charcoal from a single fire may be introduced into lake sediment over a period of years, which makes it more difficult to isolate individual fire events (Whitlock and Millspaugh, 1996). Convectional fire activity, atmospheric winds and water are the main agents responsible for the transportation of charcoal material.

The aerial transport of charcoal is dependent upon the fire size, intensity and severity (Whitlock and Larsen, 2001). During a fire, indrafts and convection cause turbulence that suspends particles in the combustion zone, and then thermal buoyancy provides the necessary energy to loft particles well above forest canopies (Clark, 1988). About 90% of these particles that are lifted upwards are microscopic (< 100 μm) and can be transported large distances by atmospheric winds (Clark, 1988). Thus, microscopic charcoal provides a more regional indication of fire activity. Macroscopic particles (>100 μm) are not as easily transported due to their larger size and thus provide a more local record of fire occurrence. Numerous studies including Whitlock and Millspaugh (1996), Clark et al. (1998), Ohlson and Tryterud (2000) and Gardner and Whitlock (2001) confirm that macroscopic charcoal is not transported far from the fire margin. However, recent observational evidence indicates that entire needles and other plant remains can be transported 10s of kilometers by intense fires that generate updrafts (Pisaric, 2002). Therefore, in mountainous regions, the "local" nature of macroscopic charcoal is not completely understood.

Following a fire, charcoal can also be transported to lakes or catchment basins by surface run-off or streams. Surface run-off may deliver charcoal for a short period after a fire, but the importance of this process diminishes as the watershed becomes revegetated

(Whitlock and Larsen, 2001). Alternatively, Clark (1988) suggests that surface water run-off does not play an important role at all in charcoal transport, due to the uncompacted nature of forest soils. Meyer et al. (1995) produced evidence that surface run-off contributed very little to the charcoal record at Yellowstone National Park due to the uncompacted nature of the soil. In burned landscapes with steep slopes, surface-runoff may play a more important role, due to increased erosion, unvegetated ground, hydrophobic soils and reduced infiltration (Whitlock and Larsen, 2001). Mass-wasting events triggered by intense rains can also transport significant amounts of charcoal to a lake basin. Hydrological processes may continue for several years after the actual fire and thus are largely responsible for the introduction of secondary charcoal in the charcoal record. It is also important to note that charcoal deposition in alpine and steep environments may be more strongly influenced by natural forces such as water and wind erosion and may change the charcoal composition found in the sediment. Consequently, the charcoal concentration may not reflect the true fire frequency at the site.

Charcoal data can be displayed in a stratigraphic profile, in which a charcoal peak represents a fire event. Lags in charcoal accumulation may produce broad peaks which still represent a single fire event, but years of charcoal accumulation (Whitlock et al., 1997; Laird and Campbell, 2000). Vegetation reconstructions may be placed along side the charcoal diagram, and analyzed with respect to fire occurrence and subsequent vegetation response.

2.4 Fundamental Principles of Paleoecological Research

The study of past environments using lake sediment relies upon a number of principles that largely developed from stratigraphical studies in the geological sciences.

Fundamental to these studies is James Hutton's principle of uniformitarianism, which states that "the present is the key to the past" (Murck et al., 1996). It implies that the physical and biological processes that link today's climate and vegetation must have operated in a similar fashion and at similar rates in the past.

Another underlying principle is the principle of superposition. This principle states that within a sequence of undisturbed sediment, the sedimentary layers are oldest at the bottom of a profile and youngest at the top (Murck et al., 1996). This principle is vital to the study of pollen and charcoal, as the examination of fossils preserved in the sedimentary layers of a lake sediment core consist of assemblages of pollen and charcoal that have changed over time due to biotic and abiotic factors. Changes that occur in the deepest sections of a lake sediment core represent changes that occurred some time in the past. Those that occur near the top of the lake sediment core represent more recent changes in vegetation and disturbance regimes.

2.5 Paleoecological Investigations of Holocene climate, vegetation and fire frequency in North America

Quantitative paleoecological research on fire frequency has not been extensive for northwestern Canada. However, studies in western Canada and the United States have investigated past climate and fire frequency using various proxy indicators. The analysis of pollen found in sediment is the most widely utilized technique to determine past climate change. Pollen grains can be used to identify specific vegetation, as they are characteristic of their parent plant species and are well preserved in an anaerobic environment. This technique has been utilized for over a century to understand past climatic changes, successional change of vegetation and vegetation response to disturbances (Bennet and Willis, 2001). Past studies have also focused on the use of an indicator species method, to estimate past climatic conditions from pollen data. The indicator species approach requires knowledge of the climatic limits of one or more plant species. The presence or absence of the species in the fossil record is used to infer past climatic conditions (MacDonald, 1990). Pollen can also be used indirectly as a proxy for fire disturbance. The increase in pollen percentages and accumulation rates of shade intolerant plants has been cited as evidence of canopy opening by fire (e.g. Cwynar, 1978, 1987).

Charcoal particles can be used as a proxy for fire occurrence. Charcoal found in sediments is used to reconstruct long-term variations in fire frequency on the basis that charcoal provides direct evidence of organic burning. Stratigraphic levels with abundant charcoal are thus interpreted as periods of past fire activity (Gardner and Whitlock, 2001).

2.5.1 Holocene climate change and its effects on vegetation and fire frequency in Northwestern Canada and Alaska

The climate of the Holocene has been dynamic, fluctuating in temperature and precipitation and resulting in variations in vegetation composition and fire frequency. Studies from northwestern and central Canada based on pollen assemblages and charcoal records, have shown that the climate of the Holocene went through a series of changes which had significant impacts on vegetation and fire activity.

<u>The Early Holocene</u>

Shortly after deglaciation in northwestern Canada, about 14,000 – 11,000 yrs BP, the climate was cold and dry. Following the retreat of ice ~12,000 yrs BP, *Betula* spp.

shrub tundra dominated throughout most areas of Alaska and Yukon Territory, while herb-tundra vegetation was established in the Northwest Territories (Cwynar and Spear, 1991; Earle et al., 1996; Szeicz and MacDonald, 2001). Warmer and drier conditions began around 11,000 - 9000 yrs BP, most likely due to an increase in summer insolation corresponding to the early Holocene thermal maximum (Ritchie et al., 1983). Summer insolation at this time peaked 8% above the present level in the high latitudes of North America, resulting in a 1- 2°C increase in temperature in these northern regions (Cronin, 1999). During this period of increased warmth, balsam poplar (*Populus balsamifera*) and white spruce (*Picea glauca*) dominated areas of north-central Alaska (Earle et al., 1996). A charcoal record from Sithylemenkat Lake in north-central Alaska revealed a high charcoal accumulation rate during this period, indicating that fire was an important agent of disturbance in this environment (Earle et al., 1996). However, in the interior of Alaska, evidence showed that fires did not become frequent until the establishment of fire-prone black spruce (*Picea mariana*) forests after 5500 yrs BP (Lynch et al., 2003).

In central Yukon around 10,000 yrs BP, a discontinous forest of balsam poplar was present, but was then replaced by white spruce between 9400 - 8900 yrs BP (Cwynar and Spear, 1991). The establishment of balsam poplar also occurred around 11,000 yrs BP in northern Yukon, and was followed by an expansion in white spruce at 10,200 yrs BP (Cwynar, 1982). Pollen influx also increased between 11,100 and 8900 yrs BP, and further supports the occurrence of an early-Holocene warm interval (Ritchie et al., 1983).

In the shrub tundra zone of the Tuktoyaktuk Peninsula, Northwest Territories, spruce pollen influx increased during the interval from 10,000 to 6000 yrs BP, suggesting that summer temperatures during the early Holocene must have been significantly warmer than mid to late Holocene climates (Ritchie et al., 1983). Evidence of an elevational increase in treeline was also found in the Mackenzie Mountains, NWT, along with an increase in spruce (*Picea* spp.) pollen, suggesting warmer conditions ~8000 yrs BP, during the early to mid-Holocene (Szeicz and MacDonald, 2001). This treeline advance has been found to be a general feature of eastern Beringia and far western Beringia (Siberia) (e.g. Pisaric et al., 2001) and likely the result of increased summer insolation due to Milankovtich orbital variations (Szeicz and MacDonald, 2001).

Hebda (1995) suggests a warmer and drier climate throughout British Columbia at the beginning of the Holocene. A study by Pisaric et al. (2002) in northern B.C. noted an increase in spruce stomata at approximately ~8000 yrs BP, indicating a shift to higher-density forests, with probably higher fuel loads. Accordingly, macroscopic charcoal concentrations were found to be highest between ~8400 and 7400 yrs BP, indicating a higher fire frequency as a result of the higher fuel loads and warmer climate of this period (Pisaric et al., 2003). In southern B.C., lake sediment charcoal records suggest that fires were more frequent in the early Holocene from 11,000 - 8200 yrs BP, when the climate was warmer and drier (Hallet and Walker, 2000; Hallet et al., 2003).

However, a brief cooling event occurred around 8200 cal yrs BP, potentially due to the final collapse of the Laurentide ice sheet and the drainage of the Agassiz and Ojibway glacial lakes through the Hudson Strait to the Labrador Sea (Seppa et al., 2003). This freshwater pulse may have led to a transient weakening or shut-down of the North Atlantic Thermohaline circulation around 8470 cal yrs BP, and a drop in temperature by about 2°C over the greater North Atlantic region (Kurek et al., 2003). At 8100 - 7900 cal. yrs BP, an abrupt decline in *Betula spp.* pollen in the lake sediments from the low arctic of continental Nunavut, suggests a decrease in temperature due to an abrupt disturbance such as the 8200 cal. yrs BP event (Seppa et al., 2003).

The Mid-Holocene

Conditions changed again at about 8000 - 6000 yrs BP, as temperatures began to decrease and precipitation increased. At ~8000 yrs BP, a decline in charcoal accumulation was associated with a decrease in white spruce and a widespread expansion of alder (*Alnus* spp.) shrubs in north-central Alaska (Earle et al., 1996). This change in vegetation was presumably caused by an increase in effective moisture due to decreasing temperatures and increasing precipitation (Anderson and Brubaker, 1994; Earle et al., 1996). An increase in black spruce ~6500 yrs BP and its regional expansion was potentially the result of a climatic cooling (Anderson and Brubaker, 1994). Although a cooler climate would not have increased fire frequency, charcoal accumulation rates increased slightly at this time, in response to the structural characteristics of black spruce forests which make them more fire-prone than the previous established species (Earle et al., 1996). An increase in fire occurrence was also noted in the interior of Alaska ~5500 yrs BP, also due to the expansion of black spruce forests (Lynch et al., 2003).

At ~6500 yrs BP, black spruce populations began increasing in central Yukon, leading to the development of white and black spruce woodlands (Cwynar and Spear, 1991). As the climate became cooler at 5000 yrs BP, the spruce forests gave way to shrub tundra, with only localized groves of white spruce (Cwyner and Spear, 1991). Similarly, evidence from the Mackenzie Mountains along the border of Yukon and NWT, show a reversion of the forest-tundra towards more open conditions and/or a drop in treeline beginning ~5000 yrs BP (Szeicz and MacDonald, 2001).

Hebda (1995) indicates an increase in precipitation at about ~7000 yrs BP in most areas of British Columbia. This onset of humid conditions allowed for the establishment of rainforest taxa in southwestern B.C. with a variable fire frequency (Hallet et al., 2003). However, evidence was found in the Rocky Mountains of a warm and dry hypsithermal period from 8200 – 4000 cal yrs BP (Hallet and Walker, 2000). Forests were dominated by pine, larch (*Larix* spp.) and open meadows of grass (*Poaceae* spp.) and fire frequency increased and reached its maximum during this period (Hallet and Walker, 2000).

Studies in central Canada have also yielded evidence of a mid-Holocene warming, with warmest temperatures occurring between 5700 – 4500 cal yrs BP (Seppa et al., 2003). Pollen analyzed from a lake sediment core from central Northwest Territories revealed a transformation from tundra to forest-tundra and an advance of the present-day treeline during this warmer period, which corresponds well with fossil pollen and paleosol studies from more easterly sites in central Canada (MacDonald et al., 1993). The warmer temperatures in this region at ~5000 yrs BP do not coincide with the early-Holocene summer insolation maximum, but instead may be attributed to the later retreat of the Laurentide ice sheet or a northerly displacement of the Arctic front (MacDonald et al., 1993; Seppa et al., 2003).

The Late Holocene and Onset of Modern-day Climate

The onset of modern climatic conditions commenced at ~4000 yrs BP across most of northwestern Canada, when conditions became relatively cool and moist. As a result, areas of the Yukon, Northwest Territories and Nunavut reverted back to tundra or shrub-tundra (Ritchie et al., 1983; Cwynar and Spear, 1991; MacDonald et al., 1993; Seppa et al., 2003). In southern B.C., fires became less frequent at about ~4000 yrs BP with the onset of a cooler and wetter Neoglacial climate (Hallet and Walker, 2000; Hallet et al., 2003).

In contrast, a number of studies have shown a more dynamic late-Holocene (Neoglacial) climate (Szeicz and Macdonald, 2001). Studies in western Canada have shown that a period of warmer temperatures occurred around ~3400 to 2600 yrs BP (Szeicz and MacDonald, 2001; Pisaric et al., 2003). Evidence was found of an increase in forest-tundra density and/or treeline advance around ~3000 yrs BP in the Mackenzie Mountains (Szeicz and MacDonald 2001). Warming during this period has also been identified in the southern Canadian Rockies, where summer temperatures may have risen as much as 1.8°C (Kearney and Luckman, 1987). This slight increase in temperatures provides evidence for century-scale climatic fluctuations superimposed on longer-term variations resulting from changes in insolation (Szeicz and MacDonald, 2001).

2.5.2 Effects of Climate Change on Fire Frequency throughout the Holocene in Southeast Yukon

Relatively few studies of the long-term frequency of fire have been carried out in northwestern Canada, especially in Yukon Territory. Therefore, the current study will examine how climate change has affected fire frequency in Yukon Territory since the beginning of the Holocene. To examine changes in fire frequency I am using macroscopic charcoal. A charcoal record of fire history will be developed and analyzed using the CHAPS (Charcoal Analysis Program) computer program. This will help quantify changes in fire frequency by statistically separating peaks in charcoal due to actual fire events from a constantly occurring background level (Long et al., 1998). The fire history will then be compared to existing vegetation reconstructions, which indirectly indicate climatic conditions across Yukon Territory throughout the Holocene.

With increased knowledge of past vegetation assemblages and disturbance regimes in the boreal region under changing climatic conditions, we can better predict the future impacts of global warming in northwestern Canada. This is especially important for communities whose social and economic livelihood is based on forestry. Forestry is an important sector of Yukon's economy and plays an important role in First Nation's culture and history (Forest Management Branch, 2005). The Forest Management Branch of the Department of Energy, Mines and Resources oversees the development and management of Yukon's forest resources. The Forest Management Branch has been working to develop Yukon forest legislation; carrying out and implementing strategic forest management plans in partnership with Yukon First Nations; and preparing for a revival of the Territory's forest industry (Forest Management Branch, 2005). The Kaska Forest Resource Stewardship Council (KFRSC) is currently discussing a plan to allow the use of certain areas of boreal forest of the Kaska traditional territory in southeast Yukon. Economic studies have shown that Yukon lumber imports could be replaced with local product, generating significant employment and approximately $5 million per year of economic activity in southeast Yukon (KFRSC, 2005). Thus, the results of this study can help Yukon communities build a better strategic forest management plan, which can include the role of fire in boreal forests and the effects of a warmer climate on fire activity.

On a larger scale, the results of this study may be incorporated into GCMs. In order to develop better models and produce more realistic outputs of future climatic conditions, GCMs should be able to reproduce both present and past conditions. GCMs can be tested and their ability to model climate can be assessed using paleoecological data such as that collected and developed by this study.

CHAPTER THREE

3.0 STUDY SITE AND METHODS

3.1 Southeast Yukon Territory

Yukon Territory is a topographically diverse region, being comprised of 23 ecoregions which are largely defined in terms of physiography, drainage, geological setting including glacial history, climate, topography and vegetation (Smith et al., 2004). The study lakes are located west and northwest of Watson Lake, in the southeast corner of Yukon Territory (Fig. 3.1). The sample sites are located within the Liard Basin and Pelly Mountains ecoregions, which are part of the Boreal Cordillera ecozone (Fig. 3.1) (Smith et al., 2004).

3.2 Vegetation

Throughout the southeast Yukon Territory, white spruce dominates the arboreal vegetation. White spruce occurs in association with black spruce, lodgepole pine paper birch and trembling aspen in lowland forests. Black spruce occurs on poorly drained surfaces in the low elevation forests, although not exclusively. White spruce, trembling aspen and paper birch occur on more favorable sites where drainage is better. Trembling aspen is also abundant on disturbed sites at low to mid-elevations, especially where fire has occurred recently. With increasing elevation, the closed canopy forests grade into forest-tundra, where open-grown white spruce occur in association with subalpine fir (*Abies lasiocarpa* (Hook.) Nutt.) and shrub birch (*Betula glandulosa* Michx.). At the highest elevations, typically above 1400 - 1500 m a.s.l.,

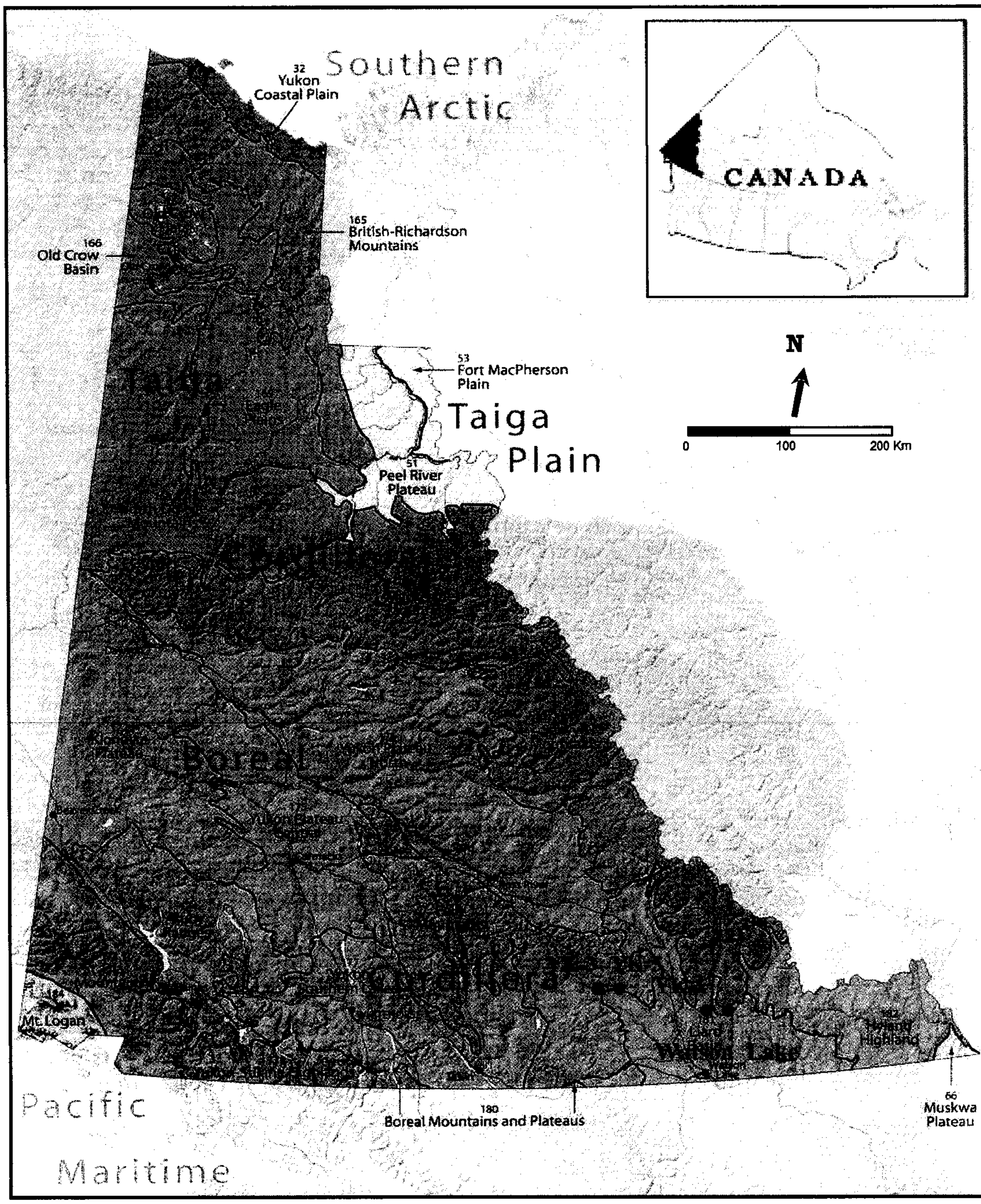


Fig. 3.1 Ecozones of the Yukon Territory (Smith et al., 2004). The study site is located in the Boreal Cordillera ecozone. The location of the Town of Watson Lake and the three core sites, YK-1, YK-3 and YK-5 are depicted on the map.

shrub tundra is the dominant vegetation cover. The dominant species include shrub birch and willow (*Salix* spp.). White mountain heather (*Cassiope tetragona*), mountain avens (*Dryas integrifolia*), *Arctostaphylos* spp. and *Vaccinium* spp. are also present, but at low densities. An assortment of herbs, including mountain sorrel (*Oxyria digyna*), one-flowered cinquefoil (*Potentilla uniflora*), various species of grass, bryophytes and lichens are also present.

3.3 Climate

The climate of this region is classified as an interior subalpine type (Environment Canada, 2005). Mean annual temperatures are about -3.5°C, ranging from a mean of -22.5°C in January to about 10°C in July (Smith et al., 2004). Extreme temperatures range from -59°C to 34°C (Smith et al., 2004). Winters are cold and long, often lasting from October to May. Summers tend to be short, but relatively warm. These temperature regimes are modified at higher elevations. Mid-winter thaws can occur in the warmer southern valleys, but are not as common as in southwestern Yukon (Smith et al., 2004).

Precipitation is moderate, with mean annual precipitation ranging from 400 - 650 mm, with 35 - 60% falling as snow (Smith et al., 2004; Environment Canada, 2005). The driest months are February to June, while most precipitation falls between September and January (Smith et al., 2004). Rain showers and thunderstorms are predominant throughout summer. Winds are generally light, but prolonged periods of moderate easterly winds can occur during the winter (Smith et al., 2004).

Above treeline, alpine weather is typical. This area is cold, windy and snowy and characterized by low temperatures during the growing season and a short frost-free period. Mean annual temperature ranges from -4 to 0° C (Environment Canada, 2005).

Frost can occur at any time and the average temperature remains below freezing for seven to 11 months each year (Environment Canada, 2005). Mean annual precipitation is 700 to 3000 mm, 70 to 80% of which falls as snow (Environment Canada, 2005).

3.4 Surficial Geology and Glacial History

During the Pleistocene, north-central British Columbia and southern Yukon Territory were repeatedly affected by at least six glacial advances originating from the Cordilleran region (Jackson et al., 1991). The surficial geology of the Laird Basin and Pelly Mountain ecoregion was mostly developed during the most recent glaciation, known as the McConnell glaciation (from about 28 000 to 15 000 yrs BP). During the McConnell glaciation, an ice cap covered all but the peaks of the Pelly Mountains and Cassiar Mountains and fed the Selwyn, Cassiar and Liard lobes of the Cordilleran ice sheet (Smith et al., 2004). Ice flowed into the Laird ecoregion from the Pelly and Selwyn mountains to the north and the Cassiar Mountains to the west (Smith et al., 2004). A trunk glacier the width of the Liard Plain flowed down the Laird Valley, where it merged with ice from the northwestern Rocky Mountains. The trunk glacier most likely contacted the retreating Laurentide ice sheet in the Mackenzie Valley about 23 000 yrs BP (Smith et al., 2004). Deglaciation was complete by 10 000 yrs BP in the Pelly Mountains and well before 9000 in the Liard Basin (Smith et al., 2004). The pattern of ice buildup was largely controlled by physiography and location relative to the prevailing westerly winds, while ice decay was largely a product of physiography (Fulton, 1991).

The topography and surficial materials of the east-central and southern Yukon reflect sculpting by the Cordilleran and montane glaciers. In general, valleys have steep sides with bottoms filled with silt, sand and gravel. Alpine areas have abundant exposed bedrock with active talus aprons, cirques and rock glaciers (Smith et al., 2004). Although the Cordilleran ice sheet did not cover the Pelly and Cassiar Mountain peaks, the lower summits that were overtopped by ice have exposed, and sometimes polished bedrock with interspersed pockets of glacial till (Smith et al., 2004). Glacial erratics and lateral moraines are found in the northwest of the Pelly Mountains ecoregion. During the post-glacial period, streams incised the glaciated terrain and created alluvial terraces and built alluvial fans. Intense mechanical weathering and mass-wasting processes created colluvial mantles on mountain slopes (Smith et al., 2004).

3.5 Soils

The soils of the study area have formed under the influence of moderate to high precipitation, warm summers, and varying elevational gradients (Smith et al., 2004). Brunisols are common throughout the Boreal Cordilleran ecozone, which form under forest cover and grasslands in southern and central Yukon. Dystric Brunisols, forested soils with acidic soil horizons, can be found in the higher elevations of the Pelly Mountains, due to high precipitation causing strong leaching of the soil. These soils support extensive conifer forests composed of subalpine fir, spruce and pine (Smith et al., 2004).

The warm summers, moderate precipitation and fine-textured parent materials in the valleys and basins of the Liard basin ecoregion, result in soils containing subsurface clay accumulations belonging to the Grey Luvisol soil group (Smith et al., 2004). The soils above treeline are acidic and often show evidence of permafrost cryoturbation and patterned ground formation (Smith et al., 2004). Permafrost is sporadic throughout the study area, and is common on north-facing slopes at higher elevations of the Pelly Mountains (Smith et al., 2004).

3.6 Charcoal Study Sites

Lake sediment cores were collected from three small lakes during the spring of 1998. The lakes are located near the Town of Watson Lake and are informally named YK-1, YK-3 and YK-5 (Fig. 3.1). The lakes were sampled across an elevational gradient (Table 3.1).

YK-1 is located at the lowest elevation and surrounded by a closed-boreal forest. Dominant arboreal vegetation includes white and black spruce, lodgepole pine, paper birch and trembling aspen. YK-5 is at a mid-elevation in the forest-tundra ecotone. Dominant vegetation includes white spruce, shrub birch and subalpine fir. YK-3 is located in the high alpine-tundra ecotone. The shrub tundra consisted of birch and willow with abundant herbs, grasses and heaths.

Table 3.1 Site characteristics of YK-1, YK-3 and YK-5.

Site	Latitude/Longitude	Elevation (m)	Area (ha)	Maximum depth (m)
YK-1	60° 22' N, 131° 09' W	920	2.3	2.4
YK-3	60° 26' N, 131° 32' W	1510	12.0	6.1
YK-5	60° 22' N, 131° 09' W	1310	5.6	3.3

3.7 Pollen Study Sites

Detailed descriptions of the main pollen study sites discussed in this thesis can be found in Cwynar and Spear (1995) (Hail Lake), Cwynar (1988) (Kettlehole Pond) and MacDonald (1987) (Snowshoe Lake). These three sites are located in southern Yukon and were chosen by this study due to their close proximity to the charcoal study sites (Table 3.2).

Table 3.2 Site characteristics of Hail Lake, Kettlehole Pond and Snowshoe Lake.

Site	Latitude/longitude	Elevation (m)	Area (ha)	Max Depth (m)
Hail Lake[1]	60° 02' N, 129° 01' W	690	1.0	7.39
Kettlehole Pond[2]	60° 02' N, 133° 08' W	600	n/a	n/a
Snowshoe Lake[3]	57° 27' N, 120° 40' W	900	8.0	2

[1]Cwynar and Spear (1995)

[2] Cwynar (1988)

[3]MacDonald (1987)

3.8 Field Methods

Sediment cores were recovered from the lakes in the spring of 1998. Coring was carried out from the ice surface. All cores were recovered using a Livingstone square-rod piston sampler with an internal diameter of 4.5 cm (Wright et al., 1984). To insure that a complete record was obtained and to capture the sediment-water interface without any significant disturbance, short surface sediment cores were also collected from each lake using a gravity corer (Glew, 1991). Long sediment cores were recovered in 1 m segments and extruded in the field. Cores were wraped in plastic and aluminum foil, placed in plastic tubing and packed in wooden boxes before shipping to Queen's University in Kingston, Ontario. Short cores recovered using the gravity corer were sectioned in the field at contiguous 0.5 cm intervals using a Glew close-interval extruder (Glew, 1991). Extruded sediment was placed in small Whirl-Pak ™ sample bags. All samples were stored in cold storage facility at Queen's University at 4°C. Samples were transferred to cold storage facilities in the Department of Geography and Environmental Studies at Carleton University in 2002.

3.9 Laboratory Analyses

3.9.1 Radiocarbon Dating

Radiocarbon measurements of fossil organic matter are based upon the radioactive decay of ^{14}C, an isotope that is incorporated into living matter. Upon death of a living organism, ^{14}C is no longer taken up. At this point, ^{14}C begins to decay at a known rate. The ^{14}C decreases at a rate known as its half-life, meaning it decomposes by one half in a particular time period. The half-life of ^{14}C is 5568 years (Bjorck and Wohlfarth, 2001). By measuring the remaining radioactivity within the organic material it is possible to determine the time that has elapsed since it was initially isolated.

A total of 15 samples were recovered from the lake sediment cores and submitted for radiocarbon dating to IsoTrace Laboratories at the University of Toronto and the radiocarbon laboratory at the University of Arizona. Bulk sediment samples were submitted for dating as macrofossils were not recovered from the cores. Samples were taken from YK-1 at 75 cm, 150 cm, 240 cm, 300 cm and 350 cm; YK-3 at 57 cm, 85 cm, 115 cm and 130 cm; YK-5 at 38 cm, 90 cm, 164 cm, and 227 cm.

3.9.2 Loss-On-Ignition (LOI)

Loss-on–ignition measures the organic and carbonate content of organic-rich lake sediment (Dean, 1974). The method is based on the premise that organic material begins to burn at approximately 200°C and is completely combusted at temperatures approaching 550°C (Santisteban et al., 2004). Higher temperatures combust inorganic constituents of sediment including, calcite (between 800 - 850°C) and dolomite (700 - 750°C) (Santisteban et al., 2004). The percent LOI from lake sediment can vary widely across spatial and temporal scales due to multiple factors operating within the lake itself and in the watershed surrounding the lake (Shuman, 2003). For example, a disturbance such as a forest fire may change the organic matter content within the sediment. After a fire, inorganic soil may be exposed, eroded and then deposited into the lacustrine environment. This will likely be reflected in the LOI record as a decrease in LOI as inorganic material is washed-in from the watershed. Rowan et al. (1992) suggest that inorganic inputs often outweigh changes in lake trophic status, therefore short-term changes in the terrestrial environment such as forest fires should leave a signal in the LOI record.

LOI is considered a fast and inexpensive means of determining carbonate and organic matter content in sediments, while not compromising precision and accuracy (Dean, 1974; Heiri et al., 2001; Beaudoin, 2003). Sediment from the lake sediment cores used in this study were examined for their organic and carbonate content using the LOI procedures outlined by Heiri et al. (2001). Sediment samples of 1 cm^3 were taken from YK-1 and YK-3 at 1 cm intervals and at 2 cm intervals for the YK-5 core. The samples were weighed and then oven-dried for 24 hours at 105°C to remove residual moisture. Samples were then heated to 550°C for four hours in a muffle furnace to combust the organic component. They were then cooled in a desiccator and reweighed. The difference between the oven dried weight and the combusted weight at 550°C is the percentage loss on ignition and is directly related to the organic carbon content of the sediment (Heiri et al., 2001). The LOI_{550} was calculated using equation 1:

$$LOI_{550} = ((DW_{105}-DW_{550})/DW_{105})*100 \qquad (1)$$

Where LOI_{550} represents LOI at 550°C (as a percentage), DW_{105} represents the dry weight of the sample before combustion and DW_{550} the dry weight of the sample after heating to 550°C (both in mg) (Heiri et al., 2001).

The samples were returned to the muffle furnace and further heated to 950°C for another 2 hours to combust carbonate minerals into CO_2. Following combustion, they were once again cooled in a dessicator and reweighed. The difference between the 550°C weight and 950°C weight is closely related to the carbonate content of the sediment. Carbonate content was determined using equation 2:

$$LOI_{950} = ((DW_{550}-DW_{950})/DW_{105})*100 \quad (2)$$

Where LOI_{950} is the LOI at 950°C (as a percentage), DW_{550} is the dry weight of the sample after combustion at 550°C, DW_{950} represents the dry weight of the sample after heating to 950°C (all in mg) (Heiri et al., 2001).

3.9.3 Charcoal Analysis

The incomplete combustion of organic matter results in the production of particles of various size classes being dispersed from forest fires. Particles greater than 150 μm are thought to represent local fires within the catchment of the lake basin (Ohlson and Tryterud, 2000; Carcaillet et al., 2001; Whitlock and Larsen, 2001), while smaller particles are likely background material transported through the atmosphere providing a more regional interpretation of fire activity (Clark and Royall, 1995; Carcaillet et al, 2001; Whitlock and Larsen, 2001). This study will focus on the analysis of macroscopic charcoal particles greater than 150 μm in size.

Where possible, sediment samples of 1 cm^3 were taken from continguous 1 cm intervals of all three surface-sediment cores. Because material had been removed for other analyses from the surface-sediment cores (i.e., pollen, diatoms), sufficient sediment was not available for each interval for the charcoal analysis. Therefore, the charcoal analyses do not extend to the sediment-water interface for each lake. Where available, the sediment samples were carefully mixed with water and mild detergent for 24 hours to disaggregate the matrix. Samples were then sieved gently through a 150 μm sieve. The particles collected on the sieve were placed onto a grided Petri ™ dish with a square grid placed on the bottom and examined using a Nikon SMZ-645 dissecting microscope. Each grid square on the Petri dish measured 0.5 mm x 0.5 mm, giving each square an area of $25.0 \times 10^4 \mu m^2$ or $0.25\ mm^2$. Macroscopic charcoal was identified and entered into size classes of $<6.25 \times 10^4 \mu m^2$, $6.25\text{-}12.5 \times 10^4 \mu m^2$, $12.5\text{-}25.0 \times 10^4 \mu m^2$, and $>25.0 \times 10^4 \mu m^2$. The number of charcoal particles/cm^3 in all the size classes was added together to determine the total number of particles (charcoal concentration) for each sample.

Charcoal counting of the Livingston piston cores was also performed using the same methodologies. YK-1 long-core was sampled at 5 cm intervals between 20 - 140 cm and at 3 cm intervals between 140 - 350 cm. The YK-3 core was sampled at contiguous 1 cm intervals throughout the length of the core, and YK-5 was sampled at 2 cm intervals. Higher resolution analysis was carried out in sections of the core where charcoal particles were more abundant.

3.9.4 CHAPS Charcoal Analysis Programs

Because the charcoal preserved in lake sediment is comprised of particles of different size classes, often representing local or regional fires, the charcoal particle fraction of the sediment needs to be separated into peaks representing local fires and an average or background component that changes more slowly (Long et al., 1998). The charcoal data from all three cores was analyzed using the Charcoal Analysis Programs (CHAPS) written by Patrick Bartlein from the Department of Geography, University of Oregon. CHAPS is designed to statistically decompose an individual charcoal-influx time series into separate series of components. The first is a low-frequency or slowly varying background component. The second is a higher-frequency, or more rapidly varying component, described as the peaks component. The data analytical phase of fire reconstruction consists of separating that component of a time series that indicates the occurrence of a fire (peak) from that related to the joint effects of sedimentation and charcoal production (background).

CHAPS consists of two parts: CHAR and DECOMPOS. The first program is CHAR, which reads in the level (m), age (years), charcoal concentration (# charcoal particles/cm^3), charcoal influx (# charcoal particles/cm^2/yr) and deposition time (years/cm) of each sample (Long et al., 1998). Concentrations are interpolated to annual values and then averaged over the specified interval to produce influx values at even age intervals (Long et al., 1998). In this study we assigned a 10-year averaging interval for charcoal accumulation rates (CHAR), which resulted in a series of average charcoal influx values interpolated for every 10 years during the period covered by the lake sediment record. This information was then input into DECOMPOS, which is designed to calculate a weighted mean, standard deviation and autocorrelation of charcoal influx data (Long et al., 1998). The output file from DECOMPOS provides information on the average charcoal influx values throughout the core and distinguishes peaks from the slower changing background levels (Long et al., 1998). Within DECOMPOS there are two significant parameters that can be adjusted by the analyst:

1. Locally weighted mean window width for calculating background
2. Peak ratio (threshold for identifying peaks)

The locally weighted mean window width controls smoothness of the resulting background component. Windows that are two wide result in a background component which does not reflect the true long-term variations in charcoal production, while windows that are too narrow produce a background component which follows the peaks component too closely (Long et al., 1998). An appropriate window width can be determined by visually inspecting the resulting background component with the CHAR time series. In this study, because information from each of the sites concerning fires during most of the 20th century was not available, the window width parameter was determined by comparing the results of YK-1 run with various window widths and then selecting the window which produced a fire return interval similar to those cited in the literature for boreal regions of northern Canada. Previous studies have shown that fire frequency or fire return intervals during the last 1000 years are on the order of 80 - 400 years in the boreal forest region (MacDonald, 2003; McCoy and Burn, 2005). The window width of 100 was thus selected and used for the analysis of all three cores.

The peak ratio or threshold identifies all CHAR values greater than the background level, which indicate that a fire event has occurred. The value is set or calibrated using dendrochronological or historical data to specify particular values of the peaks component that when exceeded, indicate a fire event has occurred (Long et al., 1998). This study chose a value of 1.10 for the threshold parameter, so influx values greater than 1.1 times the background level were identified as peaks.

Two other parameters that are chosen by the analyst in DECOMPOS are the window width for calculating locally weighted peak frequency and the base period (in years) for reporting peak frequency. A value of 200 was chosen for the window width for calculating locally weighted peak frequency and a standard value of 1000 was chosen for the base period.

CHAPTER FOUR

4.0 RESULTS

4.1 Radiocarbon Dating

Conventional ^{14}C dates on all three sediment cores are provided in Table 4.1. The radiocarbon dates for each core are also plotted on an age vs. depth profile in Figures 4.1 (YK-1), 4.4 (YK-3) and 4.7 (YK-5). One radiocarbon date at depth 300 - 301 cm from YK-1 was considered to be anomalous and was rejected. This date had the same value as the radiocarbon date at depth 240 - 241 cm, which would result in 60 cm of deposition in a single period. However, no evidence was found for a sudden influx of sediment between 240 and 300 cm after a visual examination of the core.

Linear interpolation between the radiocarbon dates provides the sedimentation rate for each core. The sedimentation rate for YK-1 was found to be 0.043 cm/yr for the last 7552 yrs BP; YK-3 was found to be 0.01 cm/yr for the last 12158 yrs BP; and YK-5 was found to be 0.02 cm/yr for the last 9330 yrs BP.

Table 4.1 Radiocarbon dates for YK-1, YK-3 and YK-5.

Lab code	Sediment interval (cm)	Material	Conventional radiocarbon age (yrs BP)
YK-1			
	65-66	White River ash	1 250*
TO-11065	75-76	bulk sediment	1 880 ± 50
TO-11066	150-151	bulk sediment	3 260 ± 50
TO-11067	240-241	bulk sediment	5 220 ± 60
TO-11068	300-301	bulk sediment	5 220 ± 60**
TO-11069	350-351	bulk sediment	8 330 ± 70
YK-3			
	20-21	White River ash	1 250*
AA39615	57-58	bulk sediment	5 979 ± 44
AA39616	85-86	bulk sediment	8 202 ± 49
AA39618	115-116	bulk sediment	9 773 ± 54
AA39619	130-131	bulk sediment	11 164 ± 71
YK-5			
	26-26	White River ash	1 250*
AA39614	38-39	bulk sediment	1 436 ± 36
AA39617	90-91	bulk sediment	3 240 ± 37
AA39620	164-165	bulk sediment	6 512 ± 45
AA39621	227-228	bulk sediment	9 746 ± 54

TO – IsoTrace
AA – University of Arizona
*** Fuller and Jackson (2002)**
**** rejected**

4.2 YK-1

4.2.1 Sediment Stratigraphy

The base of YK-1 (379.5 cm) up to 374 cm is a mixture of light brown-olive colored organic sediment and clay (Fig. 4.1). From 373 cm to 305 cm, the sediment stratigraphy is predominantly gyttja with numerous layers of aquatic moss interspersed. Moss layers occurred at 347 - 346.5 cm, 343 - 342.5 cm, 340.5 - 338.5 cm and 336 - 334 cm. From 305 cm to the top of the core (5 cm) the sediment is predominantly composed of gyttja. There is a visible change in the colour from brown gyttja to dark brown gyttja at 205 cm. From 205 cm to the top of the core the sediment is organic rich gyttja except for a 1 cm thick layer of ash at 66.5 cm. This represents the White River Ash (Fuller and Jackson, 2002).

4.2.2 Loss-on-ignition

LOI_{550} analysis reveals that the organic content of the YK-1 core remains relatively unchanged throughout the length of the record (Fig.5.1). There is a general decreasing trend in LOI_{550} from ~335 cm (82% LOI) to the top of the core (~70%). Very low organic content occurs from the base of the core to 335 cm and again at approximately 65 cm depth. At the base of the core of the core, LOI is less than 10% while at 65 cm, LOI decreases from ~78% near 70 cm to ~24% at 65 cm depth.

LOI_{950} analysis shows a more variable trend in carbonate content for the YK-1 core (Fig. 4.1). There is a general increase in carbonate content throughout the record, except for a marked decrease at the base of the core. The carbonate content decreases at 363 cm (~2.25%) to about ~334 cm (0%). The carbonate content then generally increases to a value of ~2% at the top of the core. There is a decrease between ~150 cm to 120 cm, with carbonate values falling from ~2.0% to 0.2% before increasing once again. An overall mean value of 1.26% was calculated for the entire core.

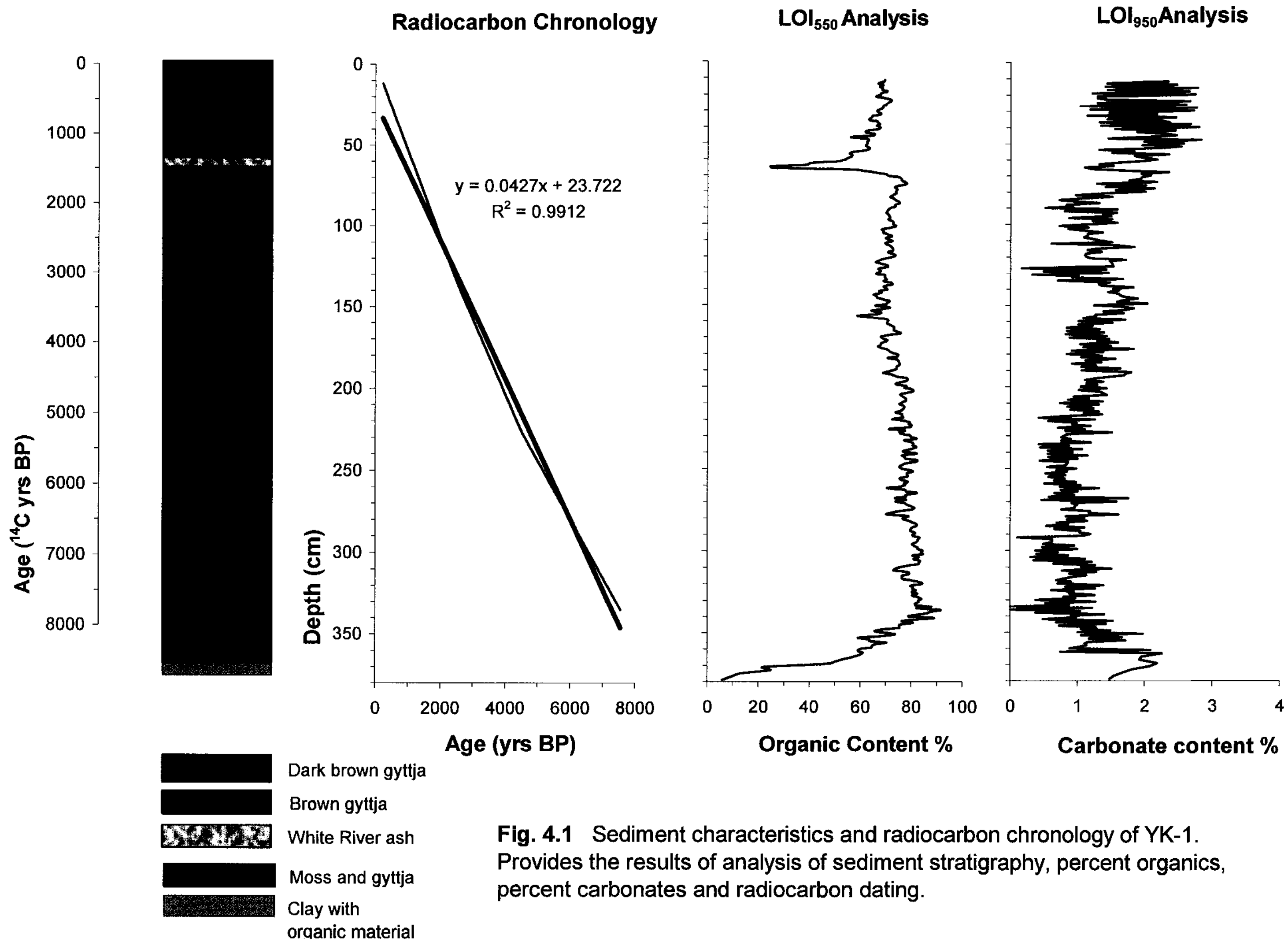


Fig. 4.1 Sediment characteristics and radiocarbon chronology of YK-1. Provides the results of analysis of sediment stratigraphy, percent organics, percent carbonates and radiocarbon dating.

4.2.3 Charcoal

Inspection of charcoal concentrations from YK-1 shows marked differences in the fire regime during the Holocene. The charcoal record from YK-1 is divided into three distinct sections: early Holocene (up to ~7000 ^{14}C yrs BP), mid Holocene (~7000 to 4000 ^{14}C yrs BP) and late Holocene (from ~4000 ^{14}C yrs BP to the present). These zones are based on previously published paleodata (Hebda, 1995) and reflect major changes in climate and vegetation.

Early Holocene Zone ~8740 - 7015 ^{14}C yrs BP (377 cm to 316 cm)

The basal section of the core is represented by increasing concentrations of charcoal (Fig. 4.2). Relatively low values of charcoal accumulation (average of 12 particles/cm^3) occur in the earliest sediments from this section between 377 cm to 342 cm (estimated ~8740 - 7750 ^{14}C yrs BP). Radiocarbon dates are uncertain at this depth and are only estimates based on the assumption that deposition remained constant to the bottom of the core. Beginning at 341 cm (estimated ~7722 ^{14}C yrs BP) charcoal concentration begins to increase and fluctuate rapidly for the remainder of this zone. A large peak occurs at 341 cm (estimated ~7722 ^{14}C yrs BP) with a value of 48 particles/cm^3. At depth 331 cm (~7440 ^{14}C yrs), charcoal values begin to fluctuate rapidly and peak at 322 cm (~7200 ^{14}C yrs BP) with 93 particles/cm^3. This peak is not only the largest in the section, but is also the largest of the entire core. The second largest peak of the core is also found in this section, at a depth of 328 cm and a value of 66 particles/cm^3. Concentration of charcoal particles decreases after the large peak at 322 cm, reaching a low value of ~6 particles/cm^3 near the top of the early Holocene zone at depth 316 cm (7015 ^{14}C yrs). The average charcoal concentration for this zone is 19 particles/cm^3.

Analysis of the CHAPS data shows that the average charcoal influx of this section is 0.497 particles/cm^2/yr (Fig. 4.3). The CHAR values are low at the beginning of this section until about ~7740 ^{14}C yrs BP (estimated), after which the CHAR values fluctuate rapidly for the remainder of the section and show a large range from 3.07 particles/cm^2/yr to 0.04 particles/cm^2/yr. The average fire frequency in the Early Holocene zone is 4.3 fires/1000 yrs. Fire frequency in this lowermost zone ranges from 2.2 to 5.9 fires/1000 yrs. Eight fire peaks are identified by the CHAPS program in the Early Holocene zone.

Mid Holocene Zone ~7015 - 3950 ^{14}C yrs BP (316 cm - 202 cm)

The mid Holocene charcoal zone is characterized by more consistent charcoal concentration values, except in the middle of the zone which experiences higher charcoal concentration and more fluctuations (Fig. 4.2). The mid Holocene zone extends from 316 to 202 cm and covers the period from approximately 7015 - 3950 ^{14}C yrs BP. Charcoal concentration remains relatively stable from 316 cm to 262 cm with an average charcoal concentration of 15 particles/cm^3. From 262 cm to 232 cm, charcoal values increase and experience more variability, with an average charcoal concentration of 25 particles/cm^3. The two largest peaks in the mid Holocene zone with values of 46 and 43 particles/cm^3 occur at depths 262 cm (~5488 ^{14}C yrs BP) and 259 cm (~5403 ^{14}C yrs BP), respectively. From 232 cm to the top of this zone, charcoal concentration decreases and becomes less variable once again. The average charcoal concentration of the mid Holocene zone is 19 particles/cm^3.

There is an increase in CHAR in the mid Holocene charcoal zone relative to the high values that characterized the early Holocene zone (Fig. 4.3). The average CHAR is 0.640 particles/cm^2/yr, with a maximum CHAR of 1.48 particles/cm^2/yr and a minimum of 0.252 particles/cm^2/yr. The CHAR values remain relatively low and consistent at the beginning of this zone until about ~5510 ^{14}C yrs BP, at which point they increase and begin to fluctuate more rapidly for the remainder of this zone. The average fire frequency also decreases to 3.1 fires/1000 yrs in the mid Holocene zone. The minimum fire frequency for this zone is 1.2 fires/1000 yrs and the maximum fire frequency is ~4.4 fires/1000 yrs. There is a total of 9 fire peaks recorded in the mid Holocene zone, with an absence of fire peaks between ~ 6810 to 5770 ^{14}C yrs BP.

Late Holocene Zone ~3950 - 225 ^{14}C yrs BP (202 cm to 12 cm)

The late Holocene charcoal zone extends from 202 - 12 cm and corresponds to the time period from 3950 - 225 ^{14}C yrs BP (Fig. 4.2). The average charcoal concentration in this zone is 18 particles/cm^3. Charcoal concentration within the late Holocene zone can be divided into two distinct periods; the first period characterized by highly fluctuating charcoal concentration values and the second period characterized by lower charcoal concentration values which remain relatively stable.

There is an initial decrease in charcoal concentration at the beginning of the zone which is followed by a period of fluctuating charcoal values and several relatively large peaks from 183 cm to 141 cm or between ~3541 - 2626 ^{14}C yrs BP. The average charcoal concentration between 183 - 141 cm is 23 particles/cm^3 with peaks ranging from 32 - 44 particles/cm^3. The highest peak in the late Holocene zone occurs during this highly variable period at depth 169 cm (~3236 ^{14}C yrs BP) with a value of 53 particles/cm^3.

There is a large drop in charcoal concentration from 37 particles to 2 particles/cm^3 between 141 - 132 cm or ~2626 to 2447 ^{14}C yrs BP. The charcoal concentration remains relatively low and unvarying from 132 cm to the top of the core, with only 13 particles being recovered in each cm^3 of sample on average. A smaller peak occurs at 62 cm (~1157 ^{14}C yrs BP) with a value of 27 particles/cm^3.

The CHAR for the late Holocene zone has very similar trends to the charcoal concentrations discussed above (Fig. 4.3). CHAR increases near the beginning of this zone at ~3730 ^{14}C yrs BP and experiences large fluctuations until about ~2420 ^{14}C yrs BP. The largest peak in the zone occurs at ~3300 yrs BP with a value of 1.56 particles/cm^2/yr. From ~2420 yrs BP to the top of the core, the CHAR values remain relatively low and are less variable. Average CHAR for the Late Holocene zone is 0.788 particles/cm^2/yr. Fire frequency is somewhat higher than the previous zone but is below the levels recorded in the early Holocene. The average fire frequency is 3.6 fires/1000 yrs during the late Holocene. The maximum fire frequency recorded in this zone is 5.8 fires/1000 yrs and the minimum fire frequency value is 2.1 fires/1000 yrs. A total of 13 fire peaks are distinguished from the background charcoal values in this section.

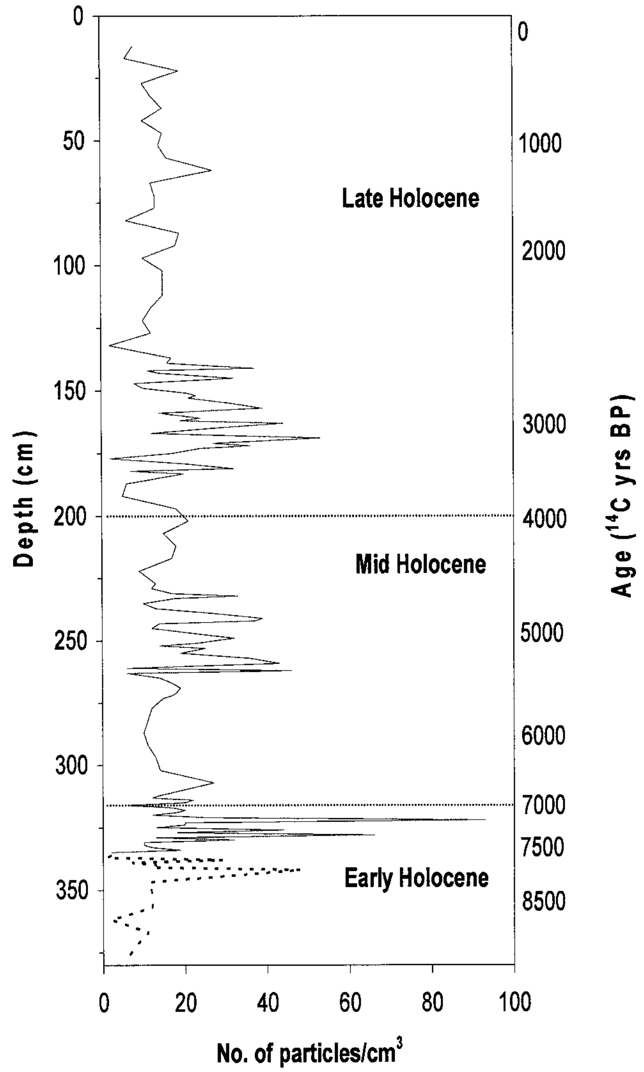


Fig. 4.2 Charcoal concentration of YK-1. Fig.4.2 shows three zones based on the major changes in climate and vegetation as reflected in previously published studies. The dashed line represents charcoal concentration data past the last radiocarbon date and age is estimated assuming deposition remained constant to the bottom of the core.

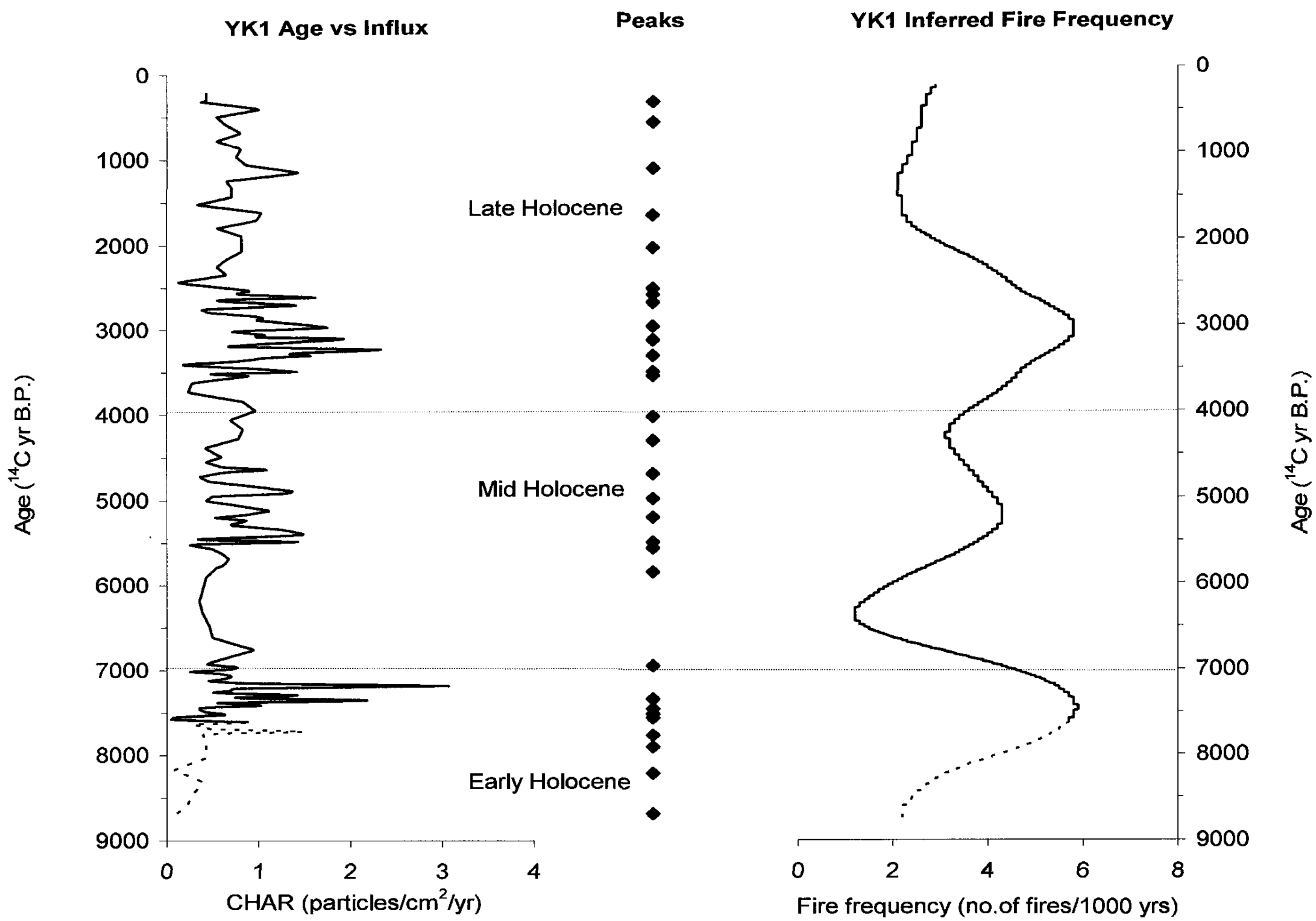


Fig. 4.3. CHAPS Analysis of YK-1. Three zones are distinguished based on the major changes in climate and vegetation as reflected in previously published studies. The dashed line represents CHAR data past the last radiocarbon date and age is estimated assuming deposition remained constant to the bottom of the core.

4.3 YK-5

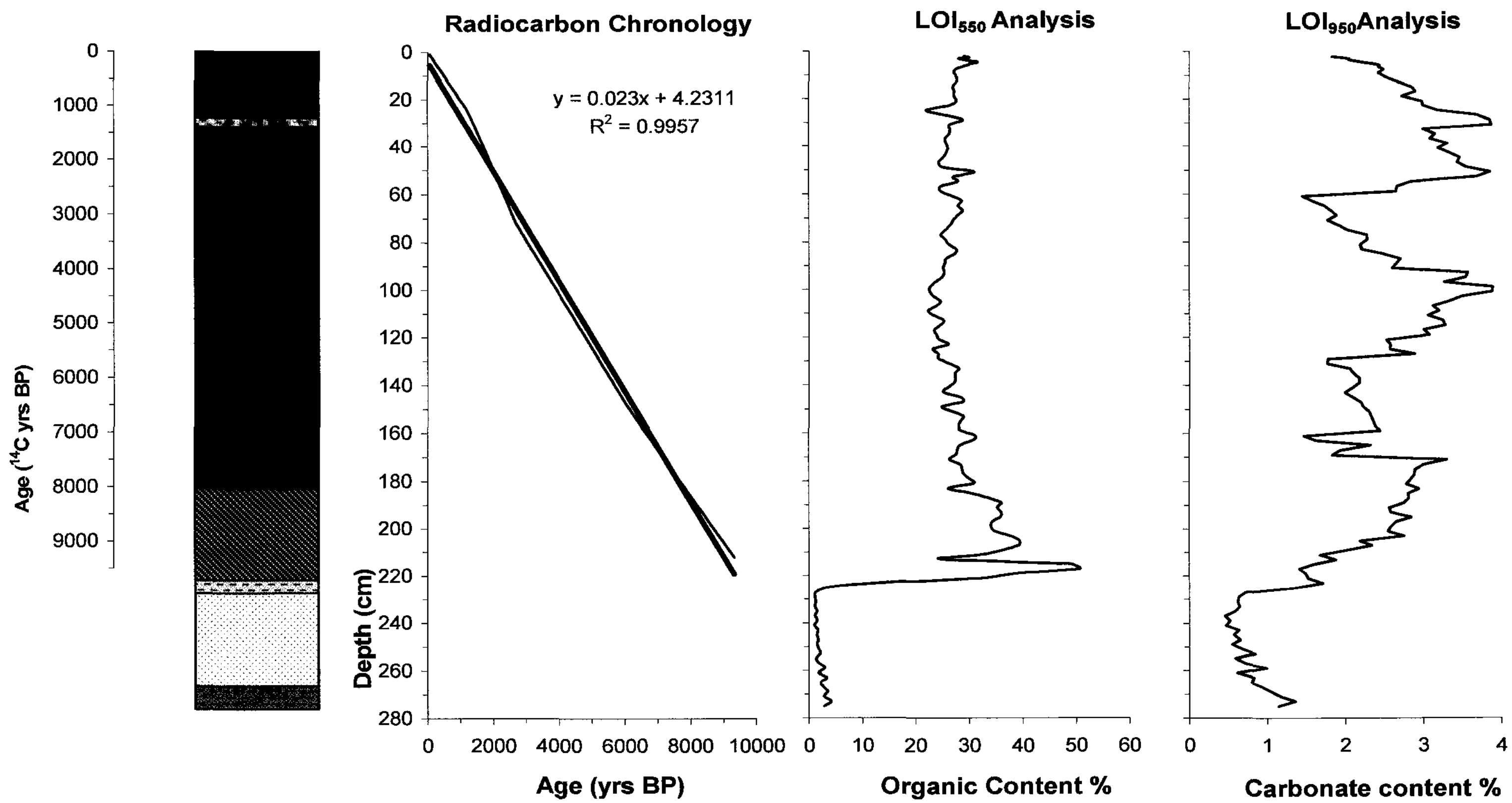
4.3.1 Sediment Stratigraphy

The bottom 10 cm of YK-5 (277 - 267 cm) is composed of clay-rich sediment (Fig. 4.4). From 267 cm to 228 cm, the sediment becomes an olive-grey mixture of sand and clay. Between 228 - 225 cm the sediment is still clay-rich but olive green in colour. A 1 cm layer of organic gyttja occurs between 225 - 224 cm. From 184 cm to the top of the core, the sediment is mostly gyttja and becomes dark brown in colour starting at 84 cm. The White River ash is found in the gyttja at a depth of 25 cm.

4.3.2 Loss-on-ignition

Throughout the length of the YK-5 record, LOI_{550} decreases from 45.3% at 216 cm to 29% at the top of the core (Fig. 4.4). The organic content at the base (275 cm) of YK-5 is initially very low (under 10%) but then increases quickly at depth 223 cm to a maximum of 51% at 217 cm. LOI_{550} then decreases to 22% at 110 cm, but subsequently increases slightly for the remainder of the core to a value of 29% at the top of the core.

LOI_{950} records a more variable trend in carbonate content for YK-5. At the base of YK-5 (depth 275 cm) the carbonate content is relatively low, with a value of 1.1%. Percentage of carbonates slightly decreases initially to 0.46% at 237 cm before increasing to 3.3% at 171 cm. There is a subsequent decrease in LOI_{950} to 1.5% at 161 cm. Carbonate content then increases to a maximum of 3.9% at 99 cm, and once again decreases to 1.5% at 61 cm. A relatively rapid rise in carbonates occurs between 61 and 51 cm, reaching 3.85% at 51cm. Despite a small peak (3.87%) at 31 cm, carbonate content decreases steadily for the remainder of the core (last 50 cm) and decreases to 1.8% at ~ 2 cm.



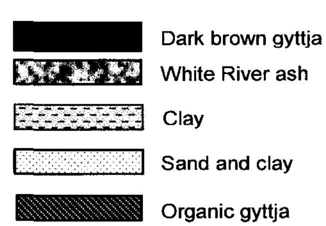


Fig. 4.4. Sediment characteristics and radiocarbon chronology of YK-5. Provides the results of analysis of sediment stratigraphy, percent organics, percent carbonates and radiocarbon dating.

4.3.3 Charcoal

Examination of charcoal concentrations from YK-5 reveals significant differences in the fire regime during the Holocene. The charcoal record from YK-5 is divided into three distinct sections which are based on results of previously published studies and reflect major changes in climate and vegetation: early Holocene (up to ~7000 ^{14}C yrs BP), mid Holocene (~7000 to 4000 ^{14}C yrs BP) and late Holocene (from ~4000 ^{14}C yrs BP to the present).

Early Holocene Zone ~12,513 - 7021 ^{14}C yrs BP (274 cm to 167 cm)

The bottom section of the core is represented by relatively low concentrations of charcoal (Fig. 4.5). The average charcoal concentration for the early Holocene zone is 2.31 particles/cm^3, which is the lowest charcoal concentration value for all three sections. Charcoal particles are absent in the basal sediment between 274 to 220 cm (12,513 - 9741 ^{14}C yrs BP. Radiocarbon dates are uncertain between 274 cm (~12,513 ^{14}C yrs BP) and 214 cm (~9433 ^{14}C yrs BP) and are only estimates based on the assumption that deposition remained constant to the bottom of the core. Beginning at 220 cm (9741 ^{14}C yrs BP), charcoal concentration begins to increase and reaches a maximum of 10 particles/cm^2 at 184 cm (7893 ^{14}C yrs BP). Charcoal values tend to fluctuate slightly in the early Holocene zone.

Analysis of the CHAPS data shows that the average charcoal influx of the early Holocene zone is 0.05 particles/cm^2/yr and is considered low compared to the other sections of the core (Fig. 4.6). The largest peak of the early Holocene zone has a value of 0.19 particles/cm^2/yr and occurs at 184 cm or 7889 ^{14}C yrs BP. Two smaller peaks also occur at 188 cm (8099 ^{14}C yrs BP) with a value of 0.131 particles/cm^2/yr and 168 cm

(7069 ^{14}C yrs BP) with a value of 0.133 particles/cm^2/yr. The average fire frequency of the early Holocene zone is 2.25 fires/1000 years with a high of 2.6 fires/1000 years at 8389 ^{14}C yrs BP and a low of 1.8 fires/1000 years at the bottom of YK-5. A total of 5 fire peaks were recorded in the Early Holocene zone.

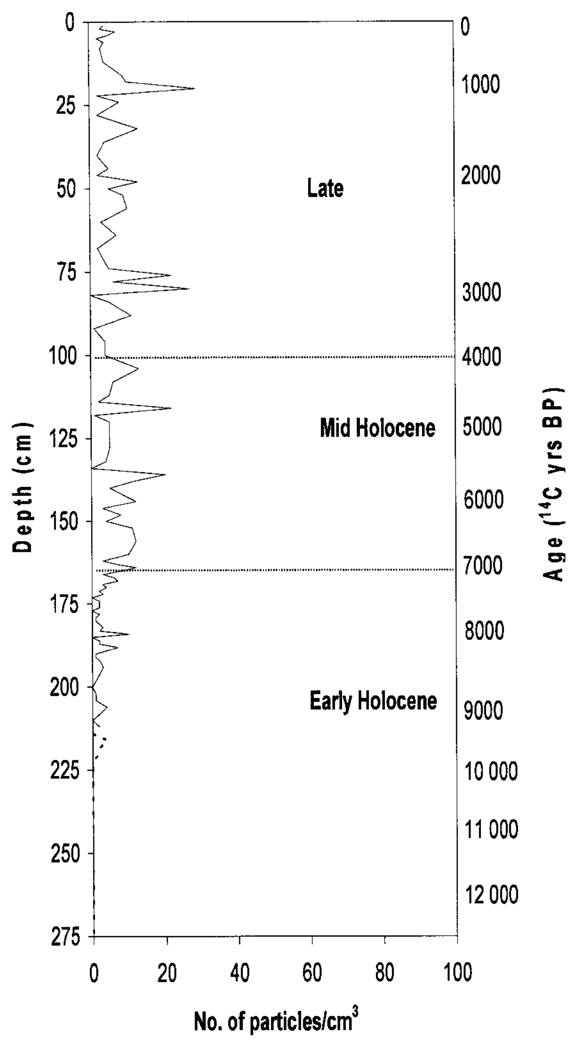


Fig. 4.5. Charcoal concentration of YK-5. Fig 4.5 illustrates the three zones representing the major changes in charcoal deposition and changes in the fire regime throughout the Holocene. The dashed line represents charcoal concentration data past the last radiocarbon date and age is estimated assuming deposition remained constant to the bottom of the core.

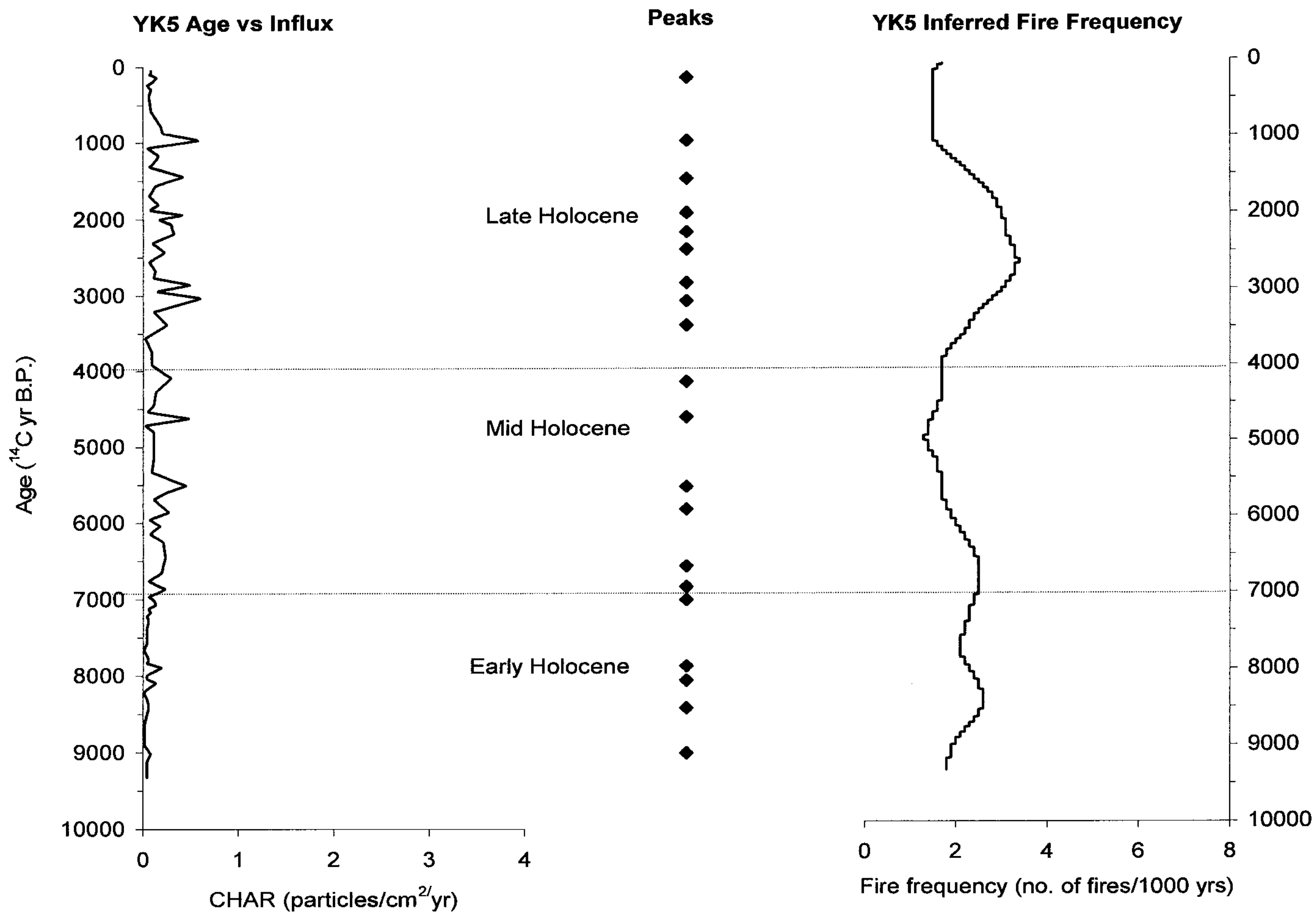


Fig. 4.6. CHAPS Analysis of YK-5. Three zones are distinguished based on the major changes in climate and vegetation as reflected in previously published studies.

Mid Holocene zone ~7020 - 3923 ^{14}C yrs BP (167 cm to 100 cm)

The mid Holocene charcoal zone is characterized by a significant increase of macroscopic charcoal (Fig. 4.5). This zone extends from 167 to 100 cm and covers the period from approximately 7020 - 3923 ^{14}C yrs BP. The average charcoal concentration of this section is 7.39 particles/cm^3 which is significantly higher than the early Holocene zone. Charcoal values are highly variable in this zone, with a minimum value of 0 particles/cm^3 and a maximum value of 22 particles/cm^3 at 134 cm and 116 cm, respectively. The second largest charcoal concentration peak in this zone has a value of 20 particles/cm^3 at 136 cm (5515 ^{14}C yrs BP). Five moderate-sized charcoal concentration peaks also occur in the mid Holocene zone which have charcoal concentrations between 12 - 13 particles/cm^3 and are found at 164, 156, 144 and 104 cm. Unlike the previous zone, the mid Holocene zone exhibits more variability with respect to the concentration of charcoal contained in the sediments.

There is a noticeable increase in CHAR in the mid Holocene charcoal zone relative to the low values that typified the early Holocene (Fig. 4.6). The average CHAR is 0.17 particles/cm^2/yr, with a maximum CHAR of 0.478 particles/cm^2/yr and a minimum of 0.028 particles/cm^2/yr. There are two large peaks notable in this section at depths of 116 cm (4629 ^{14}C yrs BP) (0.478 particles/cm^2/yr) and 136 cm (5509 ^{14}C yrs BP) (0.447 particles/cm^2/yr). The fire frequency is relatively high at the beginning of the mid Holocene, with a value of 2.5 fires/1000 yrs, but then decreases to a value of 1.3 fires/1000 yrs by the middle of the zone. The fire frequency increases again to a value of 2.2 fires/1000 yrs by the end of the mid Holocene at 100 cm (3923 ^{14}C yrs BP). Overall this section has an average fire frequency of only 1.7 fires/1000 yrs, which is surprisingly lower than the average fire frequency for the previous section. A total of 6 fire peaks are identified by the CHAPS program in the mid Holocene zone.

Late Holocene Zone ~3923 - 49 ^{14}C yrs BP (82 cm to 1 cm)

The late Holocene zone spans an 81 cm depth between 82 - 1 cm and represents the last 4000 years of deposition (Fig. 4.5). This zone is comparable to the mid Holocene zone, with large fluctuations in the charcoal concentration values. The average charcoal concentration in the late Holocene zone is 6.81 particles/cm^3. The maximum charcoal concentration is 29 particles/cm^3, at 20 cm (980 ^{14}C yrs BP). Two other large peaks are notable at 80 cm (3038 ^{14}C yrs BP) (27 particles/cm^3) and 76 cm (2861 ^{14}C yrs BP) (22 particles/cm^3). Three moderate peaks are also identified in the late Holocene zone, with values ranging between 13 - 10 particles/cm^3 at depths of 32 cm (1451 ^{14}C yrs BP), 48 cm (1944 ^{14}C yrs BP), 56 cm (2191 ^{14}C yrs BP). The minimum charcoal concentration in the late Holocene zone varies between 0 - 2 particles/cm^3 and occurs at various depths within the zone.

CHAR remains quite similar to the mid Holocene zone, with an average CHAR value of 0.18 particles/cm^2/yr (Fig. 4.6). Greater fluctuations in the CHAR values are evident compared to the mid Holocene charcoal zone. The two largest peaks in this section have values of 0.572 particles/cm^2/yr and 0.596 particles/cm^2/yr and correspond to the years of 3039 ^{14}C yrs BP and 979 ^{14}C yrs BP, respectively. Three moderate-sized peaks in the late Holocene zone are also notable, with values of 0.413 particles/cm^2/yr (1 449 ^{14}C yrs BP), 0.408 particles/cm^2/yr (1939 ^{14}C yrs BP) and 0.485 particles/cm^2/yr (2859 ^{14}C yrs BP). The lowest CHAR value in this zone is 0.04 particles/cm^2/yr at ~5 cm (239 ^{14}C yrs BP). The average fire frequency is 2.33 fires/1000 yrs, and is the highest average fire frequency value among all three zones. The fire frequency peaks at 2629 yrs BP with a value of 3.4 fires/1000 yrs, and then drops to only 1.5 fires/1000 yrs at 22 cm (1069 ^{14}C yrs BP). A total of 9 fire peaks are identified as fire events in the late Holocene zone.

4.4 YK-3

4.4.1 Sediment Stratigraphy

From the base of YK-3 (149 cm) up to 146 cm sediment is composed of a mixture of clay and sand (Fig. 4.7). Clay rich sediment, which changes from olive green to grey in colour occurs between 146 and 138.5 cm. Dark brown gyttja occurs from 138.5 to 107 cm, but is not found to be very organic in composition. The remainder of the core (107 - 0 cm) was composed of dark brown gyttja. The White River ash is represented by a thin layer (0.5 cm) at 20 cm depth.

4.4.2 Loss-on-ignition

Throughout the length of the YK-3 record, LOI_{550} decreases from ~26% at 135 cm to less than 10% at 20 - 25 cm (Fig. 4.7). From ~20 cm to the top of the core, LOI_{550} values increase slightly to approximately 12 - 13% at the top of the core. An overall mean value of 12.3% was calculated for the entire core.

LOI_{950} analysis reveals a slightly fluctuating carbonate content throughout the YK-3 core. Carbonate content increases initially from 1.5% at the base of the core to 3.3% at 115 cm depth. LOI_{950} values then decrease to a value of 0.9% at 20 cm, with the exception of one relatively large peak (4.7%) at 24 cm. Carbonate content increases once again from depth 20 cm (0.9%) to the top of the core (3.7 %). An overall mean value of 2.2% was calculated for the entire core.

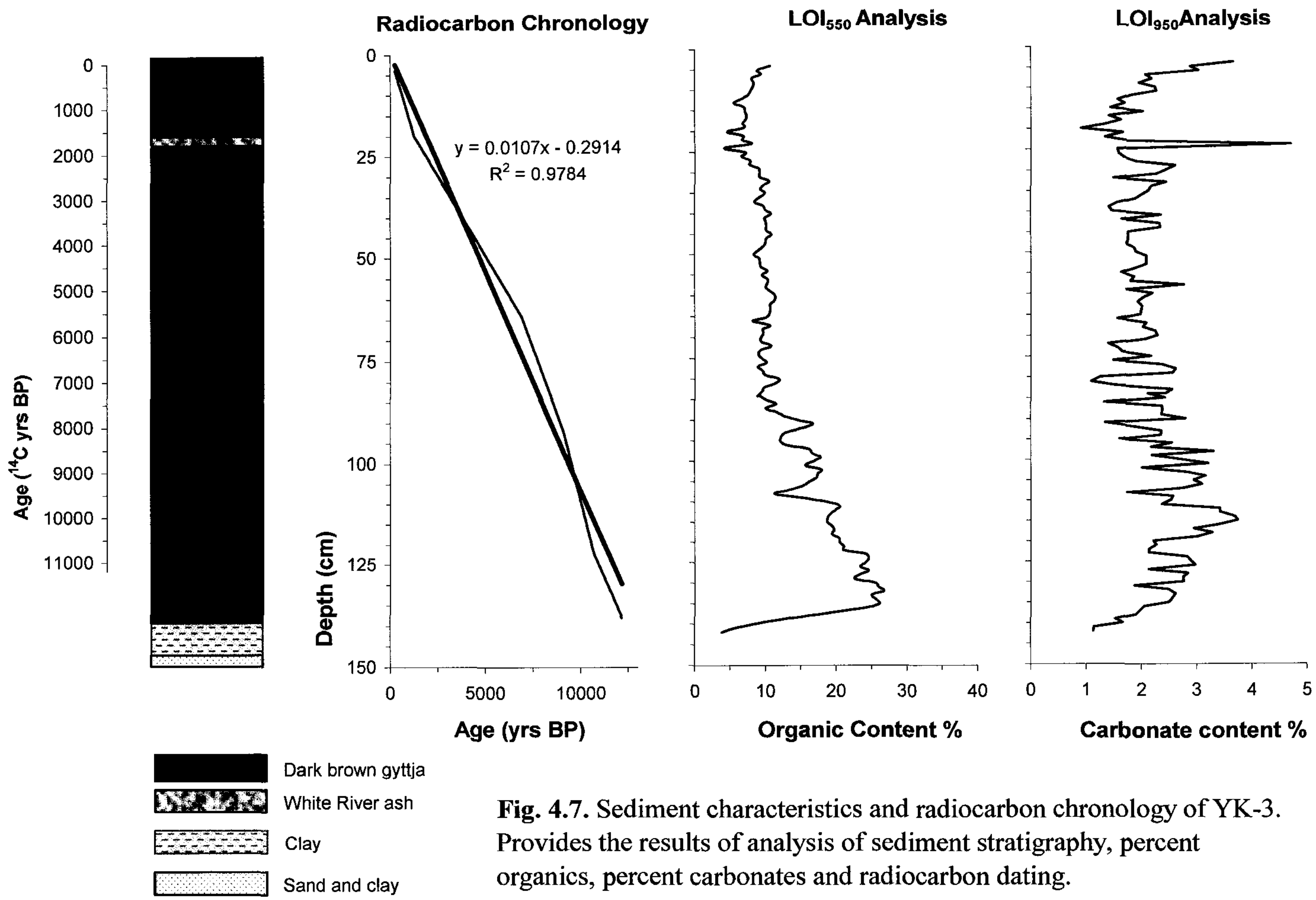


Fig. 4.7. Sediment characteristics and radiocarbon chronology of YK-3. Provides the results of analysis of sediment stratigraphy, percent organics, percent carbonates and radiocarbon dating.

4.4.3 Charcoal

The charcoal concentration from YK-3 illustrates differences in the fire regime throughout the Holocene (Fig. 4.8). The charcoal record from YK-3 is divided into three distinct sections as previously mentioned for YK-1 and YK-5: Early Holocene (up to ~7000 ^{14}C yrs BP), Mid Holocene (~7000 to 4000 ^{14}C yrs BP) and Late Holocene (from ~4000 ^{14}C yrs BP to the present).

Early Holocene Zone ~13,085 – 7033 ^{14}C yrs BP (148 cm to 66 cm)

The early Holocene zone encompasses slightly more than half of the YK-3 core, and spans a timeframe of ~13,085 to 7033 ^{14}C yrs BP, or about 6000 years (Fig. 4.8). The radiocarbon chronology is uncertain between 148 cm (~13,085 ^{14}C yrs BP) and 138 cm (~12,157 ^{14}C yrs BP) and are only estimates based on the assumption that deposition remained constant to the bottom of the core. Average charcoal concentration for this zone is 7.88 particles/cm^3, which is the highest of the three zones of YK-3. The bottom 20 cm of YK-3, from 148 to 128 cm, is characterized by the absence or minimal amounts of charcoal particles, with an average charcoal concentration of only 0.43 particles/cm^3. The sediment chronology reveals that the sediment samples at this depth are largely composed of clay and fine silt, which may explain the absence of charcoal from this section.

Charcoal concentration begins to increase significantly at 127 cm (11,138 ^{14}C yrs BP). For the next 61 cm or the remainder of the early Holocene zone, charcoal concentration experiences large fluctuations between values. The average charcoal concentration between 127 and 66 cm is 10.4 particles/cm^3. The highest peak in the early Holocene zone occurs at 98 cm (9417 ^{14}C yrs BP) and has a value of 46 particles/cm^3.

Several moderate peaks are also notable, with values of 20, 25, 23 and 21 particles/cm^3 at depths of 120 cm, 118 cm, 107 cm, 97 cm, respectively. At the top of the early Holocene zone at depth 66 cm (7032 ^{14}C yrs BP), charcoal concentration begins to decrease.

Analysis of the CHAPS data shows that the average charcoal influx in the early Holocene zone is 0.135 particles/cm^2/yr (Fig. 4.9). The highest CHAR values occur in this zone, which also has the largest variability between CHAR values of any zone of YK-3. From the bottom of the core up to about ~11,000 ^{14}C yrs BP, CHAR remains negligible, which is consistent with the absence of charcoal particles in the sediments at this time as mentioned above. Beginning at about ~11,000 ^{14}C yrs BP, CHAR values increase significantly and peak at 98 cm (9410 ^{14}C yrs BP) with a maximum value of 0.849 particles/cm^2/yr. CHAR values experience relatively large fluctuations between ~, to ~8500 ^{14}C yrs BP. Three smaller peaks are found between ~11,000 to 8500 ^{14}C yrs BP, with values of 0.465 particles/cm^2/yr (10,460 ^{14}C yrs BP), 0.421 particles/cm^2/yr (9880^{14}C yrs BP) and 0.367 particles/cm^2/yr at 8460 ^{14}C yrs BP). From about ~8500 ^{14}C yrs BP, CHAR values are relatively low and decrease to 0.076 particles/cm^2/yr by the top of the early Holocene zone at ~66 cm (~7000 ^{14}C yrs BP).

Increases in charcoal concentration and CHAR are also highlighted in the fire frequency values for the early Holocene zone. Average fire frequency in this lowermost zone is 2.4 fires/ 1000 yrs, which is the highest fire frequency value for all of YK-3. Fire frequency initially increases significantly within the zone until approximately 10100 ^{14}C yrs BP, when it reaches the maximum value of 4.1 fires/1000 yrs. Fire frequency then decreases to a value of 2.1 fires/1000 yrs and increases once again to a value of 3.6 fires/ 1000 yrs at around ~7600 ^{14}C yrs BP. The lowest fire frequency value of 0.9 fires/1000 yrs occurs at the bottom of the section at an approximate depth of 127 cm (11140 ^{14}C yrs BP). A total of 13 fire peaks are distinguished from background charcoal levels in this section.

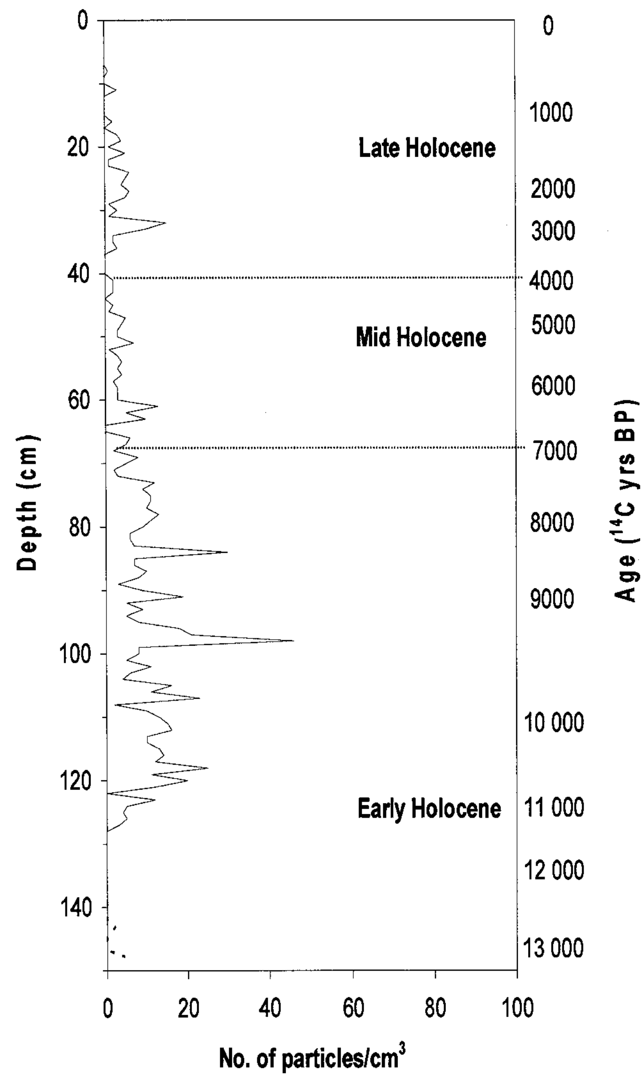


Fig. 4.8. Charcoal concentration of YK-3. Fig. 4.8 shows three zones representing the major changes in charcoal deposition and changes in the fire regime. The dashed line represents charcoal concentration data past the last radiocarbon date and age is estimated assuming deposition remained constant to the bottom of the core.

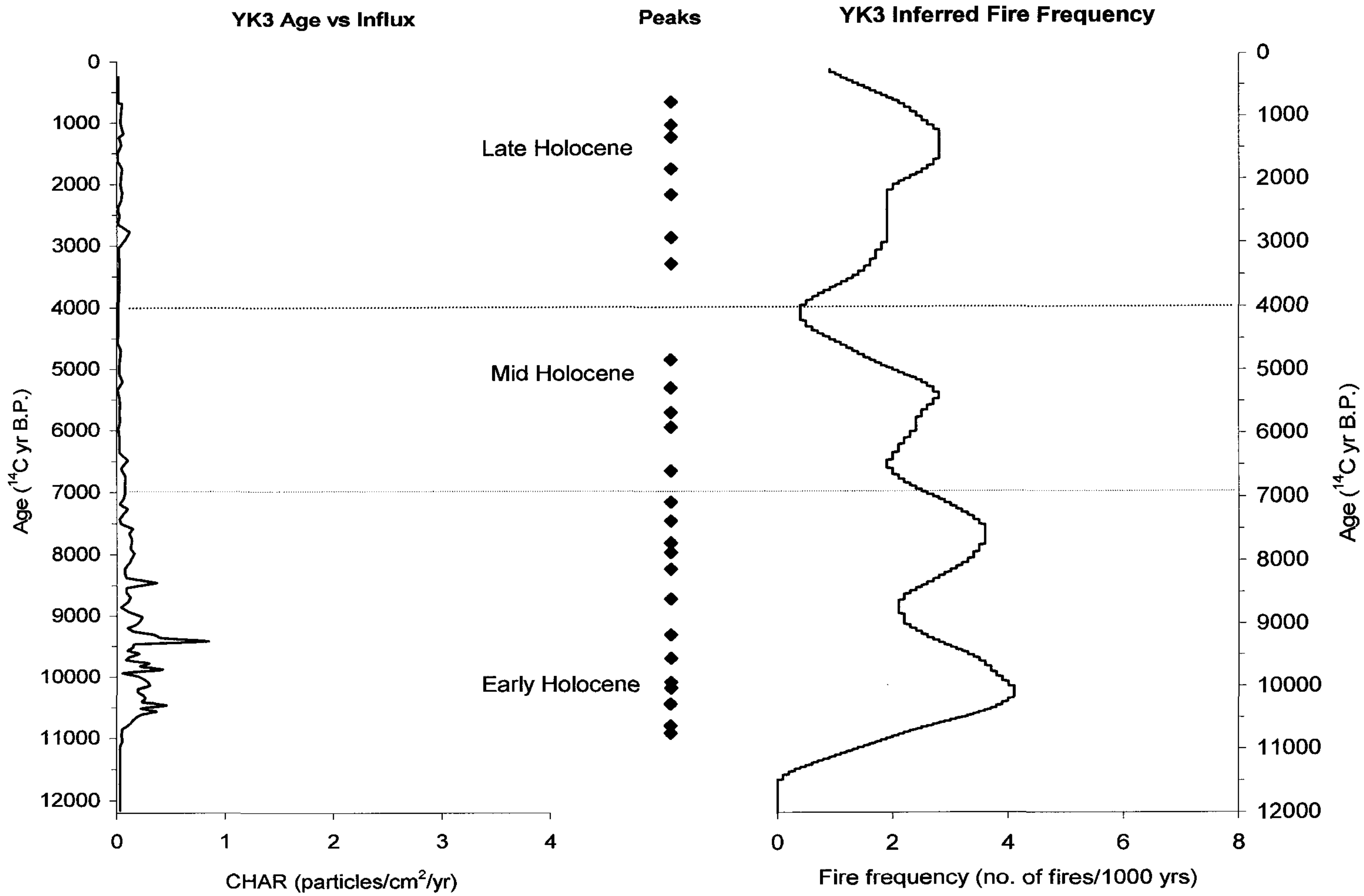


Fig. 4.9. CHAPS Analysis of YK-3. Three zones are distinguished based on the major changes in climate and vegetation as reflected in previously published studies.

Mid Holocene zone ~7032 - 4061 ^{14}C yrs BP (66 cm to 42 cm)

The mid Holocene charcoal zone of YK-3 is characterized by low concentrations of macroscopic charcoal (Fig. 4.8). This zone extends from 66 cm (7032 ^{14}C yrs BP) to 42 cm (4061 ^{14}C yrs BP), or approximately 3000 years. The average charcoal concentration of this section is 3.46 particles/cm^3. The maximum concentration of charcoal particles in the mid Holocene zone occurs at 61 cm (6490 ^{14}C yrs BP) with a value of 13 particles/cm^3 and 63 cm (6746 ^{14}C yrs BP) with a value of 10 particles/cm^3. The minimum charcoal concentration of value 0 particles/cm^3 occurs at 64 - 65 cm (6953 to 6874 ^{14}C yrs BP) and 44 cm (4317 ^{14}C yrs BP). The charcoal concentration values in the mid Holocene zone are much less variable than in the early Holocene zone.

The CHAR values in the mid Holocene zone decrease dramatically compared to those in the early Holocene zone, as the average CHAR falls to only 0.033 particles/cm^2/yr (Fig. 4.9). CHAR values are highest at the beginning of the mid Holocene zone and peak at a value of 0.1 particles/cm^2/yr at depth 61 cm (6490 ^{14}C yrs BP). CHAR values are low for the remainder of the section, and decrease to 0.016 particles/cm^2/yr by the top of the zone ~4000 ^{14}C yrs BP. Average fire frequency for the mid Holocene zone also decreases to 1.9 fires/1000 yrs, with a range of 2.8 to 0.4 fires/1000 yrs. There are five peaks identified in the mid Holocene zone.

Late Holocene Zone ~4061 - 250 ^{14}C yrs BP (42 cm to 4 cm)

The uppermost section of the core (top 42 cm) represents about ~3800 years of sediment deposition (Fig. 4.8). Charcoal concentration values within this zone are comparable to those from the previous zone. The average charcoal concentration for the late Holocene zone is 2.53 particles/cm^3. There is an absence of charcoal at the beginning of the late Holocene zone, between 40 and 31 cm (3806 to 3423 ^{14}C yrs BP). Following this quiescent period, there is a sudden increase in charcoal concentration, with a peak of 15 particles/cm^3 at 32 cm (2784 ^{14}C yrs BP). Macroscopic charcoal concentration steadily decreases throughout the remainder of the zone and becomes completely absent at the very top of the core at 7 cm (438 ^{14}C yrs BP).

CHAR in the late Holocene zone remains consistent with the mid Holocene zone, with an average CHAR value of 0.03 particles/cm^2/yr (Fig. 4.9). CHAR values remain negligible at the beginning of the late Holocene zone until a sudden increase which peaks at a maximum value of 0.116 particles/cm^2/yr at ~32 cm (~2780 ^{14}C yrs BP). CHAR values remain low for the remainder of the core, with an average value of 0.03 particles/cm^2/yr. The CHAR value drops to a minimum value of 0.16 particles/cm^2/yr for the top 11 cm of the core. The average fire frequency in the late Holocene zone is 1.9 fires/1000 yrs, with a maximum value 2.8 fires/1000 yrs at ~21 cm (1400 ^{14}C yrs BP) and a minimum value of 0.4 fires/1000 yrs at the bottom of the zone at ~40 cm (3990 ^{14}C yrs BP). Seven fire peaks are identified in this section of the sediment record.

CHAPTER FIVE

5.0 DISCUSSION

5.1 Influence of Elevation on the Fire Regime

Elevation indirectly influences the fire regime by determining the type and structure of plant communities in a given region. Properties such as slope, aspect, elevation and soil type affect site energy and water budgets, which in turn influence the growth of specific plant communities. As one moves upslope in mountainous regions, vegetation cover or above ground biomass decreases significantly. In the valley bottoms, closed canopy forests provide abundant biomass and fine fuels that help fuel wildfires. Further, the physical structure of the closed canopy forests enhances the spread of wildfire. For example, black spruce trees have abundant lower branches which allow ground fire to quickly climb into the canopy and transform into large conflagrations. At mid elevations in alpine regions, vegetation cover becomes a mix of plants with forest and tundra affinities. Shrubs and herbaceous species become more abundant and arboreal cover decreases. As a result, fine fuel accumulation decreases reducing the ease which fire can initially ignite and subsequently spread through these mid-elevation ecosystems. At the highest elevations above treeline, aboveground biomass and fine fuel accumulation is lowest and thus ignition and spread of fire is reduced compared to lower elevations. As expected, lower valley regions dominated by closed canopy forests experience higher fire frequencies than upper elevations dominated by alpine tundra. While fire frequency certainly differs across elevational gradient in an alpine region, it is less certain if these typical fire frequencies can be discerned in lake sediment records because of the compressed vegetation zones in alpine environments. Markgraf (1980) has shown that pollen is easily transported across major vegetation units in alpine regions. Further, Pisaric (2001) documented the long-distance transport of charred macrofossils across considerable distance in mountain and valley terrain in southwestern Montana. Thus in an alpine setting there appears to be a possibility that macroscopic charcoal particles could also be transported across vegetation boundaries and confound the charcoal record.

To further explore the potential differences between the fire regimes at different elevations, I compared the macroscopic charcoal concentrations from each of the three cores from this study, as they were obtained from three different ecotones and elevations. YK-1 was obtained from a lake within the low-elevation forest at an elevation of 920 m a.s.l.. The dominant vegetation at this site consists of closed canopy forests comprised of white and black spruce, lodgepole pine, paper birch and trembling aspen. YK-3 was recovered from a mid-elevation (1310 m a.s.l.), with open-grown white spruce occurring in association with subalpine fir and shrub birch. YK-5 was obtained from a tundra lake at an elevation of 1510 m a.s.l., with the dominant vegetation consisting of shrub birch and willow, an assortment of herbs, various species of grass, bryophytes and lichens.

As initially predicted, the macroscopic charcoal concentration is not consistent between the three cores. The forest core (YK-1) has the highest average charcoal concentration of 19 particles/cm^3, with a maximum of 93 particles/cm^3 and a minimum of 2 particles/cm^3. This is due to the higher frequency of fires in the lower elevation closed-canopy forests. The CHAPS analysis of YK-1 resulted in an average fire frequency of 4 fires/1000 yrs, or 1 fire every 250 years. This value is consistent with fire frequency values obtained from previous studies, which suggest fires in closed-canopy forests of the boreal ecosystem occur once every 80 - 400 years (MacDonald, 2003).

The forest-tundra core (YK-5) has a relatively low average charcoal concentration of 5.5 particles/cm^3, with a maximum of 29 particles/cm^3 and a minimum of 0 particles/cm^3. However, throughout the majority of the record from YK-5 (12–173 cm) average charcoal concentration is slightly higher with 7 particles/cm^3. The average fire frequency determined using the program CHAPS analysis is 2.1 fires/1000 yrs, or about ~1 fire every 500 years. As expected, this mid-elevation core has a lower fire frequency than YK-1.

Similar to the forest-tundra charcoal record, the charcoal record from the tundra lake YK-3 has low average charcoal concentration (6 particles/cm^3). The average charcoal concentration for the tundra core is slightly higher than the forest-tundra core, which is likely a result of the influence of the large influx of charcoal between depths 127 - 60 cm (11,140 - 6360 ^{14}C yrs BP). The average charcoal concentration between 60 - 127 cm is 10 particles/cm^3, while the remainder of the core has an average charcoal concentration of only 2 particles/cm^3. The large influx of charcoal around 11,140 ^{14}C yrs BP may be due to an increase in fire frequency, as a result of a higher treeline during a period of warmer temperatures. A number of studies throughout the Canadian Rocky Mountains have documented higher treeline in the early Holocene based on a number of proxy data sources including fossil pollen and stomata (Kearney and Luckman, 1987; Pellat et al., 1998; Mazzuchi et al., 2003; Pisaric et al., 2003). Thus when conditions cooled and tundra vegetation prevailed around 6360 ^{14}C yrs BP, macroscopic charcoal values decreased significantly. The average fire frequency for YK-3 is 2.1 fires/1000 yrs or ~1 fire every 500 years. This value is slightly higher than expected, as the literature states alpine tundra typically burns only once every thousand years. As suggested above, the average fire frequency value is probably skewed, due to the higher fire frequency (2.96 fires/1000 yrs) at the bottom of the core (60-127 cm). The higher fire frequency during the early Holocene as suggested by the charcoal record from YK-3 reflects changes in the Earth's orbital parameters which altered the receipt of solar radiation.

5.2 Climate, vegetation and fire during the Holocene in southeastern Yukon Territory.

A number of studies have investigated the paleoenvironments of northwestern Canada and the possible factors that have influenced the changes in vegetation. These factors include: climate, fire, soil development, competition, pathogens and seed dispersal ability. More than one factor, or a combination of factors, may result in a similar change in vegetation and thus it is difficult to isolate a single factor which is responsible for a specific change in vegetational structure. Pollen records from northwestern Canada provide detailed information regarding the vegetation and climate of the region throughout the Holocene. I will draw upon the published pollen percentage diagrams from Hail Lake (Fig. 5.1) (Cwynar and Spear, 1995), Kettlehole Pond (Fig. 5.2) (Cwynar, 1988) and Snowshoe Lake (Fig. 5.3) (MacDonald, 1987) to aid in the interpretation and implications of the charcoal records obtained from my study lakes.

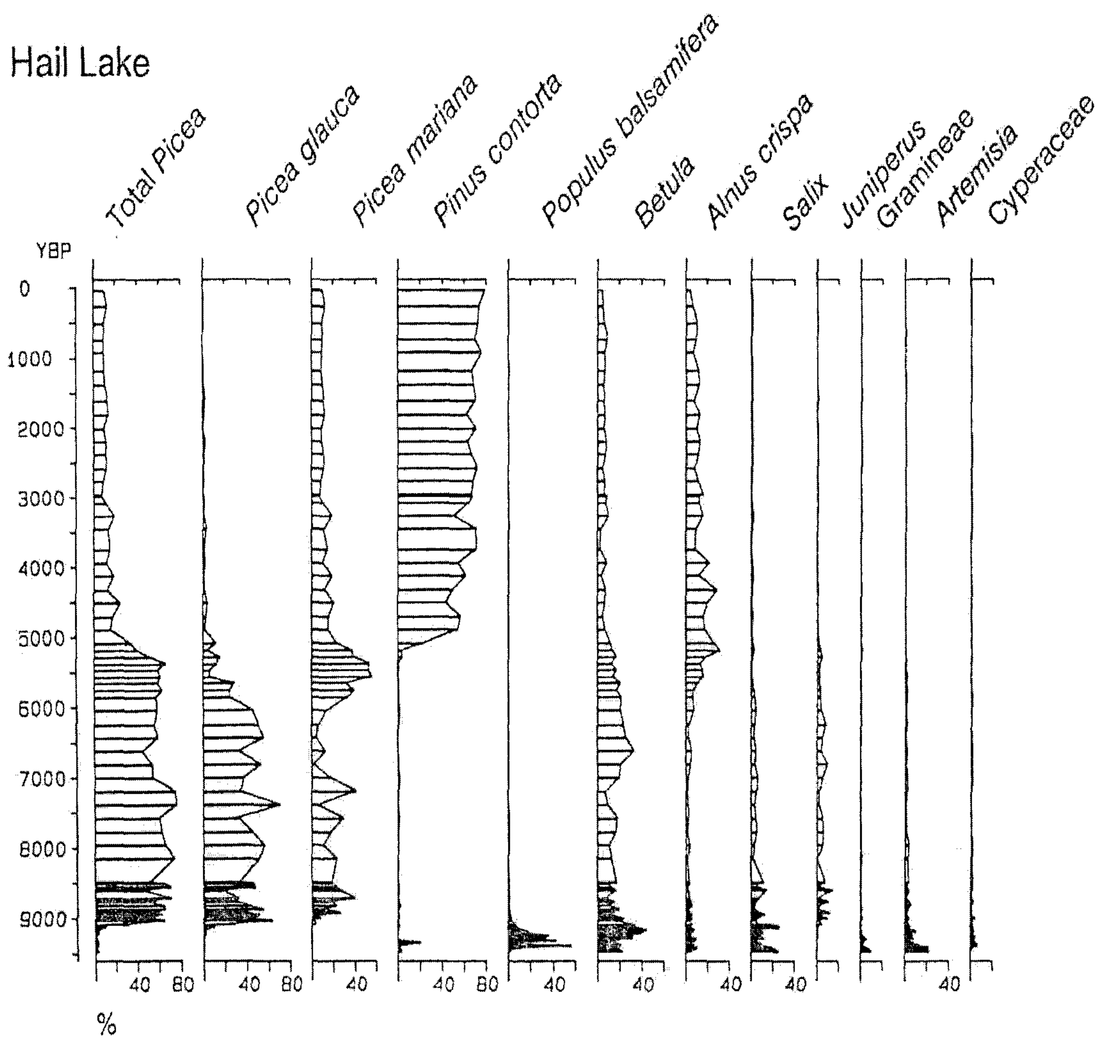


Fig. 5.1 Summary pollen percentage diagram from Hail Lake (Cwynar and Spear, 1995.)

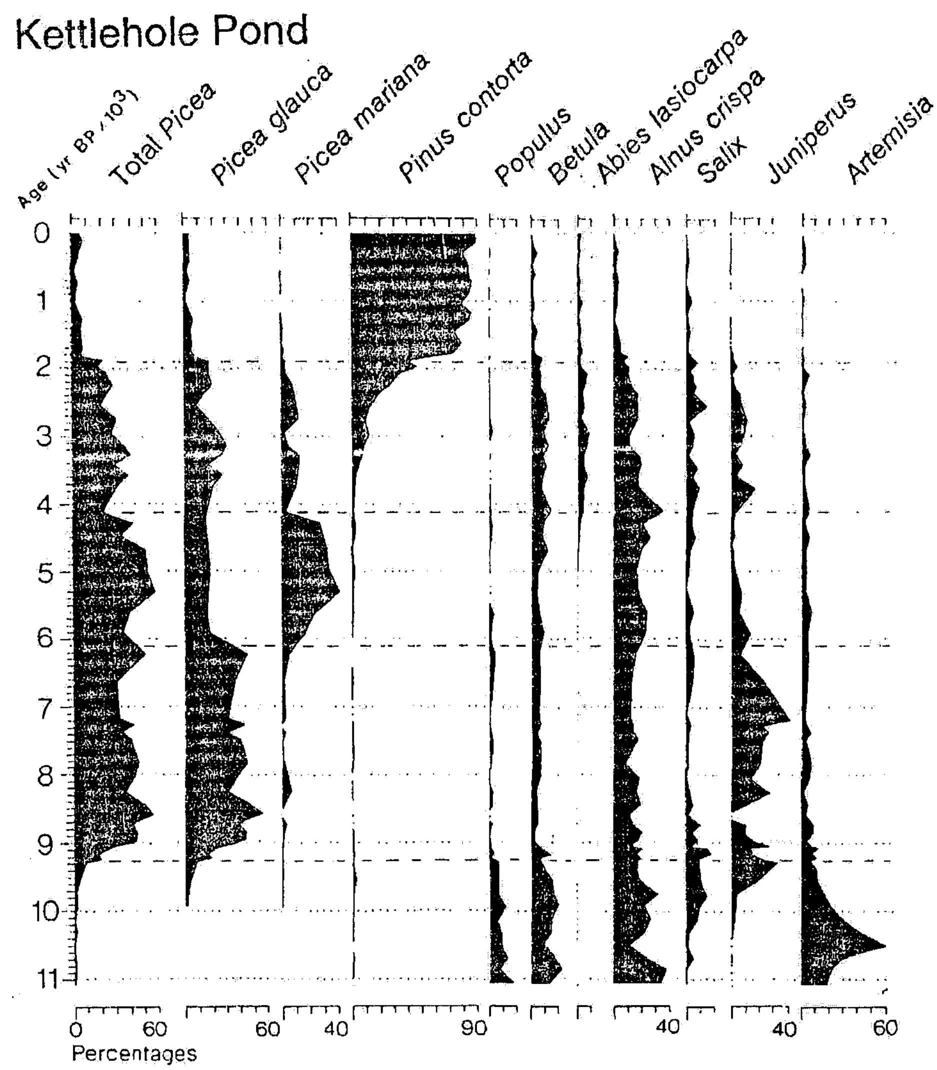


Fig. 5.2 Summary pollen percentage diagram from Kettlehole Pond (Cwynar, 1988).

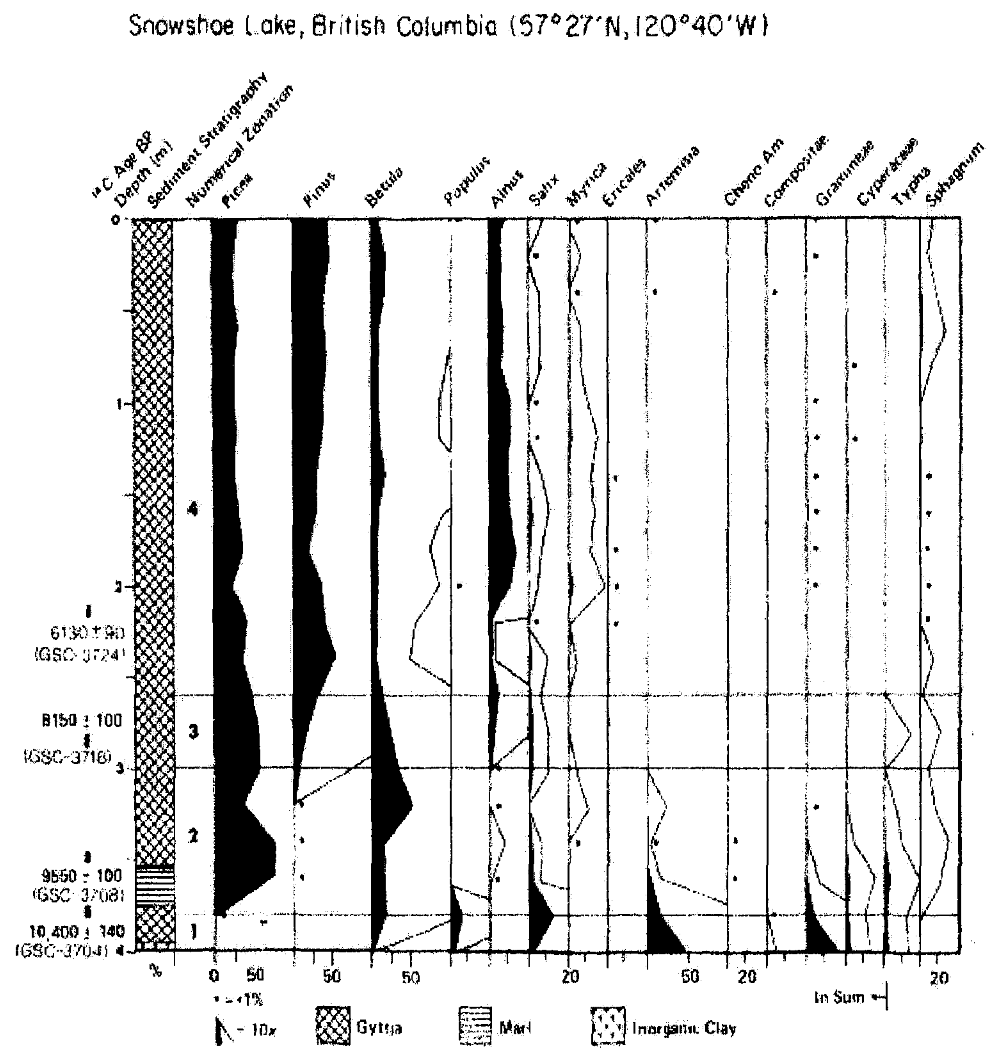


Fig. 5.3 Summary pollen percentage diagram from Snowshoe Lake (MacDonald, 1987).

5.2.1 Late Pleistocene-early Holocene Transition

Traditionally viewed as a time of relatively stable climate compared to that of the glacial and deglacial periods of the Pleistocene epoch, the Holocene has in fact experienced significant climate changes over various time scales (Cronin, 1999). The most recent glaciation peaked approximately 20,000 yrs BP and terminated by ~10,000 yrs BP. Following the retreat of ice ~12,000 yrs BP, the increased summer insolation led to warmer summer temperatures and decreased effective moisture, corresponding to the early Holocene Milankovitch thermal maximum (Ritchie et al., 1983). These astronomical characteristics of the earth's orbit and rotation are external forcing mechanisms which are the underlying cause of glacial-interglacial cycles and play an important role in controlling the cyclicity of climatic variations (Bradley, 1999). Orbital induced changes in the receipt of solar insolation were an important forcing agent during deglaciation and also for early Holocene climates. Solar insolation in the Northern Hemisphere was approximately 8% higher than at present during the summer months and approximately 8% lower than at present in the winter months between 12-9000 yrs BP (Cronin, 1999). This resulted in greater seasonality during the early Holocene than at present. Summers were at least 1-2°C warmer than present in northern regions and likely warmer based on recent palynological and biological data (Cronin, 1999; Pellatt et al., 1998; Smith et al., 1998; Palmer et al., 2002; Pisaric et al., 2003).

Evidence of an appreciable climatic warming in southern Yukon by 11,000 yrs BP is provided by the presence of *Typha latifolia* in the Snowshoe Lake pollen record (Fig. 5.1) (MacDonald, 1998). *Typha* spp. is a common aquatic plant in western Canada but does not occur north of the arctic or alpine treeline (MacDonald, 1988). *Typha* spp. pollen is not dispersed over large distances and thus its presence is indicative of post-glacial warmer temperatures throughout northwestern Canada (MacDonald, 1988).

Between 11,000 to 9000 yrs BP, pollen records from southern Yukon indicate a displacement of the early herb and shrub dominated vegetation by expanding *Populus* spp. and *Betula* spp. populations. Initially, the region was dominated by *Artemisia* spp. which reached values up to 50 - 60% of the pollen sum by ~10,500 yrs BP. *Gramineae* spp. reached a maximum value of 25% around ~9500 yrs BP, but then disappeared completely from the pollen record by ~9200 yrs BP. As the abundance of grasses and herbs declined around ~9500 yrs BP, poplar, birch and willow populations increased and dominated the region by ~9300 yrs BP. Studies in central Yukon have concluded that poplar likely was restricted to gallery forests and localized groves and did not form a continuous forest (Cwynar and Spear 1991). Birch pollen derived from dwarf birches was present in the pollen records from surrounding study sites, which provides further evidence of an open landscape (Cwynar and Spear, 1991). Discontinuous birch and poplar forests were also widespread in northern Yukon during this time (Cwynar, 1982). A period of peak abundance of poplar centered around 10,000 yrs BP is a common feature of pollen diagrams from Alaska, Yukon Territory and Northwest Territories (Ritchie et al., 1984, 1987 as in Cwynar and Spear 1991). Apart from the changing climate, the abundance of poplar was likely due to the bare mineral soils left by the retreating glacial ice (MacDonald, 1987). Disturbed mineral soils provide the best seed-bed for poplar, resulting in its dominance over early vegetation throughout the region

(MacDonald, 1987). Juniper (*Juniperus* spp.), which grows in arid climates, also appears in the pollen record at ~9500 yrs BP and reaches 30% of total pollen by 9200 yrs BP.

As post-glacial vegetation was relatively sparse and discontinuous, fires were not frequent or widespread during this time as indicated by the negligible charcoal influx of YK-3 at 12,000 – 11,000 yrs BP. Fire frequency between 12,160 – 11,140 yrs BP was found to be only 0.132 fires/1000 yrs. Fires at this time may have been small and not stand-replacing, and thus would not have produced much charcoal. Alternatively, the abundance of deciduous species such as poplar and birch may have controlled the burning temperature of these early Holocene fires. The fires may have burned at much lower temperatures resulting in more complete combustion of the organic material and thus less charcoal production. The charcoal record of YK-3 indicates higher levels of CHAR beginning at 10,500 yrs BP, which suggests a change in fuel conditions or fire severity in the area. Charcoal influx peaks at 9500 yrs BP with a value of 0.8 particles/cm^2/yr. Fire frequency reaches a high of 4.1 fires/1000 yrs at about 10,200 yrs BP. YK-3 is presently surrounded by tundra vegetation, with an average fire return interval of only 1.0 fires/1000 yrs. Thus, the changes in charcoal influx may imply an increase in fire activity due to an expansion of forest vegetation beyond present-day alpine treeline. The warmer and drier climate during the early Holocene may have triggered the replacement of alpine tundra with forest, providing sufficient fuel for fire activity. Pollen and macrofossil records from other areas of western Canada also provide evidence of early to mid-Holocene increases in treeline elevation (Kearney and Luckman, 1983; MacDonald, 1983; Ritchie et al., 1983). Warmer temperatures at this time may have influenced the treeline to move upslope at least 235 m higher than today (Pisaric et al., 2003). The main factor driving this change in treeline was likely increased summer warmth due to Milankovitch orbital variations (Ritchie et al., 1984).

A significant shift in vegetation occurs at ~9000 yrs BP, with the arrival of *Picea* spp. and a decline in the relative abundance of *Populus* spp.. Just prior to 9000 yrs BP, *Picea* spp. appears in the pollen record throughout the region, with *Picea glauca* being more prolific than *Picea mariana.* As white spruce grows best in warm and dry environments, its presence in the pollen record provides evidence of an early Holocene warm interval. The early dominance of white spruce may also have been aided by the prevalent soil conditions. The optimal seed-bed for white spruce germination is mineral soil with a thin veneer of duff, which would have been common in the post-glacial environment (MacDonald, 1987). In northern and central Yukon, a discontinuous forest of balsam poplar was replaced by extensive white spruce woodland between 10,200 - 9400 yrs BP (Cwynar and Spear, 1991;Cwynar, 1982). Thus, there is regional evidence of a change from open forests of poplar and dwarf birches to spruce-dominated forests by about 9000 yrs BP which was likely triggered by the onset of a warm and dry climate.

From 8000 to 6000 yrs BP, white spruce forests dominated most of the region with values up to 50% of the pollen sum. Birch continues to be present with values of 20% in the lower elevations (600 m) and up to 40% at higher elevations (900 m) (MacDonald, 1987; Cwynar and Spear, 1995). The percentage of *Juniperus* spp. increases with a value up to 40% of total pollen. The widespread establishment of white spruce and juniper suggests the climate was drier and warmer between 8000 - 6000 yrs BP than in the past. The fire record from YK-1 indicates a high level of fire activity at this time. Large peaks in CHAR between 7500 - 7000 yrs BP indicates an increase in fire occurrence, which is consistent with higher amounts of woody fuel build-up associated with closed forests. A mean fire interval of 175 years is estimated for this time, which is consistent with that observed in modern boreal forests (MacDonald, 2003). The vegetation and fire records therefore agree with warmer and drier conditions in the early Holocene.

The increased biomass at these alpine locations would have provided fine fuels to feed wildfires in areas that today lack a sufficient fuel load to carry frequent wildfires. The charcoal record from the tundra lake YK-3 suggests that was the case in this region. The highest fire frequency at this site occurs during this early Holocene episode. At the lower elevation site, YK-1 the early Holocene is also a period of increased fire occurrence. The forest-tundra record (YK-5) is more complicated and does not suggest more frequent fires in that location during the early Holocene. Given the very patchy nature of forest fires on the landscape, perhaps that location was not affected by a large fire during that time interval. Microclimatic conditions at these sites and YK-5 in particular may have minimized the occurrence of fire at that site. Further analysis of other proxy data (i.e. pollen and stomata) is needed from this site to draw more definitive conclusions. Combined with palynological and biological data from other studies in western Canada, the charcoal records from YK-1 and YK-3 provide additional evidence for a warm and probably drier early Holocene episode. These warmer and drier conditions were driven by changes associated with the Earth's orbit.

5.2.2 Middle Holocene

As previously mentioned, the early Holocene was a period of increased warmth, from about ~10,000 yrs BP to ~6000 yrs BP, due to the effects of increased solar insolation. By about ~6000 yrs BP, maximal insolation began to decrease to present-day levels consequently temperatures decreased and precipitation increased. Throughout the Yukon, this transition from warm/dry to cool/wet climate is characterized by the spread of green alder (*Alnus crispa*) and black spruce (Cwyner and Spear, 1995). Both of these species were present throughout the Yukon Territory prior to 6500 - 6000 yrs BP, so their expansion at this time suggests a change in environmental conditions. Pollen records from the region also indicate that the percentages of white spruce and juniper also declined during this transition period. White spruce populations decrease significantly from about 60% to only about 20% of the total pollen sum. Concurrently, the percentage of black spruce increases, reaching values of 60% of total pollen by ~5500 yrs BP (Cwynar and Spear, 1995). As black spruce generally grows on cooler and moister sites than white spruce, the expansion of black spruce suggests that climate must have become cooler, increasing effective moisture (Cwynar and Spear, 1995).

Green alder percentages increase with the rise in black spruce at 6000 yrs BP. An increase in alder, a plant which grows in moist habitats, also supports the interpretation of increased effective moisture (Cwynar and Spear, 1995). Alders have shallow root systems with nitrogen-fixing nodules and are commonly found on moist and organic soils (MacDonald, 1987). In northern Yukon, the rise of alder occurs a little earlier at about 8900 yrs BP (Cwynar, 1982). Across central Canada and the Northwest Territories, this rise has been found to usually occur between 5500 and 7800 yrs BP (MacDonald et al., 1993; Seppa et al., 2003).

The increased influx of *Sphagnum* spores around 6000 yrs BP also indicates the development of organic soils and peatland, which are common in cool and moist climates. The ability of black spruce to germinate and reach maturity on sphagnum surfaces is well known and it could have effectively colonized developing peat surfaces and resulted in the extensive *Sphagnum-Picea mariana* muskeg of the region during this time (MacDonald, 1987). Thus, there is a regional shift from white spruce forest to a more mesophytic black spruce woodland, beginning at 6000 yrs BP (Cwynar and Spear, 1995). By the end of the middle Holocene, black spruce populations begin to decrease and are replaced by lodgepole pine, which becomes well established by 3000 yrs BP.

Even though the climate appeared to become more mesic at this time, the charcoal records from YK-1 and YK-3 suggest that fire frequency increased. Prior to ~6500 yrs BP, fire frequency at YK-1 was less than 2 fire events/1000 yrs. Similarly, fire frequency was also lower for the alpine lake YK-3. Between 5500 and 5000 yrs BP, fire frequency increased at both lakes. At the low elevation lake YK-1, fire frequency increased to ~4 fire events/1000 years while in the alpine environment fire frequency increased to ~3 fire events/1000 years. This increase in fire occurrence may be the result of the widespread establishment of fire-prone black spruce. The growth form of black spruce, with many lower branches near the ground, promotes the spread of fire through a stand of black spruce trees. Similar to the current study, recent work in Alaska indicates higher fire frequencies after the establishment of black spruce around 6000 - 5500 yrs BP (Lynch et al., 2003, 2004). The increase in fire frequency at the same time that black spruce was becoming established in Alaska, suggests that vegetation type is also an important driving factor of fire regimes in boreal ecosystems (Rupp et al., 2002; Lynch et al., 2003, 2004).

5.2.3 Late Holocene

The onset of the transition to modern climate conditions commenced at ~4000 yrs BP. Climatic conditions throughout northwestern Canada continued to cool and some regions experienced further increases in moisture. The late Holocene period features increased values of *Pinus* spp. with a simultaneous decrease of *Picea* spp., *Alnus* spp. and *Betula* spp.. *Pinus* first appeared in the pollen record as early as 8800 yrs BP, but did not become widespread throughout southern Yukon until about 3000 yrs BP, reaching values of 80% of the total pollen sum (Cwynar, 1988; Cwynar and Spear, 1995).

The charcoal records from YK-1 and YK-3 suggest that fire activity increased at 3500 yrs BP until 2500 yrs BP. Lodgepole pine is tolerant of fire, and the increase in pine populations at this time corresponds with the increase in fire activity as determined from the charcoal record. Lodgepole pine has serotinous cones and germinates best on recently burned, well-drained soils. The resinous components of lodgepole pine are highly flammable and promote fire activity (MacDonald, 1987). Thus, the increased fire activity at this time, as noted in the charcoal record of YK-1 and YK-3, may be the result of the widespread establishment of lodgepole pine. Previous studies have also suggested a slight warming between 3400 to 2600 yrs BP (Kearney and Luckman, 1987; Szeicz and MacDonald, 2001), which could also influence the fire regime.

CHAPTER SIX

6.0 CONCLUSIONS AND RECOMMENDATIONS

6.1 Conclusions

This study examined the macroscopic charcoal from three sediment cores from the southeastern Yukon Territory. The charcoal data was analyzed by the use of CHAPS to distinguish peaks representing fire events from the slowly-varying background level of charcoal influx. The charcoal record was also compared to pollen stratigraphies from lakes from the surrounding area, to examine the relationship between vegetation composition, climate and fire frequency throughout the Holocene. As initially hypothesized, the results from the charcoal records from the three cores indicate that the fire regime of the southeast Yukon was dynamic and fluctuated throughout the Holocene. The results from the lower elevation forest core (YK-1) and the alpine-tundra core (YK-3) illustrate that fire frequency responded to changes in climate; the fire frequency increased during the Early Holocene when the climate was warmer and drier than present, and then decreased during the mid-Holocene when the climate became more cool and moist. In contrast, results from the forest-tundra core (YK-5) did not support this conclusion. As the occurrence of fire is not only influenced by climate but is governed by a set of complex interactions between abiotic and biotic factors, this study also looked at the type of vegetation that was present at times of increased and decreased fire frequency. As hypothesized, fire frequency facilitated changes in the composition of forests in southeast Yukon Territory throughout the Holocene. The charcoal record from YK-1 and YK-3 provides evidence that periods of increased fire frequency during the late Holocene were associated with extensive stands of fire-prone species such as black spruce and lodgepole pine, despite the onset of a more cool and moist climate. This study also yielded the following conclusions:

1. The comparison of charcoal records between YK-1 and YK-3 reveals that fire frequency changes with elevation. Elevation indirectly influences the fire regime by determining the type and structure of plants in a given region. The fire frequency was found to be highest in the lower elevation closed-canopy forests and lowest in the high elevation alpine tundra.
2. The large influx of macroscopic charcoal during the early Holocene in the high elevation tundra core (YK-3) suggests an increased fire frequency as a result of the expansion of vegetation upslope beyond the present-day alpine treeline. Warmer temperatures at this time would have provided a more conducive environment for the replacement of alpine tundra with more arboreal vegetation, providing additional fuel for fire activity. This increase in fire frequency during the early Holocene as suggested by the charcoal record of YK-3, reflects changes in the Earth's orbital parameters which altered the receipt of solar radiation and resulted in a warmer climate.
3. The early Holocene was a period of increased warmth and drier conditions from about ~10,000 yrs BP to 7000 yrs BP, corresponding to the early Holocene Milankovitch thermal maximum. Large peaks in CHAR in the fire records of YK-1 and YK-3 indicate high fire occurrence during this period. Pollen diagrams from the surrounding areas also indicate the widespread establishment of white spruce and juniper, which is also indicative of a warmer and drier climate.

4. The climate of the mid Holocene from about ~7000 to 4000 yrs BP became cool and moist as maximal insolation began to decrease to present-day levels. This transition from warm/arid to cool/moist climate is characterized by the spread of green alder and black spruce, which favor such an environment.
5. The late Holocene period from ~4000 yrs BP to the present is typified by the onset of modern-day climate. Climatic conditions throughout northwestern Canada continued to cool and some regions experienced further increases in moisture, noted by the decrease of pollen percentages of spruce, alder and birch. Lodgepole pine is widespread at this time, which corresponds with an increase in fire activity as determined by the charcoal records of YK-1 and YK-3.
6. The charcoal record of the forest-tundra core YK-5 is complicated and does not provide any substantive evidence of trends in fire activity throughout the Holocene. Microclimatic conditions and/or topographic parameters at this site may have been different than the surrounding areas, and perhaps may not have been suitable to render charcoal and establish a substantive record of fire occurrence at the site. Attention must be brought to the many processes which charcoal is subject to, prior to being deposited in a lake basin and which may change the charcoal composition found in the sediment. Consequently, the charcoal concentration may not reflect the true fire frequency at the site.
7. The use of the CHAPS program proved to be a very useful tool for differentiating fire events from background levels of charcoal. The program provided a statistical analysis of the charcoal data, which allowed for a more sound

interpretation of the results and conclusive evidence of the fire frequency of the study site area.

6.2 Future Recommendations

This study provides evidence that changes in climate and vegetation can be related to changes in the fire regime of a region. Already many effects of climate change are becoming noticeable in the boreal forests of northwestern Canada, including larger and more extensive fires (ACIA, 2004). The boreal forests are extremely important globally for their economic and environmental values, such as timber production, wildlife habitat and freshwater collection and distribution. Thus, there is a need to study the future impacts of global warming on the boreal forest and surrounding environment. Additional charcoal and pollen studies can be done throughout the Yukon to further understand the relationship between climate, vegetation and fire occurrence and to accumulate more data which can be used to model and predict the potential outcomes of a warmer climate on this ecosystem.

The sediment cores from this study could also be analyzed for their pollen content, in order to have a better understanding of the vegetation of the surrounding area and provide a better analysis of the fire and vegetation relationship. Also, the examination of charcoal and pollen from various sediment cores from the surrounding area would also provide a better regional picture of fire activity and reduce the effects of microclimatic conditions at individual sites.

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