CLIMATE CHANGE AND POSTGLACIAL ENVIRONMENTAL HISTORY OF PERMAFROST PEATLANDS IN THE MACKENZIE DELTA AREA, N.W.T.

by

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A thesis
presented to the University of Waterloo
in fulfilment of the
thesis requirement for the degree of
Doctor of Philosophy
in
Earth Sciences

Waterloo, Ontario, Canada, 1997

0-612-21394-3

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ABSTRACT

The Holocene environmental histories of two peatlands in the Mackenzie Delta area were reconstructed in order to gain an understanding of how climate change, permafrost and build-up of ground ice affect the processes of peat accumulation in an area of continuous permafrost. Multi-proxy paleoecological analyses, including peat stratigraphy, pollen, plant macrofossils and radiocarbon dating were carried out on cores from Kukjuk peatland, on the Tuktoyaktuk Peninsula 75 km north of present treeline, and Campbell Creek peatland, 20 km south of treeline. Stratigraphy and physical characteristics (bulk density and content of moisture, organic material and inorganic carbon) were analyzed for six cores from Kukjuk peatland and six cores and two peat sections from Campbell Creek peatland. Two cores from each site were selected for pollen and plant macrofossil analyses.

The paleoecological reconstructions are compared with a well established record of regional postglacial environmental history available from previous lake sediment studies, to investigate how past climate changes affected the peatlands. This independent paleoenvironmental record, in combination with the analyses of several cores from each of the peatlands, aids in distinguishing stratigraphic changes in the peat deposits resulting from autogenic succession and local environmental factors from those that are more likely to reflect regional climatic change. An early Holocene period of warmer than present climate has been inferred, based on evidence from lake pollen records and spruce macrofossils that the treeline was further north from 9000-5000 BP, covering at least part of the Tuktoyaktuk Peninsula.

Organic deposition began in Kukjuk peatland by 7200 BP, and in Campbell Creek peatland by at least 9000 BP. Both sites were initially occupied by open water mineral wetlands with emergent and submergent aquatic vegetation. A subsequent fen stage is identified in both peatlands, but with some differences in plant communities. A switch to *Sphagnum*-dominated ombrotrophic conditions occurred in both peatlands between approximately 4000-5000 BP. This coincides with a deterioration of regional climate and the retreat of treeline to its present position. The changes in the peatland ecosystems may have been an indirect response to this climate change, linked to permafrost aggradation which altered the surface hydrology, allowing the establishment of ombrotrophic vegetation.

A theoretical model of peatland development by terrestrialization in continuous permafrost areas is presented, based on the reconstructed peatland histories. Comparison with published stratigraphical and paleoecological records from other peatlands, mostly in northwestern Canada, suggests that similar peat accumulation processes have occurred at many other sites. However, the model will not apply to all permafrost peatlands, since variations in environmental conditions affecting peat accumulation occur at regional and local scales, and even within a single peatland. The multi-proxy, multi-core approach applied here is recommended for studies aimed at improved understanding of peat accumulation processes, in permafrost or non-permafrost environments. It provides a more complete picture of the history of the entire peatland ecosystem than can be attained by applying a single technique to one core.

ACKNOWLEDGEMENTS

This thesis is a contribution to PACT (Paleoecological Analysis of Circumpolar Treeline), a Canadian-Russian collaborative, interdisciplinary study of circumboreal treeline fluctuations. I am grateful for the opportunity to participate in this project, which is headed by Drs. Les Cwynar, Glen MacDonald and Andre Velichko and includes an international team of scientists applying their individual areas of expertise towards a common goal. Financial support for my research was provided in part by an NSERC Special Collaborative Grant to PACT. Additional support for field work was provided by the Inuvik Research Centre of the Science Institute of the Northwest Territories, the Polar Continental Shelf Project (Tuktoyaktuk base) and the Northern Scientific Training Programme.

Isotope data cited in Chapter 6 was provided by Ramon Aravena. Bob Ingleton and Jason Wolki provided invaluable assistance and always pleasant companionship in the field. Stephen Marsters, Mark van Bakel and Jean-Francois Ouillet helped with preparation of some samples, and other past and present lab-mates provided thought-provoking discussion, much encouragement and occasional, well-timed comic relief.

Thanks are due to Dr. Barry Warner for financial support in the form of research assistantships, for providing the opportunity to fulfill a life-long dream of working in the Arctic, and especially for his patience and endless enthusiasm. I also thank the other members of my supervisory committee, Drs. Jocelyn Legault, Sherry Schiff, Rick Soulis and John Jeglum, for their advice and encouragement; and external examiner Dr. Dorothy Peteet for her interest and helpful feedback.

I am especially grateful to Ian for a great deal of moral and technical support, and to my parents, John and Annie Vardy, for not trying to dissuade me from dropping out of business school to go back to my childhood occupation of playing with mud.

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1. INTRODUCTION

1.1 INTRODUCTION

Peatlands are landforms composed of the remains of organic material, characteristic of waterlogged situations in which, because of anoxic and cool conditions below the surface, organic detritus accumulates to depths greater than 40 cm and often up to several metres (Clymo, 1984, 1991). Peat is composed of the preserved remnants of plants that grew on or near the peatland. As peatlands evolve, they therefore preserve a stratigraphic record of changes in the vegetation on the peatland and in the surrounding area, which may reflect environmental changes at either local or regional scales. Regional changes in climate or terrain features and local changes in hydrology, water chemistry, vegetation and the activities of animals (especially beavers) have all been recognized as important factors in determining peatland development.

Peatlands cover extensive areas in subarctic and low arctic regions, where the majority are underlain by permafrost. Zoltai (1988) estimated that peatlands cover a total of 26409 x 10^3 ha in the Yukon and Northwest Territories. These peatlands are concentrated mainly in the western Northwest Territories from the Alberta boundary north to the Tuktoyaktuk Peninsula, and the vast majority are within the zones of continuous or widespread discontinuous permafrost (Tarnocai *et al.*, 1995). Some areas of the extensive peatlands covering more than

72% of the land area in the Hudson-James Bay lowlands of northern Manitoba, Ontario and Quebec are also affected by permafrost (Tarnocai *et al.*, 1995). The presence of permafrost can have significant impact on peatland ecosystems, and should therefore be included in the list of factors influencing their development.

In general, peat accumulates in waterlogged environments and is associated most commonly with areas of high precipitation or low evaporation rates, or both. While precipitation is low in many arctic and subarctic areas, widespread permafrost is conducive to the development of peatlands. Permafrost acts as a barrier to water movement, so that most available moisture from precipitation and snow melt is retained on or near the surface. Organic materials tend to be well preserved in permafrost peat because once they become part of the perennially frozen layer, decomposition ceases.

It is important to improve our knowledge of the effects of past climate change in arctic and subarctic regions, since most General Circulation Models (GCMs) and vegetation response models based on GCM output suggest that past and future climatic fluctuations and related vegetation changes are likely to be most extreme at high latitudes (Emanuel *et al.*, 1985; Hansen and Lebedeff, 1987; Houghton *et al.*, 1990; MacCracken *et al.*, 1990; Maxwell, 1992; Rizzo and Wiken, 1992). Future climate warming is also expected to have significant effects on northern peatlands (Gorham, 1988, 1991). Some of the strongest evidence for large scale changes in climate at high latitudes comes from paleoecological records from lakes and peatlands, which show latitudinal fluctuations in the position of northern treeline during the Holocene. One of the most definitive records of treeline movement in North America is from the Mackenzie Delta region, where a dense network of lake sites to the east of the delta and on the Tuktoyaktuk Peninsula provide evidence of a significant northward advance of treeline during the early Holocene. This has been attributed to a period of warmer than present climate, possibly related to a Milankovitch insolation maximum (Ritchie *et al.*, 1983). Part of the Tuktoyaktuk Peninsula, which now has a shrub tundra vegetation cover, was occupied by forest or forest-tundra vegetation and summer temperatures are thought to have been significantly higher than present. The limits of forest subsequently shifted southwards, reaching its present position at about 4500 BP (Ritchie, 1984).

1.2 OBJECTIVES AND RATIONALE

The objectives of this thesis are:

1) to compare trends of peatland development in the Mackenzie Delta region with the independent record of Holocene paleoenvironmental and climate history from lake sediments, in order to gain an understanding of the response of arctic and subarctic peatlands to climatic change; and

2) to investigate the general processes of peat accumulation in subarctic and low arctic peatlands during the Holocene, and the effect of permafrost and ice accumulation on peatland development.

The Mackenzie Delta area was chosen because of the well established record of postglacial environmental history available from several lake sediment studies. Application of the paleoclimatic reconstructions derived from these records to the patterns of peat stratigraphy in the region will help us understand how climate change affects wetland dynamics and peat accumulation in permafrost peatlands. The study area covers a steep climatic gradient, among the steepest in North America, and spans two major biogeoclimatic zones, the forest-tundra and the tundra. It also encompasses parts of two wetland zones, as defined by the National Wetlands Working Group (1988): the High Subarctic and Low Arctic Wetland Zones. It therefore offers an ideal opportunity to study peatlands which have developed in different local environmental settings, yet in close enough geographical proximity to be affected by the same regional climatic changes.

Multi-proxy paleoecological analyses including peat stratigraphy, pollen, plant macrofossils and radiocarbon dating were carried out on cores from one peatland near the present treeline in the Inuvik area and another on the Tuktoyaktuk Peninsula 75 km north of treeline (Figure 1.1), to determine the relationship between peat development and the climatic changes that apparently caused Holocene treeline fluctuations in the region. A substantial body of evidence indicates that during the early Holocene, forest or forest-tundra occupied much of the Tuktoyaktuk Peninsula. This would require environmental conditions considerably different from those experienced by the tundra-covered peninsula today, with significantly warmer summer temperatures (Ritchie, 1984). A peatland developing on the peninsula may have been affected by different environmental conditions than another which has been within the forest-tundra zone for the entire Holocene, but may also have been subjected to regional climatic changes.

By analyzing multiple peat profiles from the different basins and comparing the results with the independent paleoenvironmental records, it should be possible to distinguish stratigraphic changes in the peat deposits due to autogenic succession and local environmental factors from those that are more likely to reflect regional climatic change.

Chapter 2 provides relevant background information and a review of previous knowledge of permafrost peatland systems. The regional setting and Holocene environmental history of the study area are described in the third chapter, followed in the fourth chapter by an explanation of the methods used. The results from Kukjuk peatland on the Tuktoyaktuk Peninsula and Campbell Creek peatland south of Inuvik are presented and discussed in Chapters 5 and 6, respectively, followed by a comparison and further discussion of the results from the two sites in terms of the stated objectives.

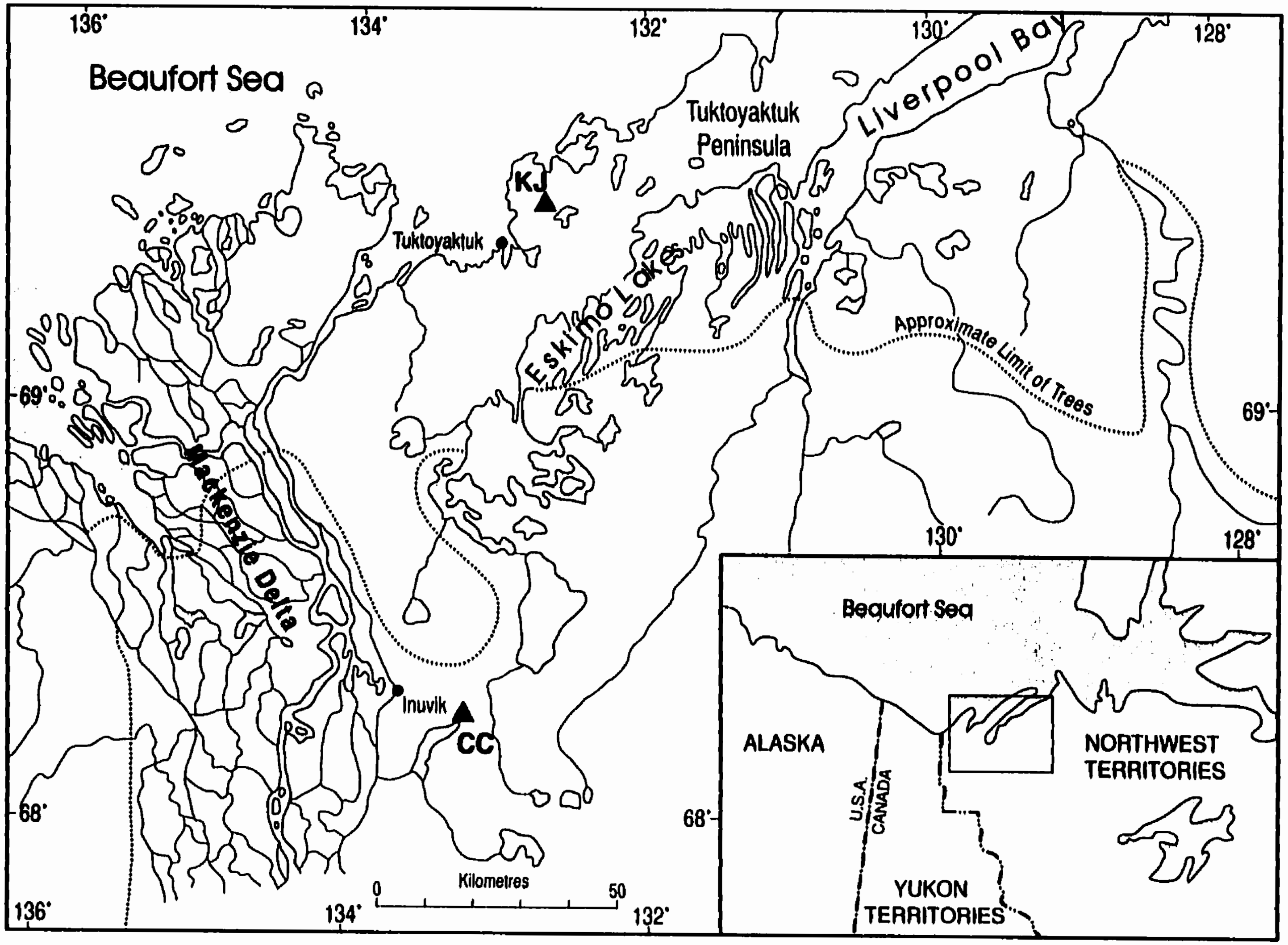


Figure 1.1. The Mackenzie Delta area, with locations of study sites indicated: KJ - Kukjuk Peatland; CC - Campbell Creek peatland.

2. PERMAFROST PEATLANDS

2.1 PERMAFROST - INTRODUCTION

Permafrost is defined as a subsurface material that remains frozen throughout a period of several years. This includes ground that freezes in one winter, remains frozen through the following summer and into the next winter (Brown, 1970). Formation of permafrost requires a negative heat budget giving soil temperatures less than 0°C for a time period of at least one summer (Railton and Sparling, 1973). In general, the overall distribution of permafrost is related to air temperature. Permafrost will form if the mean annual air temperature of an area cools to about 0°C or lower (Brown, 1973).

Where perennially frozen ground prevails everywhere except below and near oceans and large lakes, the area is described as being in the zone of continuous permafrost. In practice, this is the area north of a line delimiting the -5°C ground temperature isotherm, measured at the depth where annual variations do not occur (Brown, 1967). South of this is a zone where permafrost is present under some surfaces but not all. The boundary between continuous and discontinuous permafrost zones corresponds to a mean annual air temperature of -8.3°C, while the southern limit of discontinuous permafrost corresponds approximately with the -1°C isotherm of mean annual air temperature (Brown, 1978).

The zone of substratum immediately below the surface that freezes and thaws each year is known as the active layer. Its thickness varies according to the thermal properties of the surface materials, the vegetation cover, the slope and aspect, and the soil moisture conditions.

Permafrost is widespread in North America, Greenland, northern Eurasia (including northern Scandinavia, Russia, Outer Mongolia and Manchuria), Antarctica, and at high elevations in mountainous regions of other parts of the world. It underlies approximately 50% of Canada's land surface, including many of its peatlands (Brown, 1970; Tarnocai *et al.*, 1995).

2.2 PEAT-PERMAFROST RELATIONSHIPS

The relationship between permafrost and peatlands is complex, since the presence and nature of peat influences the formation and aggradation of permafrost, and permafrost in turn has considerable impact on the dynamics of peatlands.

It is well known that in the present discontinuous permafrost zone, permafrost is more likely to develop in peatlands than in mineral soil (Brown, 1963, 1970, 1973, 1980; Zoltai, 1972; Zoltai and Tarnocai, 1975; Payette *et al.*, 1976; Tarnocai, 1982; Couillard and Payette, 1985, Seppälä, 1988; Vitt *et al.*, 1994). Brown (1970) explains that this phenomenon is a result of changes in the thermal conductivity of peat throughout the year. *Sphagnum* peat, in particular, has been found to retard thawing by keeping soil temperatures low all year. During the summer, when the surface layers of peat are dried out by evaporation, the thermal conductivity of the peat is low (approximately 0.00017 g cal/sec cm^2 °C cm; Brown, 1963) and warming of the underlying soil is impeded. In the fall, the surface layers of the peat are wetter because of the melting of early snow and lower evaporation rates. Wet peat has a much higher thermal conductivity, estimated at 0.0007g cal/sec cm^2 °C cm for unsaturated peat and 0.0011g cal/sec cm^2 °C cm for saturated peat (Brown, 1963), so heat is conducted at a more rapid rate, permitting greater heat loss from the ground. When peat freezes, its thermal conductivity is further increased (approximately 0.0056 g cal/sec cm^2 °C cm for saturated frozen peat), allowing deep penetration of seasonal frost (Brown, 1963). As a result, peat offers less resistance to cooling of the underlying soil in winter than to warming of it in summer, creating conditions favourable for the development and expansion of permafrost.

While any vegetation cover affects the transfer of heat to and from the underlying soil, *Sphagnum* in particular seems to be an effective insulator. In the discontinuous permafrost zone, permafrost occurrence is rare in wet sedge-grass areas with little or no moss cover, and much more common in better-drained peat with *Sphagnum* cover. The depth of thaw decreases with an increase in combined thickness of living *Sphagnum* and peat (Brown, 1963).

The importance of moss and peat in insulating permafrost is demonstrated by the fact that little change is found in the depth to permafrost if only trees and shrubs are removed from a peatland surface (Brown, 1970). Trees and other vegetation are also important influences on ground temperature and permafrost, however, since they shade the ground from solar radiation in summer, intercept some of the insulating snowfall in winter, and affect ground surface wind velocities (Brown, 1970). Cowell *et al.* (1978) found that in the Hudson Bay Lowlands, near the southern margin of the discontinuous permafrost zone, *Picea mariana* islands in fens and bogs always had a layer of frozen peat. They identified vegetation cover, peat type, and water movement as the three main factors promoting frozen layers in wetlands, and found that in the Hudson Bay Lowlands, permafrost was most common in bogs, which have the slowest rates of water movement, and was rare in fens, swamps, and marshes, where surface waterflow is more rapid.

Observations of numerous permafrost peatlands lead Zoltai and Tarnocai (1975) to conclude that in areas of discontinuous permafrost, perennially frozen layers are initiated as particular vegetation conditions develop. Most commonly, small *Sphagnum* cushions develop on the surface of fens, insulating seasonal frost against rapid summer thawing. Eventually a small lens of frost under the cushion fails to thaw and becomes permafrost. This lens gradually thickens, and water in the peat changes to ice, further elevating the surface above the water table so that it becomes drier and insulation further improved. Small trees may become established on the cushions, intercepting some snow, therefore reducing the insulting effect of the winter snow layer. This process appears to be the main mechanism of permafrost initiation in peatlands of the discontinuous permafrost zone. Nearly all permafrost peatlands are covered by at least a thin surface layer of peat which differs in origin from the underlying material (Zoltai, 1995).

While permafrost is present under all land surfaces in the continuous permafrost zone, similar processes probably lead to the aggradation of permafrost as peatlands develop. The distribution of peat also seems to influence the occurrence of ground ice in both the continuous and discontinuous permafrost zones. Heginbottom *et al.* (1978), in a study of 11,600 boreholes drilled for geotechnical purposes in the Mackenzie Valley, found that poorly drained sites with a thick cover of peat contained more ground ice than adjacent drier sites.

Permafrost, in turn, modifies peatlands by raising them above the water table, so that previously saturated surface layers become drier and completely different vegetation communities can develop (Zoltai and Tarnocai, 1975). As well, internal drainage may be blocked by frozen lenses, altering the drainage pattern of the peatland. Permafrost also alters the environment of vegetation growing on the peatland surface, since plant roots are restricted to the active layer, and temperatures in the root zone during the growing season are low (Johnson, 1963).

Permafrost impedes drainage of water derived from precipitation and runoff, since it creates an impervious layer similar to bedrock. As a result, large areas of land in the low Arctic and subarctic are associated with wetlands and water saturated soils, even in areas where precipitation is low, as it is in the Mackenzie Delta area (Tarnocai and Zoltai, 1988). The presence of masses of ground ice lead to surficial irregularities, while melting of this ice produces thermokarst topography.

2.3 *FORMS OF PERMAFROST PEATLANDS*

A variety of peatland forms are unique to areas of permafrost, including palsas, peat plateaus, low- and high-centre polygonal peat plateaus, and a variety of fen types. Palsas are mounds of peat with permafrost cores, rising 1-10 m above the surrounding peatland and having diameters on the order of 10-100 m (Zoltai and Tarnocai, 1975; Railton and Sparling, 1973; Zoltai *et al.*, 1988). They occur as islands in wet, unfrozen fens or ponds, or as peninsulas extending into these non-permafrost, wet areas. They can develop individually, but sometimes coalesce into contorted ridges and swales of peat occupying several hundred hectares. Their internal structure includes a surface layer of *Sphagnum* peat up to 50 cm thick, underlain by fen peat usually composed of sedge and brown moss remains. The basal peat is humified organic matter or aquatic detritus. Thin (up to 10 cm) ice lenses are often found in the frozen peat, but large accumulations usually occur at the peat-mineral soil interface. Palsas are found mostly in areas of discontinuous permafrost and are common in the low subarctic of North America as well as in Fennoscandia and extensive areas of northern Russia.

Peat plateaus are generally flat, perennially frozen deposits elevated about 1 m above the lowland water table, occurring as small (few m^2) to large (several km^2) islands in fens or as thick peat deposits on slightly sloping mineral terrain (Brown, 1970; Zoltai and Tarnocai, 1975; Zoltai *et al.*, 1988). The surface vegetation and morphology vary widely, but the internal structure generally consists of a cap of *Sphagnum* peat overlying the main deposit, which is composed of brown-moss or moss-sedge peat, often with shrub remains or with aquatic peat near the base (Zoltai and Tarnocai, 1975).

In the peat plateaus of the Continental High Subarctic Wetland Subregion, the main shrub species are *Ledum decumbens* in the open woodlands and *Ledum groenlandicum* in the forested areas, as well as *Andromeda polifolia* and *Betula glandulosa*. *Vaccinium vitis-idaea* and *Rubus chamaemorus* are also widespread. *Sphagnum fuscum* may grow in small, moist depressions, but the main ground cover is provided by *Cladina mitis, C. stellaris, C. rangiferina*, and *C. amauracraea* (Zoltai *et al.*, 1988).

Polygonal peat plateaus are also elevated approximately 1 m above the adjacent fens. They are relatively flat, but have a polygonal network of ice trenches surrounding polygons 15-30 m in diameter. A wedge of clear ice extends downwards under each trench for 2-4 m. These peatlands are completely perennially frozen beneath the seasonal active layer, with permafrost extending into the mineral soil below (Zoltai and Tarnocai, 1975). There are two main morphological distinctions, depending on whether the centres of the polygons are lower or higher than the edges. The middle of low-centre polygons are usually wet, with standing shallow water during most of the summer period, and supporting *Carex* spp. and *Eriophorum* spp. and mosses such as *Calliergon giganteum* and *Drepanocladus revolvens*. The raised shoulder along the trenches of low-centre polygons often support *Betula glandulosa* and ericaceous shrubs. The central part of a high-centre polygon is domed and therefore well-drained, and is either bare or have surface vegetation of *Betula glandulosa* and lichens. High-centre polygons are believed to evolve from low-centre polygons and therefore represent a later stage of peatland development (Zoltai and Tarnocai, 1975; Tarnocai and Zoltai, 1988).

Low-centre polygons are characterized by shallow sedge and brown moss peat. As the peatland develops into a high-centre form, drainage improves and there is a resulting increase in shrub species. The peat associated with high centre polygons is usually moderately or well-decomposed and acidic, ranging in pH from 3.4 to 6.9. Mineral content ranges between 7.6 and 39.7%, generally higher than that of peat materials in southern Canada (Tarnocai and Zoltai, 1988).

The ice wedges in polygonal peat plateaus are similar to those which develop in mineral soils in Arctic regions, which have been studied intensively. Lachenbruch (1962) describes the contraction theory of ice-wedge polygons first proposed by Leffingwell in 1915. During winter, narrow vertical fractures form in frozen soil, as a result of tension caused by thermal contraction at the tundra surface. In the spring, these cracks are filled by precipitation, melting snow and hoarfrost, which freezes into the cracks and produces a vertical vein of ice that penetrates the permafrost. In following winters, renewed thermal tension re-opens the crack, and additional increments of ice are added when spring meltwater enters the renewed crack and freezes. Such a cycle, acting over centuries, is theorized to produce the vertical wedge-shaped masses of ice. The polygonal pattern is presumed to be a natural consequence of contraction origin. Horizontal compression caused by re-expansion of the permafrost during the following summer results in the upturning of strata adjacent to the ice wedge, creating a ridge of ground on each side of the ice wedge (Péwé, 1963).

Price (1972) hypothesized that high centre polygons represent an eroding, melting phase in which the ice wedges are inactive. However, Tarnocai and Zoltai (1988) point out that high-centre polygons have a consistently greater thickness of peat, suggesting that peat formation also influences the surface morphology. A complete range of low-centre to high-centre polygons is found in arctic peatlands. The polygon shoulders become thicker, and the enclosed pools smaller as the peat accumulates, until the surface becomes level (Zoltai and Tarnocai, 1975). The development of a domed centre, characteristic of high-centre polygons, may be due to partial melting of the ice wedge during a late stage (Price, 1972).

Fen types found in the subarctic and low arctic include northern ribbed fens, which have narrow, drier peat ridges running at 90° to the direction of water movement (Zoltai *et al.*, 1988). Fens occur even in the High Arctic Wetland Region, in places where deep snow drifts accumulate on the lee side of hills, leading to the formation of small 'snowpatch' fens on lower, gentle slopes, nourished by the meltwater which usually contain wind-borne mineral particles. The peat in these fens is seldom more than about 20 cm thick, and is composed of sedge and moss remains (Tarnocai and Zoltai, 1988).

If the thermal balance of a portion of a peat plateau changes enough for the permafrost to thaw, the surface of the peat plateau drops, forming a 'collapse scar' fen. These are usually a few tens of square metres in size, and are characterized by dead trees protruding from the fen, and a cover of partially submerged *Sphagnum riparium*.

2.4 *DYNAMICS OF PERMAFROST PEATLANDS*

A unique interplay of climate, hydrology and other environmental factors leads to permafrost and ice formation in arctic and subarctic peatlands. How the development of these peatlands has been influenced by presence of permafrost and other factors is not well understood. Previous paleoecological studies of permafrost peatlands have provided some insight, although most have been restricted to analysis of a single core from a peatland (e.g. Ovenden, 1982; Couillard and Payette, 1985; Eisner, 1991; Wang and Guerts, 1991).

According to Zoltai *et al.* (1988), in the subarctic and low arctic, wetlands often develop in depressions, then go through a series of developmental stages which may include open water, marsh, fen, and bog. The basal deposit in many peatlands is a thin (1-15 cm), well-humified organic mud that may be mixed with mineral soil, typical of wet *Carex-Eriophorum* meadows. This may be overlain by a fen peat, with various proportions of *Carex* spp., *Drepanocladus* spp., and the remains of other fen species. In some cases, the fen peat rests on detrital, organic lacustrine deposits, indicating the infilling of a pond. The fens may evolve into peat plateau bogs with shrub and trees, but repeated reversion to open fens are sometimes apparent in the peat stratigraphy (Zoltai and Tarnocai, 1988).

Ovenden (1982) analyzed pollen, plant macrofossils and matrix composition of a 221 cm core from a polygonal peatland in northern Yukon, and found evidence of a hydroseral succession of wetland communities in the early

Holocene, beginning with a submerged vegetation followed by a marsh, fen, and eventually *Sphagnum-Ledum* bog. She attributed a transition to a wetter *Sphagnum balticum- Andromeda* carpet at 9600 BP to the formation of permafrost and polygonal ice wedges. This community persisted until approximately 3000 BP when the polygon became high-centred and peat growth declined. Zoltai and Tarnocai (1975) also linked the transition from sedge to *Sphagnum* dominated vegetation to the initial invasion of peatlands by permafrost.

Couillard and Payette (1985) documented the development of a peat plateau bog complex in northern Quebec by plant macrofossil analysis and radiocarbon dating, and found that a minerotrophic herbaceous vegetation originally colonized the depression, but the peat plateau began forming 1000 years after the establishment of the wetland, following a fen stage. The internal structure of the peat suggested that permafrost developed at a later stage, and palsas within the peat plateau had developed only in the last 700 years.

Eisner (1991) conducted pollen analysis on a peat core from the North Slope of Alaska to determine the usefulness of such studies for reconstructing regional vegetation and climate change. She concluded that while a regional signal was obtainable, a transect of cores allowing correlation of changes in hydrology, vegetation and climate would be more useful for deciphering both the climate record and the history of the peatland.

Zoltai (1995) reported stratigraphic and macrofossil analysis results of a total of 161 radiocarbon dated cores of peatlands in west-central Canada, and used the reconstructed paleoenvironments to indicate the presence or absence of permafrost at the time of peat formation. He concluded that in areas where permafrost is discontinuous today, most peatlands were fens without permafrost at 6000 BP, but in the continuous permafrost zone of the Arctic, permafrost was present at or before this date, and peat accumulation occurred under permafrost conditions. Permafrost development in the fens of the more southerly areas was associated with the development of a *Sphagnum*-dominated surface on the fens, caused by the onset of a cooler and moister climate. The insulation provided by surface peat layer and associated tree cover initiated permafrost development in small lenses that coalesced into large permafrost bodies according to the prevailing climatic conditions.

The time of permafrost development in peatlands cannot often be determined in sufficient detail to relate it to climate events or changes. At 6000 BP, permafrost was present in some peatlands, but distribution zones were shifted 300-500 km to the north, relative to the present zonation (Zoltai, 1995). Zoltai estimates that this corresponds to a mean annual temperature that was about 5°C warmer than at present.

Accumulation rates in frozen peatlands are generally lower than in non-frozen wetlands (Ovenden, 1990). In the western subarctic and low arctic, peat accumulation in many cases virtually ceased after permafrost elevated the peatland, which consequently became dry at the surface. Available dates from such peatlands show old ages for near-surface peat. For example, peat from 5.5-8.5 cm depth in a polygonal peat plateau bog was dated at 1145 ± 65 BP (Ovenden, 1982); peat from 35 cm depth in another polygonal peat plateau bog was dated at 2710 ± 60 BP (Zoltai and Tarnocai, 1975); and the age of peat from a depth of 30 cm at Natla River, N.W.T., was 3000 ± 50 BP (MacDonald, 1983). Zoltai *et al.* (1988) point out that these ages give minimum dates for permafrost development in these peatlands, and it is possible that widespread permafrost formation and aggradation took place after a general cooling of the climate around 5000 BP (Ritchie *et al.*, 1983).

3. REGIONAL SETTING AND HOLOCENE ENVIRONMENTAL HISTORY

3.1 *REGIONAL SETTING*

3.1.1 Geology and Physiography

The entire study area falls within the Arctic Coastal Plain geological province, which is characterized by a shallow homocline of carbonates, shales and sandstones of Paleozoic and Mesozoic age, overlain by thick deposits of unconsolidated Quaternary and recent material (Bostock, 1970; Vincent, 1989). There are few bedrock outcrops in the region, with the notable exception of the Campbell-Dolomite Uplands south of Inuvik, which consist of a dome of argillite overlain by Devonian limestone and dolomite (Vincent, 1989). The most apparent influence of the structural geology on the landscape is the northeast-southwest alignment of the Eskimo Lakes and Liverpool Bay, which are believed to be an extension of the Aklavik Arch, a tectonic feature that extends from Bell Basin in the west, across the Mackenzie Delta to near Sitidgi Lake. However, the bedrock is overlain by thick deposits of Quaternary fluvial and deltaic silts, sands and gravels from an ancestral Mackenzie River, which flowed through a graben that extends from Campbell Lake through to the Sitidgi and Eskimo Lakes (Ritchie, 1984).

The most imposing feature of the landscape is the modern Mackenzie River Delta, a complex network of shifting channels and thermokarst lakes 65 km wide and 200 km in length. The topography of the entire study area is low and flat, with the highest points, south of Inuvik, less than 300 m above sea level. West of the Mackenzie Delta the surface rises sharply to the Mackenzie and Richardson Mountains.

The surficial deposits of the Tuktoyaktuk Peninsula consist of a complex of unlithified and poorly lithified sand, silt and gravel of glacial, glaciofluvial, fluvial-deltaic, lacustrine and eolian origin (Vincent, 1989). The entire peninsula is flat and low-lying (less than 60 m above sea level) with the only significant relief resulting from several hundred closed system pingos (Mackay, 1979). The interior is poorly drained with a high density of shallow thermokarst lakes and wetlands. On most areas of the peninsula, lakes cover 30-50% of the surface (Mackay, 1963, p. 98). Significant areas are covered by deposits of peat 1-2 m thick, which are especially common in areas of lacustrine and intertidal deposits (Mackay *et al.*, 1983). Ice-wedge polygon development is abundant in both mineral and organic terrain. Vegetation covered earth hummocks, which develop in areas of extensive thermokarst activity, are also common (Rampton, 1988).

The Kukjuk peatland falls within the Involuted Hills physiographic section described by Mackay (1963, p. 138) as resembling, on aerial photographs, "the wrinkled skin of a well-dried prune." The 'wrinkles' are branching ridges, up to several hundred metres in length, 10s of metres in width, with individual ridges 5-10 m in height and a total relief of 30-50 m. These involutions are caused by the *in situ* development of extensive ground ice in fine-grained soil, followed by partial melting and slumping (Mackay, 1963).

The surficial deposits in the vicinity of the Kukjuk peatland consist of a complex of morainal and lacustrine deposits. The morainal deposits are till and associated gravel and sand deposited directly, or with minor reworking, by glaciers, probably during the Early Wisconsinan (see section 3.1.2). In most areas they have been modified by cryoturbation. The lacustrine deposits are of Quaternary age and include silt, sand, organic sediments and minor gravel, 2-8 m thick, deposited mainly in thermokarst basins (Rampton, 1988).

The southern part of the study area, including Inuvik and the Campbell Creek peatland area, is characterized by intermediate relief (40-300 m) with rolling, fluted and irregular topography resulting from Quaternary glaciations. The Campbell Creek peatland is within Mackay's (1963) Campbell Lake Hills Physiographic Region, which is a mainly upland area where bedrock is close to the surface and outcrops in escarpments. Both bedrock and morainal deposits are heavily fluted. Rapid changes of glacial flow patterns are shown by abrupt swings in the direction of fluting. The peatland is within the Campbell-Sitidgi Lake Lowland section of this region, which is part of the course of the former river which flowed through the Campbell-Sitidgi Lake depression. The area is low and flat, with the divide between Campbell and Sitidgi Lakes at about 10 m asl. Drainage is poor, and shallow lakes and wetlands are abundant. To the north and south of the lowland the surface rises gradually to about 250 m asl. Campbell Lake, to the immediate southwest of the peatland, occupies part of a glacially modified channel of probable tectonic origin, surrounded by rocky cliffs composed in part of Devonian limestone.

The surficial deposits in the Campbell Creek peatland area consist of Late Wisconsin lacustrine deposits, including silt, sand, and some gravel, deposited mainly through thermokarst during a high water phase of the Eskimo Lakes (Rampton, 1988). The uplands to the north and south have a cover of Late Wisconsin morainal deposits, comprised of a blanket (2-5 m thick) of till and associated gravel and sand deposited directly or with minor reworking by glacier ice, with some modification by cryoturbation (Rampton, 1988).

3.1.2 Climate

Climate normals for Tuktoyaktuk and Inuvik are summarized in Table 3.1. In general, the entire study area experiences long, very cold winters and short summers. The Tuktoyaktuk Peninsula has an arctic coastal climate, with late

Table 3.1. Climate data, Inuvik Airport and Tuktoyaktuk . Source: Canadian Climate Normals, 1961-1990 (Environment Canada, 1993). Data cover the period 1957-1990 for both sites.

	Inuvik Airport	Tuktoyaktuk
Temperature (°C)		
Daily mean	-9.5	-10.5
July mean	13.8	10.9
January mean	-28.8	-27.2
Annual degree days		
above 18°C	18	4
above 5°C	682	410
Precipitation (annual means)		
Rainfall (mm)	116	75.4
Snowfall (cm)	175.2	66.8
Total (mm)	257.4	142.1
Days with		
max. temp. >0°C	154	138
measurable pptn.	129	69

springs and prevalence of arctic air in summer. The transition between this climate zone and the continental subarctic zone, which includes the Inuvik-Campbell Creek area, roughly coincides with the ecological transition from forest to tundra. The Inuvik area experiences earlier springs, and summer weather is largely influenced by the cyclonic activity of the frontal belt, as opposed to the arctic air masses which dominate on the Tuktoyaktuk Peninsula (Ritchie, 1984). The most significant factor is perhaps the difference in growing season temperatures. Inuvik has a mean July temperature 3°C higher than Tuktoyaktuk, and also experiences considerably more growing degree days above 5°C and 18°C (Table 3.1). Precipitation is low in the entire area, but higher at Inuvik, with roughly half accounted for by winter snowfall at both locations.

The extreme maximum and minimum temperatures recorded at Inuvik from 1961-1990 are 31°C and -57°C, respectively. Tuktoyaktuk has a more maritime climate, with slightly warmer winters and cooler summers, and temperature extremes of 28°C and -50°C. The prevailing wind directions are northwesterly and easterly throughout the region (Environment Canada, 1993).

3.1.3 Vegetation

The arctic treeline crosses the study area just north of Inuvik, separating the two major ecological complexes of the region, the boreal forest and the tundra, each with its distinctive diversity of flora. The low arctic tundra of the lower twothirds of the Tuktoyaktuk Peninsula (including the vicinity of the Kukjuk peatland), is characterized by a variety of low shrubs, predominantly *Betula glandulosa* and *Ledum decumbens,* as well as *Arctostaphylos alpina, Empetrum nigrum* ssp. *hermaphroditum, Vaccinium vitis idaea* var. *minus, V. uliginosum, Rubus chamaemorus, Salix reticulata* and occasionally *Salix glauca* on moderate to well-drained sites (Ritchie, 1972, 1984). Moderate to poorly drained areas generally have a cover of *Eriophorum vaginatum, E. angustifolium* and a variety of *Carex* sp., while wetlands are covered by sedge meadows, with *Carex* sp., *Eriophorum Scheuchzeri,* the mosses *Drepanocladus* and *Sphagnum* and some *Betula glandulosa* and other shrubs. Rare patches of krummholz *Picea glauca* and *P. mariana* occur, as well as clumps of *Populus balsamifera* which are probably clonal (Ritchie, 1984).

The forest-tundra in the Inuvik area is dominated by open stands of spruce, mostly *Picea glauca* on uplands and *P. mariana* in wetlands. Where *P.mariana* grow on severely cryoturbated soils they are usually tilted by frost heaving of the ground (Ritchie, 1977). *Betula papyrifera* and *Larix laricina* are also common, along with local stands of *Poplar balsamifera*. *Alnus crispa, Betula glandulosa* and *Salix* spp. are important shrubs, as are a variety of heath, especially *Vaccinium vitis-idaea, V. uliginosum , Ledum decumbens, L. groenlandicum, Arctostaphylos rubra* and others on peatlands.

Northward from Inuvik on the uplands, the subarctic forests gradually become restricted to small clumps of *Picea mariana* or *P. glauca* growing in sheltered locations in an increasingly dominant tundra. The actual limit of *Picea glauca* can be as much as 50 km north of treeline. Spear (1983, 1993) reported several patches of spruce krummholz on the Tuktoyaktuk Peninsula near Black Ice Lake, roughly 40 km north of treeline, and Kittigazuit, 60 km beyond treeline. These krummholz patches are extremely rare and of low density (Spear, 1993). *Larix laricina* reaches its northern limit in the vicinity of the Inuvik airport (Ritchie, 1977).

3.1.4 Wetlands

The Tuktoyaktuk Peninsula is in the Low Arctic wetland region, while the Inuvik area falls within the High Subarctic region (Fig. 3.1; National Wetlands Working Group, 1988). Wetlands are widespread, particularly on the peninsula where low- and high-centred polygons, fen marshes and shallow water are the most common types (Tarnocai and Zoltai, 1988). A large percentage of the land area is covered with wetlands and saturated soils because moisture from incoming precipitation and snowmelt is retained in the active layer as a result of the underlying permafrost acting as a barrier to water movement. A *Carex-Eriophorum* vegetation is dominant in wetlands of this region, with various species of *Sphagnum* also abundant.

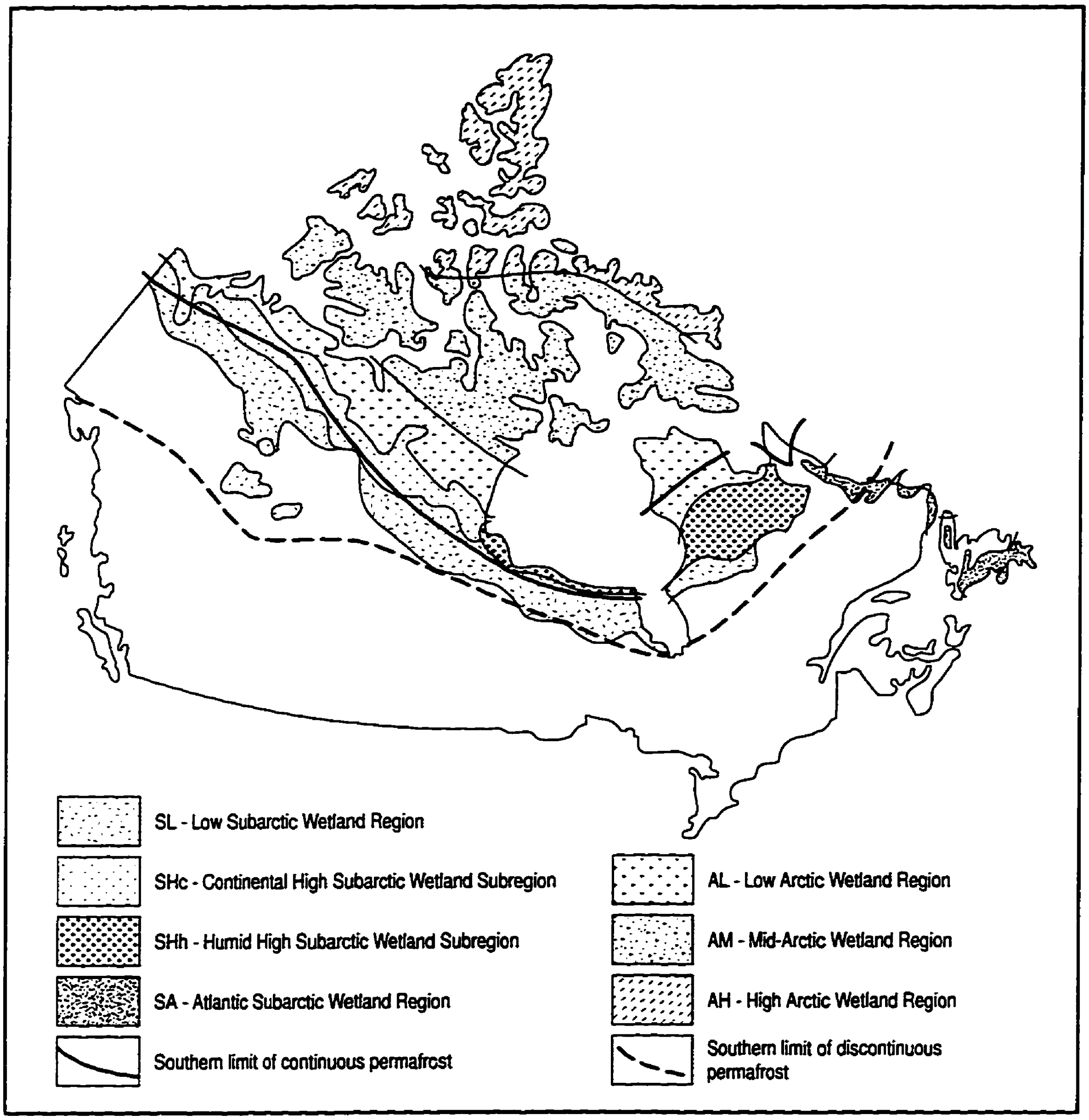


Figure 3.1. Canadian arctic and subarctic wetland regions (after National Wetlands Working Group, 1988). Southern boundaries of continuous and discontinuous permafrost are also indicated (after Brown, 1970).

Wetlands are common in Tuktoyaktuk Peninsula part of the Low Arctic wetland region, with low- and high-centre polygons, fens, marshes and shallow lakes being very numerous (Zoltai and Tarnocai, 1988).

3.1.5 Permafrost

Permafrost underlies most of area, with the exception of large, deep water bodies. The depth of permafrost in the Tuktoyaktuk area is estimated at greater than 400 m in areas that have not recently been covered by lake waters or whose surface has not had a mean annual ground temperature of greater than -8°C in the recent past (Rampton and Bouchard, 1975). The southern limit of continuous permafrost passes to the south of the study area at approximately 67.5°N (Figure 3.1). Large bodies of massive ground ice are widespread, especially on the Tuktoyaktuk Peninsula. Some of these ice bodies may be buried glacier ice but most probably developed as segregated ice after deglaciation (Rampton, 1974a; Mackay 1986; Vincent, 1989).

3.2 *QUATERNARY HISTORY*

3.2.1 Wisconsinan Glaciations

The region was repeatedly invaded during the Quaternary by continental glaciers, probably originating west of Hudson Bay. However, the timing and extent of the glaciations is uncertain. Records of Wisconsinan glaciations in the area come from morainal deposits and drift erratics that provide evidence of the extent of cover and the glacial limits, and from stratigraphic evidence found in exposures or cores. Unfortunately, there are still too many gaps in both types of information to establish a definite glacial history.

While some Quaternary geologists propose that the strongest advance was during the Early Wisconsinan and that Late Wisconsinan ice was less extensive (Figure 3.2), others contend that the maximum Wisconsinan advance occurred in the Late Wisconsinan, and was followed locally by a readvance. If this latter scenario is accurate, then the limit shown in Figure 3.2 as Early Wisconsinan becomes the Late Wisconsinan limit and the illustrated Late Wisconsinan limit is considered to represent a readvance known as the Tutsieta Lake Phase (Hughes, 1987). Vincent (1989) has reviewed the evidence for both arguments in detail and concluded that the Early Wisconsinan maximum is the most likely scenario, but notes that the evidence available to date is inconclusive.

Most of the lower Mackenzie River region was glaciated during Early Wisconsin ice expansions. East of the Mackenzie Delta, the limit of glaciation can be traced along the Tuktoyaktuk Peninsula and the western edge of the uplands east of the Anderson River (Mackay *et al.*, 1972; Heginbottom and Tarnocai, 1983). Rampton (1988) assigned this ice limit to the Toker Point Stade. There is substantial evidence that it is of Early Wisconsinan age. The only till in the Tuktoyaktuk area is early Wisconsin or pre-Wisconsin in age, according to Mackay *et al.* (1972; also Rampton and Bouchard, 1975; Rampton, 1981, 1988).

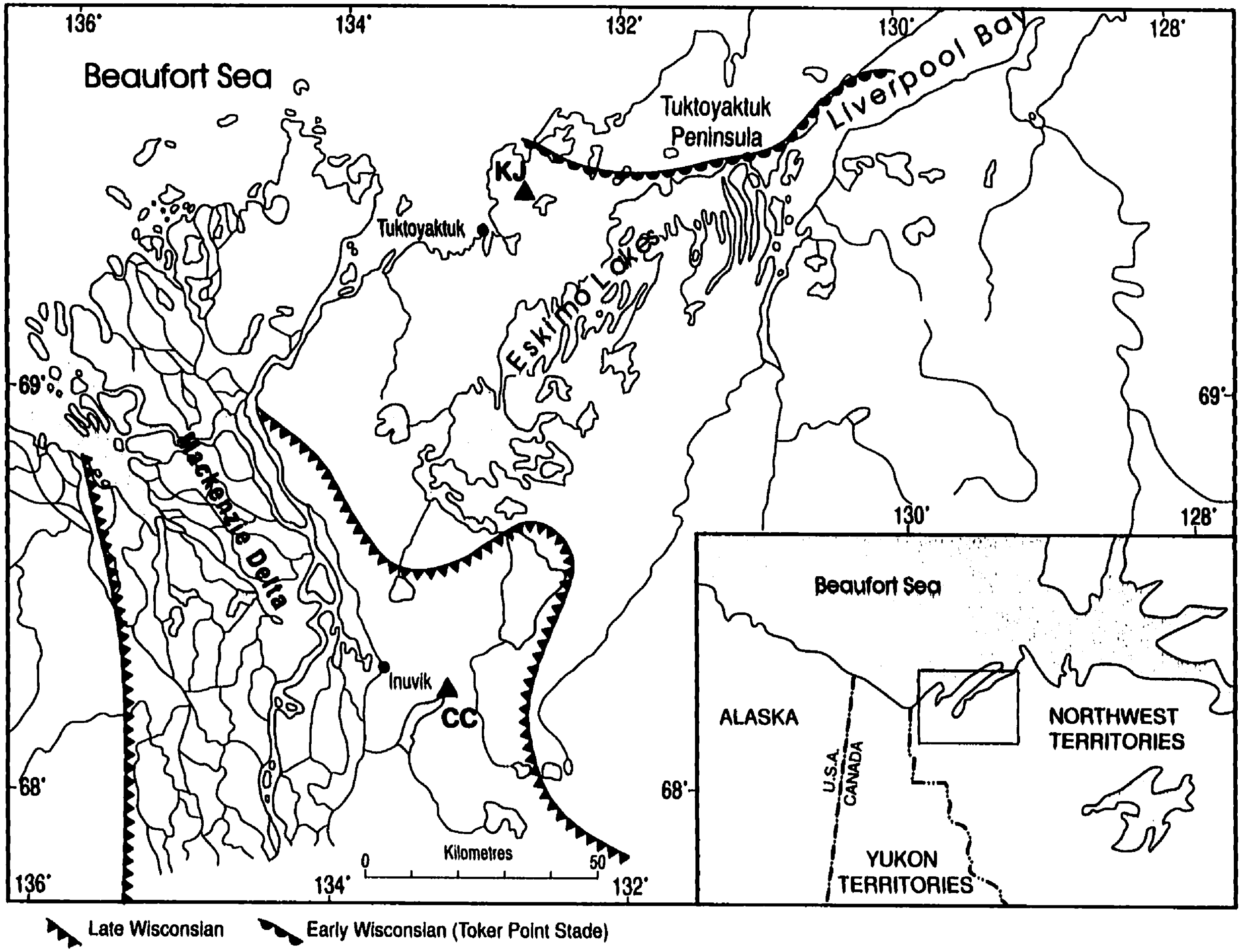


Figure 3.2. Proposed glacial limits in the study area, after Vincent (1989).

Wood from glaciofluvial sand at Ibyuk Pingo, near Tuktoyaktuk, was found to be beyond the age of radiocarbon dating (Fyles *et al.*, 1972). This lower glaciofluvial sand is the lowest deposit exposed at Tuktoyaktuk, underlying the only till on this part of the Tuktoyaktuk Peninsula. The regional extent of this sand or similar sand units indicate that much of the area was a large outwash plain during their deposition (Vincent, 1989).

Dates on shell fragments deposited when the area was isostatically depressed during deglaciation have also been used to indicate that the advance was pre-Late Wisconsinan (>35,000 BP (GSC-562) on Garry Island and >37,000 (GSC-690) on Kendall Island, [Rampton, 1988]; 43,550 ± 470 BP (TO-796) on Garry Island and 48,000 ± 1100 BP (RIDDL-801) from south of Kendall Island, [Vincent, 1989]). In addition, many Late Wisconsinan dates from sediments above Toker Point Stade deposits are available, the oldest being 17,860 ± 260 BP (GSC-481) for *in situ* peat overlying till that was tilted during growth of the Ibyuk Pingo near Tuktoyaktuk (Vincent, 1989).

The proposed limit of Late Wisconsinan glaciation is also shown in Figure 3.2. Ice flowed westward and northwestward from the Keewatin sector of the Laurentide Ice Sheet (Vincent, 1989). According to this reconstruction, Late Wisconsinan ice was confined mainly to a trench underlying the Mackenzie Delta and adjacent lowlands. The modern Mackenzie Delta has formed since the retreat of Late Wisconsin ice. West of Aklavik, the Late Wisconsin ice limit is at approximately 90 m (Heginbottom and Tarnocai, 1983). An ice margin with many well-developed terminal moraines is found east of the Mackenzie River. For part of its length this margin coincides with the Tutsieta Lake Moraine, which was named and assigned to the Tutsieta Lake Phase by Hughes (1987), who did not agree that this is the Late Wisconsinan limit, but considered it a still-stand during the retreat of Laurentide ice in Late Wisconsin time (also Hughes *et al.*, 1981). The Late Wisconsinan limit would then have been at the all time glacial margin, at what Rampton (1988) considered the Early Wisconsinan Toker Point Stade limit on the Arctic Coastal Plain. After reaching its limit, Late Wisconsinan ice receded to the southeast, leaving ice marginal features such as major meltwater channels or moraines like the Tutsieta Lake Moraine.

As a potential solution to the controversy Vincent (1989) proposes three possible Wisconsinan advances, with the oldest (Early Wisconsinan) corresponding to the Toker Point Stade and the youngest to the Tutsieta Lake Phase being "late" Late Wisconsinan. He suggests that an intermediate advance may have occurred late in the Middle Wisconsinan or early in the Late Wisconsinan.

Based on radiocarbon dates from lakes, the entire study area is believed to have been ice-free for at least the last 13,000 years (Ritchie *et al.*, 1983; Ritchie, 1984; Rampton, 1988), and possibly as early as 15,500 ± 440 BP (GSC-3646, Blake 1987; Vincent, 1989). If the limits shown in Figure 3.2 are correct, then the southwestern portion of the Tuktoyaktuk Peninsula, including the Kukjuk peatland site, was glaciated during the Early Wisconsinan, but not during the Late Wisconsinan, while the Inuvik-Campbell Creek area experienced both glaciations.

3.2.2 Holocene Environmental History

Pollen, macrofossil and radiocarbon results have been published for a number of lake sites within the study area and give clear evidence of a northward displacement of treeline and warmer than present climate during the early Holocene. Corroborating changes have been recorded in several sites with detailed records spanning up to 13,000 years: Tuk-5 (Ritchie and Hare, 1971), Sleet Lake (Spear, 1983), and Bluffer's Pingo (Spear, 1993) on the Tuktoyaktuk Peninsula; and M Lake (Ritchie, 1977), SW Lake (Ritchie, 1984b), Reindeer Lake (Spear, 1993), and Kate's Pond (Ritchie, 1987) near or south of treeline in the Inuvik-Eskimo Lakes area (Figure 3.3). Pollen analysis of several other radiocarbon dated but incomplete sections of organic sediment from the region provide further evidence (Ritchie, 1984a).

Ritchie (1984a) identified three main biostratigraphic units in the area based on the paleoecological records from these sites: 1) *Betula-Salix-Shepherdia* (13,000 to 9500 BP); 2) *Picea* (9500 to 6000 BP); and 3) *Betula-Alder*-ericad (6000 BP to present). In addition to the pollen records, the discovery of *Picea* macrofossils from at least five sites on the Tuktoyaktuk Peninsula (all between 69°07'N and

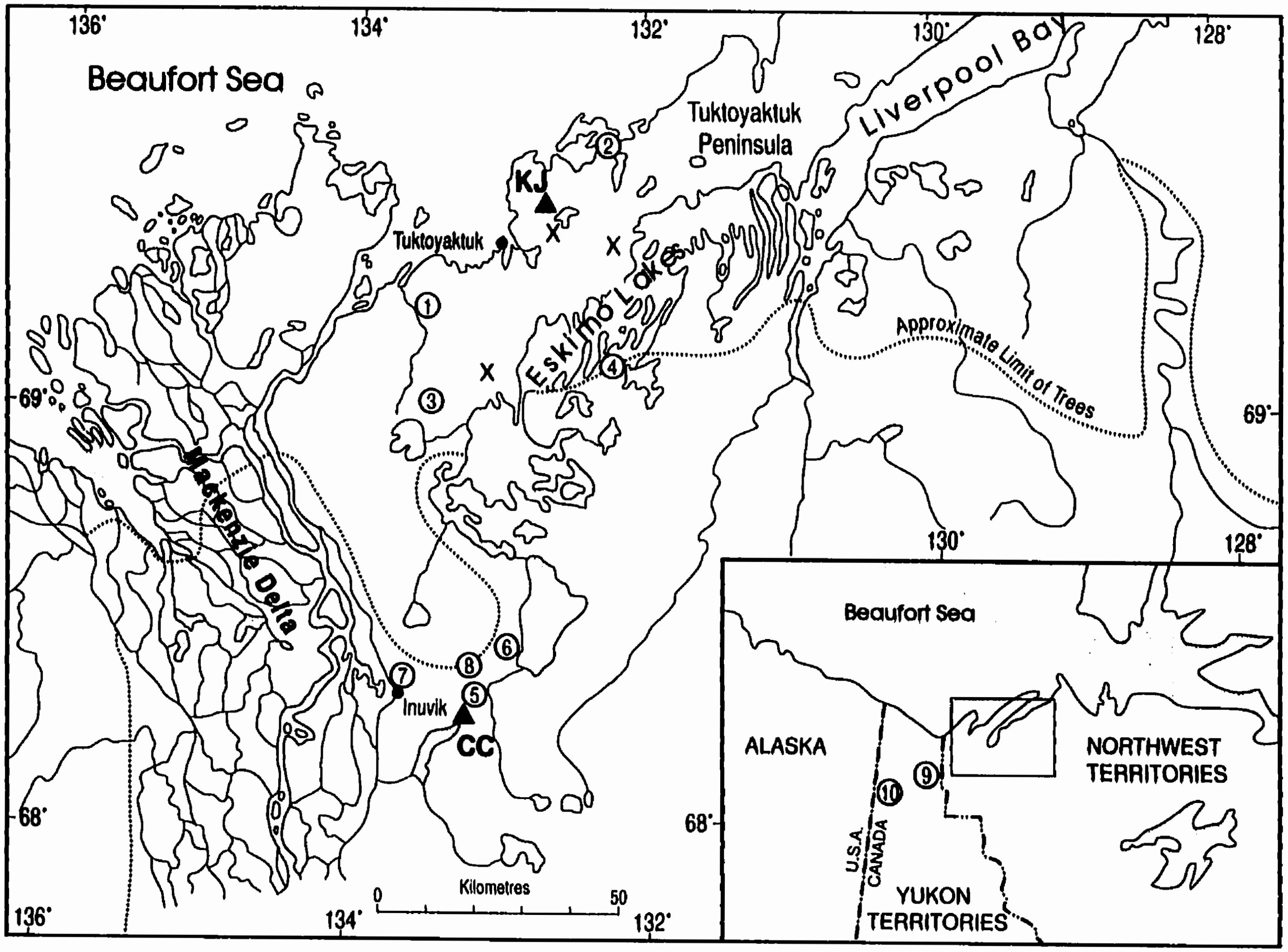


Figure 3.3. Sites mentioned in text: 1) Sleet Lake (Spear, 1983); 2) Bluffer's Pingo (Spear, 1993); 3) Tuktoyaktuk-5 (Ritchie and Hare, 1971); 4) Reindeer Lake (Spear, 1993); 5) M-Lake (Ritchie, 1977); 6) Twin Tamarack Lake (Ritchie, 1985); 7) Twin Lakes peat deposit (Mackay and Terasmae, 1963); 8) Twin Lake Hill (Ritchie, 1984); 9) Hanging Lake (Cwynar, 1982); 10) Polybog ((Ovenden, 1982). X = *Picea* macrofossils (see Table 3.2).

69°28'N; 132° and 134°W), ranging in radiocarbon age from 5000 to 6600 BP, and frequent *Picea* seeds and needles in the Sleet Lake sediments from 9800-5900 BP, provide evidence for a strong presence of *Picea* trees on the peninsula during the early Holocene (Figure 3.3, Table 3.2). The northern limit of trees was as much as 70 km north of its present location during the early Holocene (Spear, 1983) and reached at least the southern Banks Island during the Early Pleistocene (Kuc, 1974; Vincent, 1984; Matthews *et al.*, 1986; Vincent 1989). A summary of the postglacial vegetation changes reconstructed from lakes sediment studies in the Mackenzie Delta, and the climate changes inferred by Ritchie (1984a; also 1972, 1977, 1985a,b; Ritchie and Hare, 1971; Ritchie *et al.*, 1983;) and Spear (1983, 1993) is presented in Table 3.3.

According to Ritchie's (1984) vegetation reconstruction for the Tuktoyaktuk Peninsula, a dwarf *Betula*-dominated tundra with frequent local *Shepherdia* and *Salix* -sedge carr or marsh vegetation in lowlands existed between 13,000 and 9200 BP, with localized herb tundra on xeric sites. A sharp *Picea* pollen rise around 9000 BP in the Tuk-5 and Sleet Lake sites is interpreted as representing a change to a continuous coniferous woodland on all upland sites, dominated by *Picea* with local occurrence of *Populus, Larix,* and later, tree *Betula* (Ritchie and Hare, 1971; Spear, 1983). At Sleet Lake, influx of *Picea* pollen rose sharply from low values at about 10,000 BP to maximum values between 9000 and 8000 BP, and declined steadily between 8000 and 6000 BP. Needles in Sleet

Table 3.2. Locations of *Picea* macrofossils north of present treeline on the Tuktoyaktuk Peninsula (after Spear, 1983 and Ritchie, 1984)

Location	Reference	Material	Radiocarbon age
69°07'N, 133°16'W	Ritchie & Hare, (1971)	stump	4940±140, GSC-1265
69°28'N, 132°35'W	Blake (1976?) (Collected by Burden *et al* 1975; pers.comm. 1994)	twigs	5945±100
69°16'N, 132°20'W	Delorme et al. (1977)	wood & cones	5700±100, BGS-215
69°16'N, 132°20'W	Spear (1983)	log	6380±70, GSC-3242
69°27'N, 132°10'W	Spear (1983)	stump	6620±70, GSC-3239

Table 3.3. Summary of regional environmental history as reconstructed based on lake sediment studies. After Ritchie (1984) and others (see text).

VEGETATION HISTORY	INFERRED CLIMATE
No significant change since approximately 4500 BP	Only minor fluctuations in the last 4500 yr
Treeline retreated to present position 4500-5000 BP	Rapid cooling to conditions similar to present
Alnus rise 5700-7800 BP	Gradual cooling after approximately 8000 BP
Forest with *Picea* and *Larix* 75-100 km north of present limit, 5000-9000 BP	Climate warmer and more humid than present during this interval
Picea present on and near Tuktoyaktuk Peninsula by 10,900-9000 BP, and *Typha* common 10,000-9000 BP	Estimated summer temperature approximately 3°C warmer than present (up to 6°C warmer based on *Typha* presence); summer precipitation also higher than present
Sharp increase in *Betula* pollen influx 11,500-11,000 BP	Rapid climate warming begins
Herb tundra 15,000-11,500 BP	Deglaciation followed by gradual warming

Lake sediments dating from 5700 BP indicate that *Picea* was still growing in close proximity to the basin. At 4500 BP there was a sharp decrease in *Picea* pollen to modern values (Spear, 1983). The woodland was replaced by a transitional forest-tundra, with increases in *Alnus* and dwarf shrubs from 5000-3500 BP. By 3500 BP, shrub tundra had developed around Sleet Lake, as evidenced by an increase in the percentages of Ericales and herb pollen and the decrease in *Picea* pollen. A dwarf *Betula*-Ericales-sedge-*Eriophorum* tundra has persisted since about 3500 BP, with *Picea* absent except for rare krummholz (Spear, 1983, 1993).

Farther northeast on the Tuktoyaktuk Peninsula, no dramatic vegetational changes are recorded in the pollen record from the Bluffer's Pingo site, which spans the last 9000 years. The tundra vegetation probably has not changed much during that period. The only evidence from this site of potential climate change is the formation of the pingo itself at approximately 2500 BP (Spear, 1993).

Several sites in the uplands of the Inuvik area have provided a consistent pattern of Late Wisconsin-Holocene vegetation change. A representative site is Twin Tamarack Lake, a small shallow pond near Inuvik studied by Ritchie (1985). Non-arboreal pollen types, including Gramineae, *Artemisia*, *Salix*, dwarf *Betula* and a diverse assemblage of herbs, dominate the basal pollen zone (14,500 to 11,800 BP), where pollen accumulation rates of all taxa are low. Ritchie interpreted this as representing a herb tundra on uplands with Cyperaceae-Poaceae-*Salix* marsh communities in the lowlands. While dwarf *Betula* spp.

dominated the early tundra on the Tuktoyaktuk Peninsula, it was relatively scarce in the Inuvik area, probably because the limestone that dominates the Campbell Hills tends to inhibit the establishment of dwarf *Betula* (Ritchie, 1984). Between 11,800 and 8400 BP this assemblage was replaced by a woodland community dominated first by *Populus*, with a brief (500 year) interval of *Juniperus* and *Populus*, and then a transition to coniferous woodlands as *Picea* spp., which had slower migration rates, displaced the *Populus* around 9000 BP. *Picea mariana* apparently expanded more slowly than *P. glauca*, becoming more abundant about 7000 BP when paludification and permafrost aggradation had increased, forming extensive wetland habitats (Ritchie, 1989). An increase in alder to its modern abundance around 6000 BP completed the establishment of modern boreal woodlands.

Ritchie (1985, 1989) explained the early to mid-Holocene expansion of *Picea* woodlands in this region in terms of an increase in summer warmth to maximum values at 10,000 BP, possibly as a result of an early Holocene Milankovitch insolation maximum, with summer insolation theorised to have been up to 9-10% greater than at present (Ritchie *et al.*, 1983; Ritchie, 1984). However, Hill *et al.* (1985) have shown that sea-level was approximately 40 m lower during the early Holocene than it is today, and the Tuktoyaktuk area was then approximately 100 km farther inland, comparable to the present distance between Inuvik and the sea coast (Pelletier, 1987; Mackay, 1992). Proximity to the coast could have had an important role in making the climate suitable for *Picea* growth. Based on Black and Bliss's (1980) study of the autecology of *Picea mariana* in the Mackenzie Delta area, it is estimated that mean July temperature must have been 3°C warmer than present (Ritchie, 1984). Ritchie also infers mean annual temperature of 3°C greater than present, as well as an increase in mean annual total degree days above 5.5°C by 200, and mean annual precipitation up to 12 cm greater than present, as derived from the modern differences between Tuktoyaktuk and Inuvik. The highest estimate of early Holocene maximum mean annual air temperature for the western Arctic coast is 10°C (Delorme *et al.*, 1977). Although a cooling trend started at about 8000 BP, warmer than present conditions lasted until about 4500 BP (Ritchie 1984). The period of modern ice-wedge growth probably started about that time (Mackay, 1992).

There is evidence of coincident periods of warmer than present climate and northward advance of treeline in other high latitude areas during the early Holocene. The northern Yukon, which has been ice-free for at least the last 30,000 years, had a herb tundra vegetation until about 14,000 BP, when an increase in dwarf *Betula* indicates a shift to shrub tundra and an increase in temperature and precipitation (Cwynar, 1982). Between 11,000 and 9000 BP, this area was much warmer, with treeline expanding northward. As in the Mackenzie Delta area, total pollen influx in sites from this area rise abruptly at about 10,000 BP. Forests of *Picea glauca* (9500-6500 BP) and later *P. glauca* and *P.*

mariana (6500-5000 BP) extended into the valley of the Blackstone River in the Southern Ogilive Ranges, Yukon Territory, a region that is now in shrub tundra (Cwynar and Spear, 1991). Treeline started to retreat southwards by about 5500 BP, and by 4000 BP climate had apparently cooled substantially (Ritchie *et al.*, 1983).

Evidence of warmer than present conditions in northern Alaska in the early Holocene includes increased eolian activity (Carter and Hopkins, 1982), melting of ice wedges (McCulloch and Hopkins, 1966), range extensions of *Typha latifolia* and *Populus balsamifera* (McCulloch and Hopkins, 1966; Hopkins *et al.*, 1981; Edwards and Brubaker, 1986), peaks in *Populus* and *Juniperus* pollen (Ager and Brubaker, 1985), and localised peaks in *Picea* pollen (Brubaker *et al.*, 1983). However, unlike in north-western Canada, the pollen data in Alaska do not show much evidence of northern expansion of tree species beyond their present limits. Brubaker *et al.* (1983) concluded that in north-central Alaska, coniferous treeline never existed beyond its current position. There is, however, some evidence that *Populus* extended into some areas of the Alaskan tundra during the early Holocene (Edwards *et al.*, 1985; Edwards and Brubaker, 1986; Anderson *et al.*, 1988; Barnosky *et al.*, 1987).

In Scandinavia, macrofossil and fossil tree evidence suggests that early to mid-Holocene climate was warmer than present, with *Pinus sylvestris* expanding to maximum altitudinal and latitudinal maxima by 8000 BP, and retreating to modern treeline by about 4000 BP (Kullman, 1992, 1993; Hyvärinen, 1993). More northerly treeline position and warmer than present climate has also been inferred for parts of north-western Russia and central Siberia between 9000 and 3500 BP (Khotinskiy, 1984).

In other parts of northern Canada, where deglaciation did not occur until after 10,000 BP, the period of maximum Holocene warmth occurs significantly later than in the Yukon and Mackenzie Delta region. In the Canadian high Arctic, abundant driftwood on raised beaches, dating to between 6500 and 4500 BP, is attributed to the presence of more open water than at any other time in the Holocene, probably due to a Middle Holocene warm interval (Blake, 1972). Andrews *et al.* (1981) concluded that July mean temperatures were well above present values from 5500 to 2250 BP, and probably longer.

On the mainland, most areas were occupied by tundra for a brief period following deglaciation. Based on pollen and other evidence, treeline is believed to have expanded beyond present limits between 5500 and 3500 BP in Keewatin (Nichols, 1967) and between 5000 and 3500 BP north of Yellowknife (Moser and MacDonald, 1990).

Alnus increased throughout north-western Quebec after 6000 BP, then declined around 3800 BP, possibly in response to a forest-tundra environment becoming more dense (Gajewski *et al.*, 1993). Maximum *Picea* density occurred between 3800 and 2000 BP in northeastern Canada, then declined, probably as a result of climatic deterioration and a failure to regenerate after fires (Payette and Fillion, 1985; Gajewski *et al.*, 1993;). Total pollen influx in the New Quebec-Labrador region, represented by a profile from Lac des Roches Moutonées, northern Quebec (McAndrews and Samson, 1977) peaked about 4000 BP, reflecting the northward migration of spruce dominated forest in response to regional climatic warming.

The mid-Holocene climatic optimum suggested for these regions occurred too late to be attributed directly to Milankovitch forcing. The persistence of ice in these regions into the early to mid-Holocene would have retarded warming in those areas. Deglaciation was followed immediately at many sites by organic deposition and immigration of trees and shrubs, implying that the early Holocene temperature gradient was especially steep in this region.

MacDonald *et al.* (1993) attributed the asynchronous advance of treeline in central Canada to a shift in the summer position of the Arctic Front caused by changes in frontal wave characteristics, whereby a northward displacement of the Arctic front can result in a southward shift of the front in areas in central and eastern Canada.

4. METHODS

4.1 FIELD METHODS

4.1.1 Site selection

Two peatlands were chosen for detailed study. One peatland was selected from north of the present limit of trees, on the Tuktoyaktuk Peninsula, and one from south of the treeline in the Inuvik area. Potential sites were identified using aerial photographs, and final selection was made after reconnaissance from the air (on the Tuktoyaktuk Peninsula) and in the field. Most peatlands on the Tuktoyaktuk Peninsula are characterized by low- or high-centre polygons. Because high-centre polygons are thought to evolve from low-centre polygons (Zoltai and Tarnocai, 1975), it seemed likely that a high-centre polygonal peatland might yield a longer record. The Kukjuk peatland was considered ideal because it has well-developed high-centre polygons over much of its surface, and is raised approximately 1 m above the local water table, so that most of its surface was dry enough to make coring possible. It is a small but complex peatland with a variety of microtopographic and vegetation units that might be expected to yield significantly different stratigraphic records. The larger

Campbell Creek peatland, in the forest-tundra 20 km southeast of Inuvik, also offered the opportunity to obtain cores from a variety of microhabitats, including *Sphagnum fuscum* hummocks; raised, lichen-covered flat areas, wet hollows filled with *Sphagnum* spp. and sedges, and exposures at the edge of a pond. Several cores were collected from representative habitats on each of the peatlands.

4.1.2 Sampling

At each coring site, monoliths approximately 30 by 30 cm square were carefully cut by hand from the unfrozen surface layers of the peat, wrapped in plastic, and packaged in individual wooden crates. The peat from the permafrost table downwards was then cored using a modified motor-driven CRREL auger with a 7.5 cm diameter core barrel (Veillette and Nixon, 1980). Cores were raised in 5 to 40 cm long segments which were scraped clean of possible contamination. A preliminary stratigraphic description of the segments were recorded before they were wrapped in plastic and aluminum foil. Cores were kept frozen in the field in large coolers which were transported for storage in freezers at the PCSP base in Tuktoyaktuk or the Inuvik Research Centre. The cores were kept frozen during transport back to the Wetlands Laboratory at the University of Waterloo, and stored in a deep-freeze until subsampling was conducted.

Samples of the most common plant species on the peatlands and surrounding uplands were collected to provide modern reference material for macrofossil identification.

4.2 LABORATORY METHODS

4.2.1 Subsampling and physical analyses

While still frozen, the cores were sawed by hand into 1 cm slices, except the ice layers, which were cut in 2-5 cm long segments. Subsamples were extracted from the peat slices for bulk density, loss-on-ignition, pollen and plant macrofossil analysis and radiocarbon dating. The stratigraphy of each core was described by noting changes in color, composition and ice characteristics. Representative samples from each stratigraphic unit in most cores were examined and described using the modified Troels-Smith system (Aaby and Berglund, 1986). Basal organic sediments from all cores and samples from levels of significant changes in the cores used for pollen and macrofossil analysis were radiocarbon dated at the University of Waterloo Radiocarbon Laboratory.

Bulk density and loss-on-ignition analyses were performed to estimate content of moisture, organic carbon and inorganic carbon. Using a brass syringe calibrated at 0.91 cm^3, two plugs (1.82 cm^3) of sediment were taken from contiguous 1-2 cm slices in the cores used for pollen and plant macrofossil analyses, and at 1 to 6 cm intervals in the other cores. After weighing to determine fresh weight, the samples were dried overnight in an oven at 100°C, and reweighed to allow calculation of moisture content and dry bulk density (Bengstton and Enell, 1986). The dried samples were then placed in a furnace at 550°C for 2-4 hours for ignition of organic matter, and weighed again to determine ash weight so that percentage (by dry weight) of organic matter could be calculated. Finally, they were ignited at 1000°C for 1 hour and again reweighed. The mass lost is used to estimate the inorganic content of the material (Dean, 1974).

4.2.2 Pollen analysis

For pollen analysis, subsamples 0.91 or 1.82 cm^3 in size were taken with the calibrated brass syringe at 3-12 cm intervals, which varied depending on ice content. Each subsample was weighed, inoculated with 40,500 *Lycopodium clavatum* spores (Stockmarr, 1971; three tablets of Batch 307862) and processed according to standard procedures as required (Faegri and Iversen, 1989). This involves treatment with a sequence of acid washes to remove the unwanted sediment matrix, except the pollen and spores. A 10% HCl wash was first used to dissolve the *Lycopodium* tablets and remove any carbonates. Samples were then boiled in 10% KOH for 5-10 minutes to deflocculate and remove humic acids and sieved through 250 μm nitex mesh to remove coarse organics. For mineral-rich samples, HF was used to remove siliceous material where necessary, before acetolysis treatment (sulfuric acid in acetic anhydride) to remove cellulose and hemi-cellulose. Where necessary, samples were treated with warm sodium pyrophosphate and sieved through an 8 µm mesh monofilament screen to remove clays (Cwynar *et al.*, 1979). Samples were stained with safranin, dehydrated with tertiary butyl alcohol and stored in silicone oil (2000 c.st.) in stoppered vials.

Pollen identification and counting was done under a binocular microscope at 400x magnification routinely, and under oil at 1000x magnification for critical identifications. Pollen and spores were identified with the aid of keys (Moore *et al.*, 1991; Andrew, 1984; McAndrews *et al.*, 1973; Warner and Chinnappa, 1986), and by comparison with modern reference material. A minimum total of 500 terrestrial pollen grains was counted, except in ice-dominated samples where pollen sums of at least 200 were obtained. At least one entire slide was counted for each sample; and in some cases up to 7 slides were required to reach the sum. Regularly spaced (2 mm) traverses of the entire cover slip were made to avoid bias that may result from differential movement of various types of pollen grains on the slides.

Pollen was identified to genus level in most cases, (except in the case of taxa such as Poaceae and Cyperaceae which are difficult to distinguish beyond family level with much confidence), and to species level whenever possible. *Betula* pollen grains were separated into grains less than 20 µm in diameter, taken to represent shrub birch, and larger grains which may represent arboreal birch (Ives, 1977) to aid in the identification of tundra and forest assemblages. This serves as an approximation only, since overlaps occur. An attempt was made to differentiate pollen of *Picea glauca* and *P. mariana* based on morphological characteristics (Hansen and Engstrom, 1985), and size (Birks and Peglar, 1980), although this was not always possible.

In the pollen diagrams, the prefix 'cf.' is used where identifications were uncertain due to poor preservation, and the suffix 'type' to indicate taxa where insufficient reference material was available for critical identification, but the given species is most likely based on ecological or phytogeographical likelihood. The 'unknown/indeterminable' category on the pollen diagrams includes all grains that are unrecognizable because of deterioration or crumpling, and all grains that are morphologically recognizable but not assignable to a known taxon. 'Undifferentiated' pollen types include those grains which belong to the genus or family indicated, but could not be confidently identified to further taxonomic levels.

Because this research is concerned with the development of local (peatland) vegetation, the pollen sum used for calculation of percentage frequencies of individual pollen taxa was comprised of the total of all land pollen including trees, shrubs and herbs. Aquatic plant pollen and moss spores are excluded from the main pollen sum. For all other groups (spores, aquatic pollen and indeterminable grains) the relative frequencies were calculated as percentages of their own sum plus the total pollen sum.

The concentration or absolute pollen frequencies of each pollen and spore type on a unit mass basis (grains/g of dry sediment) was determined based on the exotic spore counts and the known numbers of exotic spores added to each sample:

$$\text{concentration of fossil pollen} = \frac{F/E \times \text{exotic pollen added}}{\text{mass of sample}} \quad (1)$$

where F = fossil pollen counted

E = exotic pollen counted

Pollen concentration is more commonly expressed as grains per unit volume, but since many of the cores collected were ice-rich at some levels, it would be difficult to obtain accurate measurements of sediment depth and volume. Using sample dry weights avoids the problem of differential ice content throughout the core.

Numerical manipulation of the pollen count data to calculate percentages and concentrations was performed using TILIA (versions 1.13 and 2.0; Grimm, 1984). TILIAGRAPH (version 1.25; Grimm, 1984) was used to plot all pollen and macrofossil diagrams. Pollen diagrams were zoned based on the results of stratigraphically constrained cluster analyses using TILIA's CONNIS data analysis programme.

4.2.3 Plant macrofossil analysis

For plant macrofossil analysis, 20 cm^3 subsamples (measured by displacement in a graduated cylinder) were taken from 2-4 cm thick contiguous core segments near the base, and at intervals of 5-10 cm higher in the core. Subsamples were soaked overnight in water to break up material, then washed with a moderate stream of water on a nest of sieves (mesh diameter 1 mm, 250 µm, and 125 µm). All identifiable plant remains were picked from the residue under a binocular microscope at 10-40x magnification. Identifications were made with the aid of modern reference specimens and keys (Artjuschenko, 1990; Berggren, 1969; Crum and Anderson, 1981; Grout, 1965; Jessen, 1955; Kelley, 1953; Lévesque *et al.*, 1988; Martin and Barkley, 1973; Montgomery, 1977; Schoch *et al.*, 1988; Tryon, 1949). The prefix 'cf.' and suffix 'type' are used in the same way described for pollen results. Identifiable animal remains, which may provide important clues about the local environment, are also included in the macrofossil diagrams.

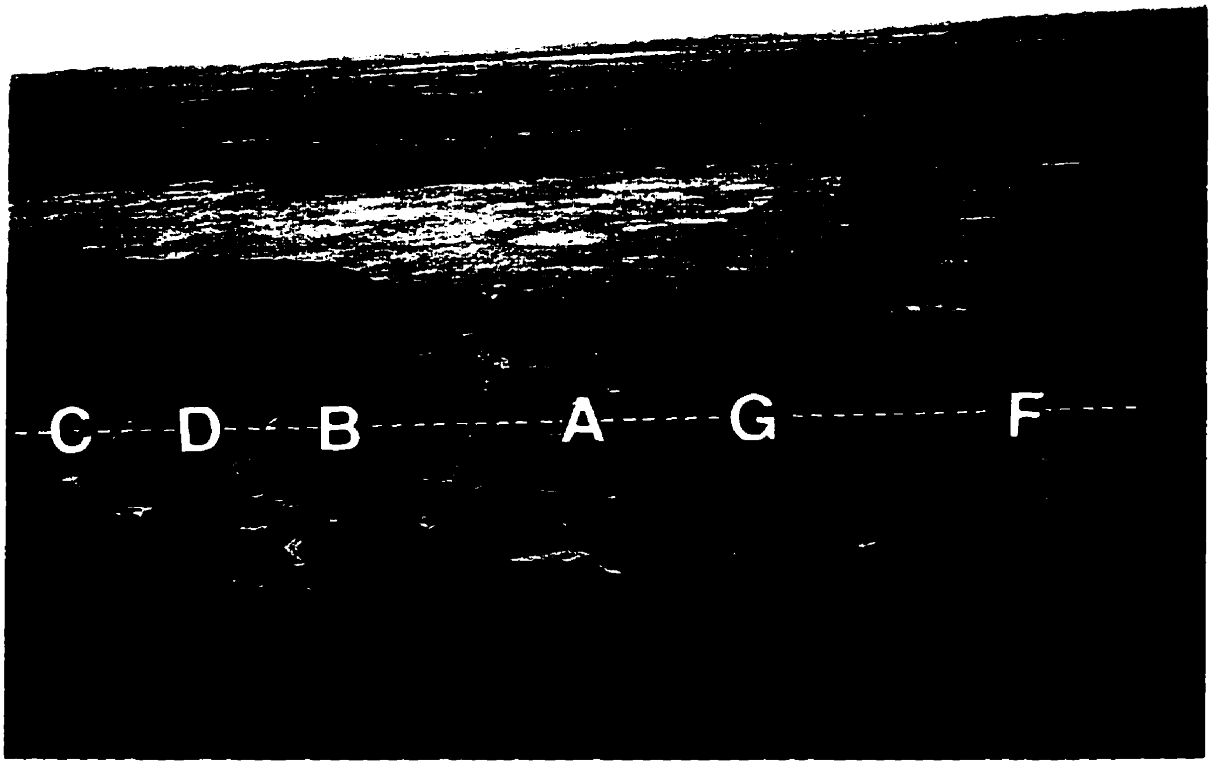
The relative proportions of different peat constituents (moss, wood, herbaceous remains) were estimated, and a sample of the moss fragments in each sample identified at least to genus or *Sphagnum* section (Janssen, 1988).

5. KUKJUK PEATLAND

5.1 SITE DESCRIPTION

Kukjuk peatland (informal name; 69° 29.6'N, 132°40.3'W) lies 17 km northeast of the community of Tuktoyaktuk, approximately 75 km north of the present treeline (Figure 1.1). Situated in the Low Arctic wetland region (National Wetlands Working Group, 1988), it occupies a small (150 m diameter) circular basin (Figure 5.1). approximately 1 km from the northwestern shore of Kukjukturijak Lake. The southwestern perimeter of the peatland is characterized by irregular high-centre polygons (10-30 m in diameter) surrounded by trenches 1-2 m wide. The high-centers of the polygons have a lichen ground cover with low shrubs such as *Rubus chamaemorus, Betula glandulosa, Ledum decumbens, Empetrum nigrum, Arctostaphylos uva-ursi* and *Vaccinium vitis-idaea* and small patches of *Sphagnum* spp. and other mosses. The trenches between the polygons are filled with slightly wetter *Sphagnum* lawns with *Eriophorum angustifolium* and *Rubus chamaemorus*. A broad, flat tussock meadow with *Calamagrostis* and *Carex* as well as *Sphagnum, Rubus chamaemorus* and various dwarf shrubs makes up the central area and north-eastern perimeter. The surface of the entire peatland was

Figure 5.1. Aerial view of Kukjuk peatland, with transect and coring sites indicated. Distance from C to F is 150 m.


C
D
B
A
G
F

dry at the time of sampling in August 1993. The peatland is bordered to the east, south and west by low (1-2 m) peat ridges with shrub *Salix* spp., *Betula glandulosa*, and a variety of heath; and to the north by an open sedge-marsh, the water table of which was approximately 1.5 m below the surface of Kukjuk peatland.

5.2 RESULTS

5.2.1 Basin Stratigraphy and Chronology

The peatland occupies a shallow depression underlain by inorganic diamicton (Figure 5.2). The permafrost table in mid-August 1993 was approximately 30 cm below the peat surface at all sampling sites. In cores KJ-A, KJ-B, and KJ-G, a fine-grained mixture of mineral and organic sediment at the base was overlain by a layer of ice 90-100 cm thick, mostly clear with vertically elongated air bubbles and discrete layers of peat 1-4 cm thick. The peat overlying this ice mass was relatively dry, but interspersed below the permafrost table with 1-4 cm thick horizontal lenses of ice which were either clear or contained 20-50% peat and silt by volume. A similar layer of dry peat with 1-4 cm thick ice lenses was found just below the permafrost table at core KJ-C. No sediments were found in the ice layer of core KJ-D, which was collected in an ice-wedge trench between high-centred polygons. In core KJ-F, from the eastern margin of the basin, the permafrost table at 25 cm depth corresponded to the boundary between a

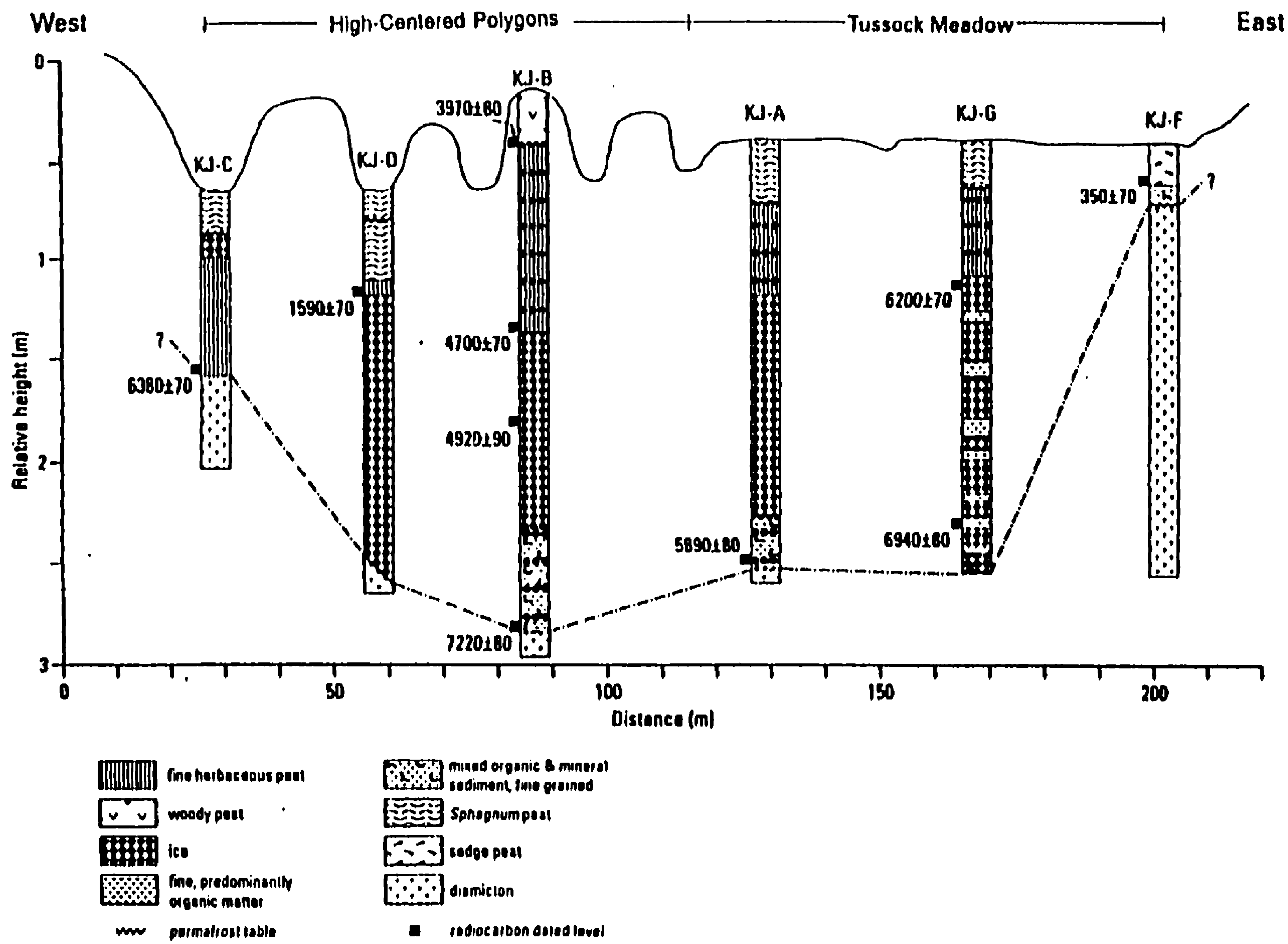


Figure 5.2. East-west stratigraphic profile of the Kukjuk basin.

diamicton (at least 2 m thick) with low ice content, and the overlying sedge peat.

The composition of the surface sediments varied depending on the habitat type in which the core was collected. The wettest habitats, in the moat surrounding the peatland and in ice-wedge cracks, contained a fine, black amorphous peat that was overlain by *Sphagnum* peat. The drier habitats in the polygons and unpatterned area contained poorly decomposed *Sphagnum* peat or highly decomposed woody or sedge peat immediately overlying the ice layers.

The radiocarbon ages (Table 5.1; Figure 5.2) suggest that the basin began to receive organic material by 7200 BP near the middle (core KJ-B), and across the rest of the basin by 5900 BP. A basal age of 350±70 BP (WAT-2768) at core KJ-F indicates much younger peat initiation than elsewhere in the basin, while a date of 3970 BP (WAT-2785) near the top of core KJ-B suggests that peat accumulation has been extremely slow since that time in the dry, lichen-covered high-centred polygons.

5.2.2 Pollen, spores and plant macrofossils

Core KJ-B (Figures 5.4 and 5.5): *Betula* is the dominant taxon throughout the entire pollen profile. Most of the *Betula* grains were 17-19 µm in diameter and, therefore, are believed to represent shrub species, although some may be from tree species. *Pinus, Abies*, and perhaps some tree *Betula*, are attributed to long-distance sources. *Picea* pollen values are consistently lower than 7%, which might also be attributable to long distance sources, and *Larix* pollen was found at

Table 5.1. Radiocarbon ages from the Kukjuk peatland.

Core	Depth (cm)	Radiocarbon age[a] (yr B.P.)	Laboratory number[b]	Dated material
KJ-A	243-250	5890±80	WAT-2788	Organic/mineral sediment
KJ-B	24-26	3970±80	WAT-2785	Peat
	116-119	4700±70	WAT-2766	Peat
	161-165	4920±90	WAT-2786	Peat
	276-280	7220±80	WAT-2765	Organic/mineral sediment
KJ-C	104-110	6380±70	WAT-2787	Peat
KJ-D	58-61	1590±70	WAT-2767	Peat
KJ-F	22-24	350±70	WAT-2768	Peat
KJ-G	75-80	6200±70	WAT-2789	Peat
	228-234	6940±80	WAT-2813	Organic/mineral sediment

[a]Ages are corrected for isotopic fractionation and are reported with 1σ.

[b]Laboratory designation: University of Waterloo Radiocarbon Laboratory.

only two levels. However, *Picea* and *Larix* needle fragments near the base (Figure 5.5) indicate that these trees were present near the coring site around 6500-7000 BP.

The pollen diagram (Figure 5.4) from this core can be divided into three zones based on differences in local pollen taxa and plant macrofossils. Zone KJB-1 is characterised by peaks in the pollen percentages for *Salix* and Cyperaceae. Macrofossils of *Empetrum, Arctostaphylos*, and *Dryas integrifolia* confirm their presence locally (Figure 5.5). These sediments also yielded seeds of submerged and emergent aquatic plants such as *Potentilla palustris, Hippuris, Myriophyllum, Ranunculus aquatilis*-type, *Carex, Typha* and oospores of *Chara* and *Nitella*. The remains of several types of aquatic invertebrates indicate shallow open water habitats.

Zone KJB-2 starts with a marked increase in *Alnus* pollen at approximately 6000 BP, a slight decrease in *Salix* and Cyperaceae pollen and a noticeable decrease in macrofossils of aquatic macrophytes which were dominant in Zone KJB-1. There is a general increase in total organic matter and a change to core material consisting largely of ice (Figure 5.3). Pollen of *Empetrum* and *Arctostaphylos* show slight increases and macrofossils confirm their presence near the coring site. Spores and macrofossils of *Sphagnum* increase noticeably in this zone.

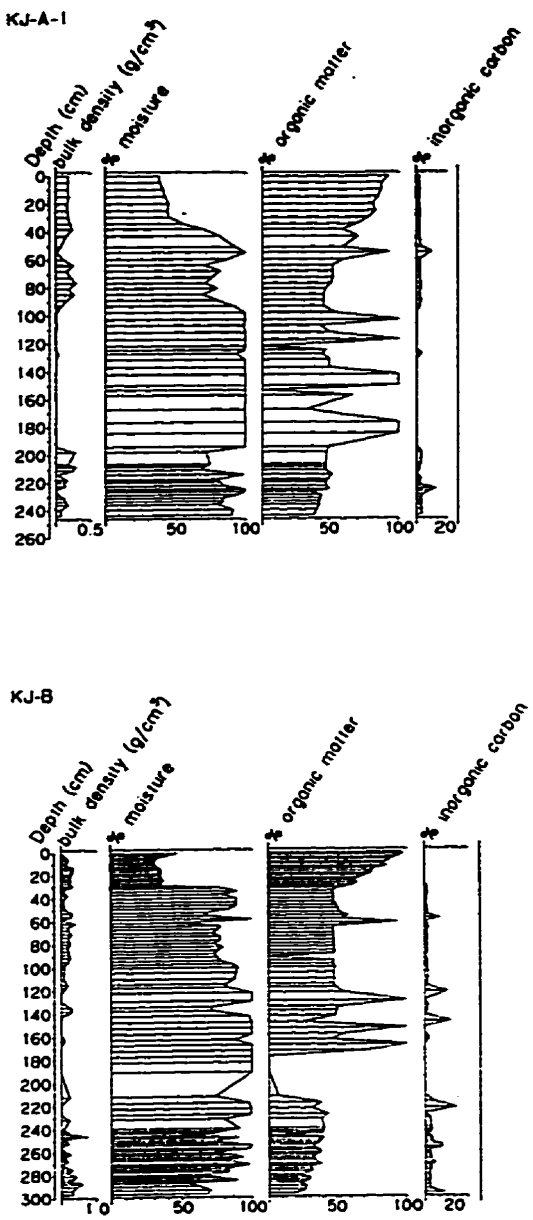


Figure 5.3. Profiles of the six cores from Kukjuk peatland showing bulk density, moisture (% of fresh weight), organic matter (% of dry weight), and inorganic carbon (% of dry weight). Zero point on each diagram represents the peat surface at the coring surface. Note that for core KJ-F, only the material overlying the diamicton was analysed. (Continued on next page)

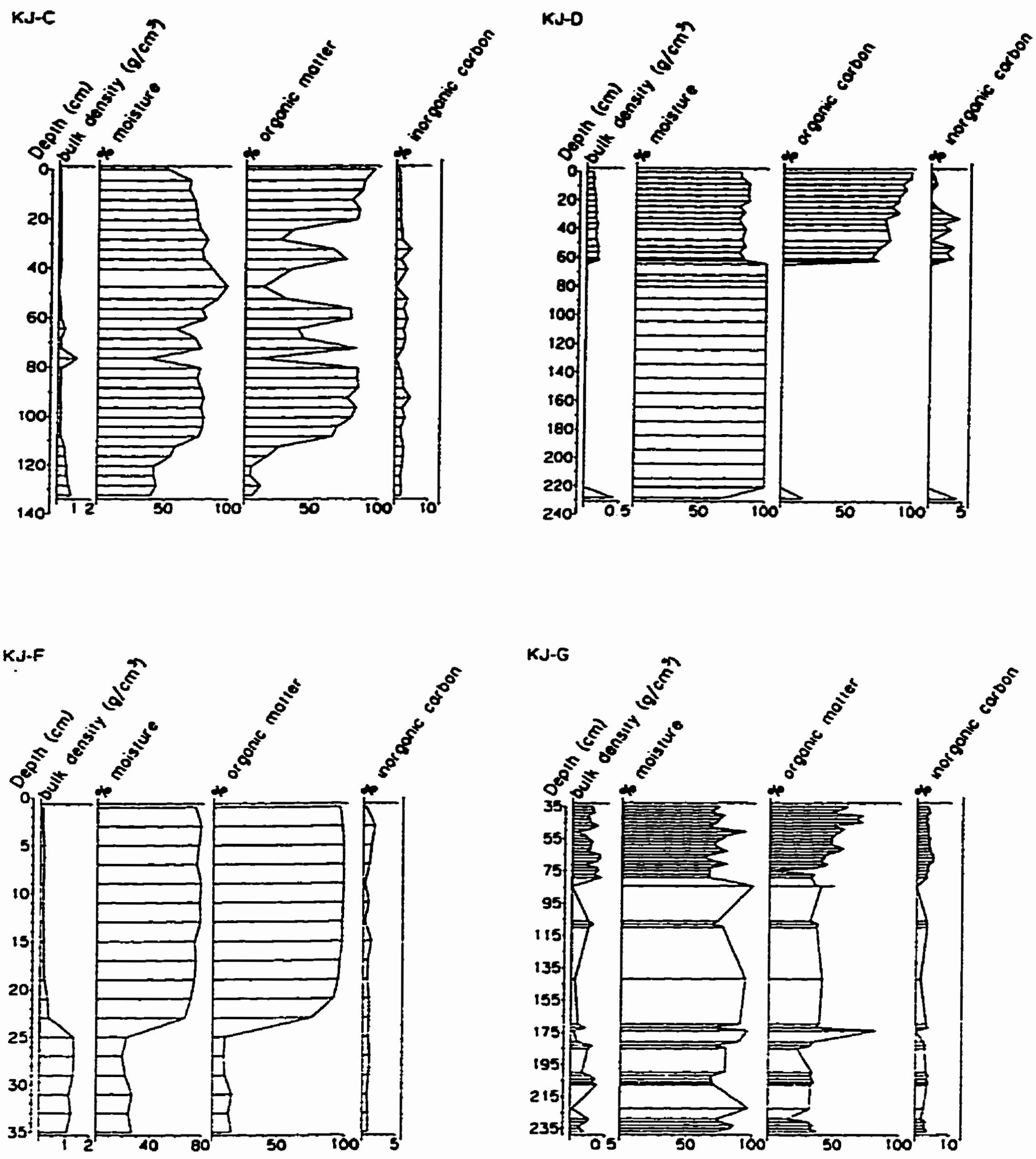


Figure 5.3. (continued from previous page).

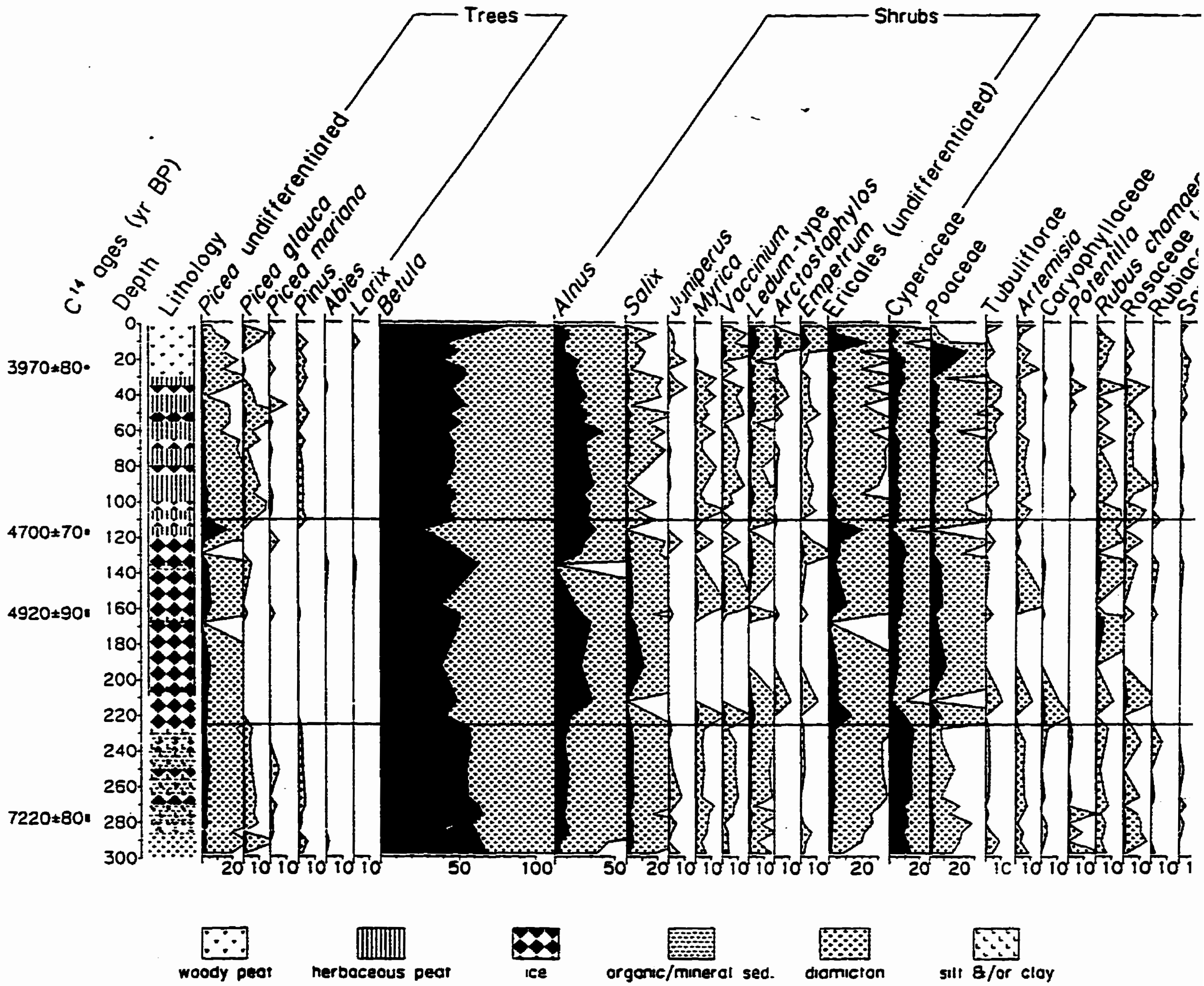
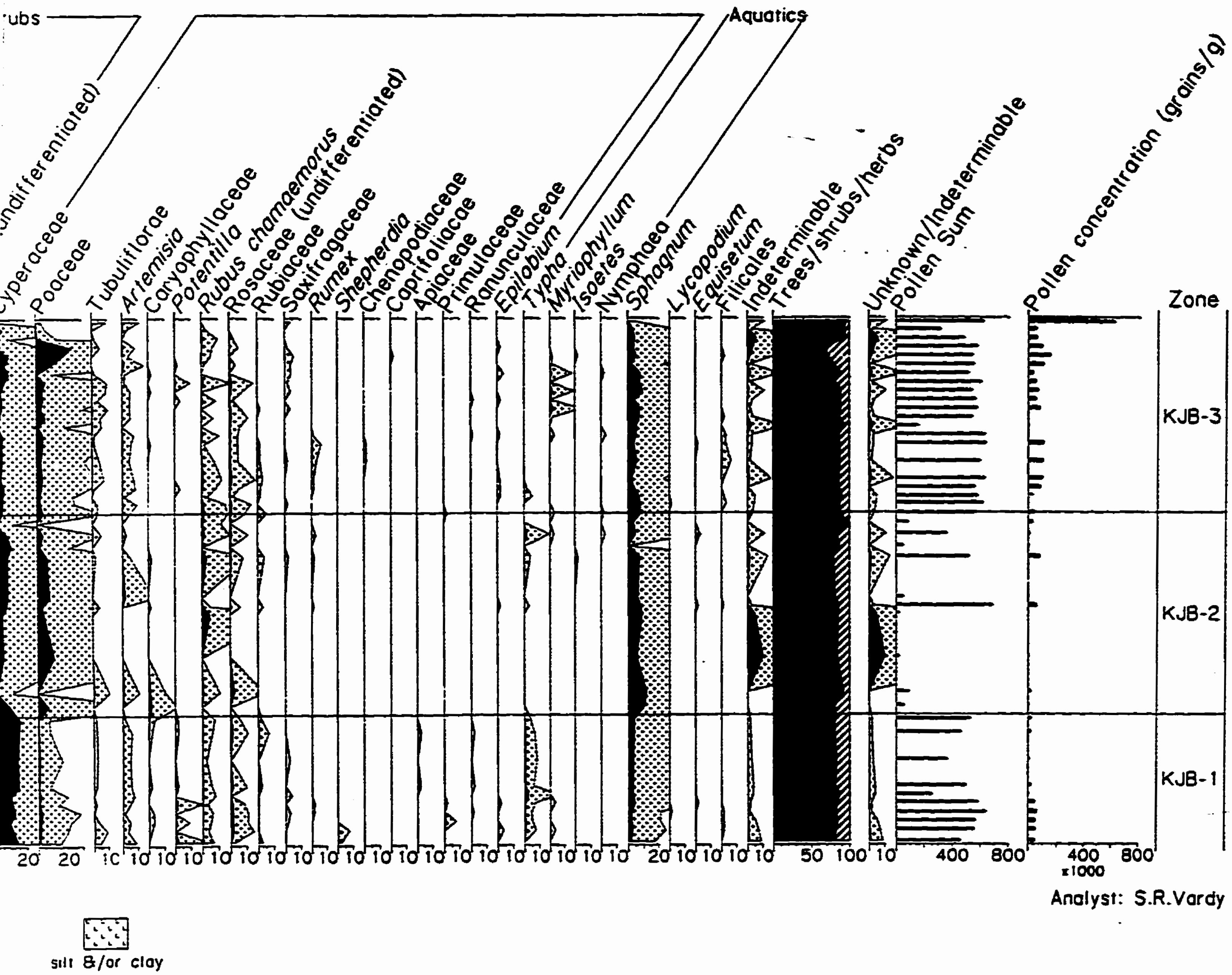


Figure 5.4. Pollen diagram for core KJ-B. Zero po
figures represents the peat surface at the coring si



ram for core KJ-B. Zero point on this and subsequent
peat surface at the coring site.

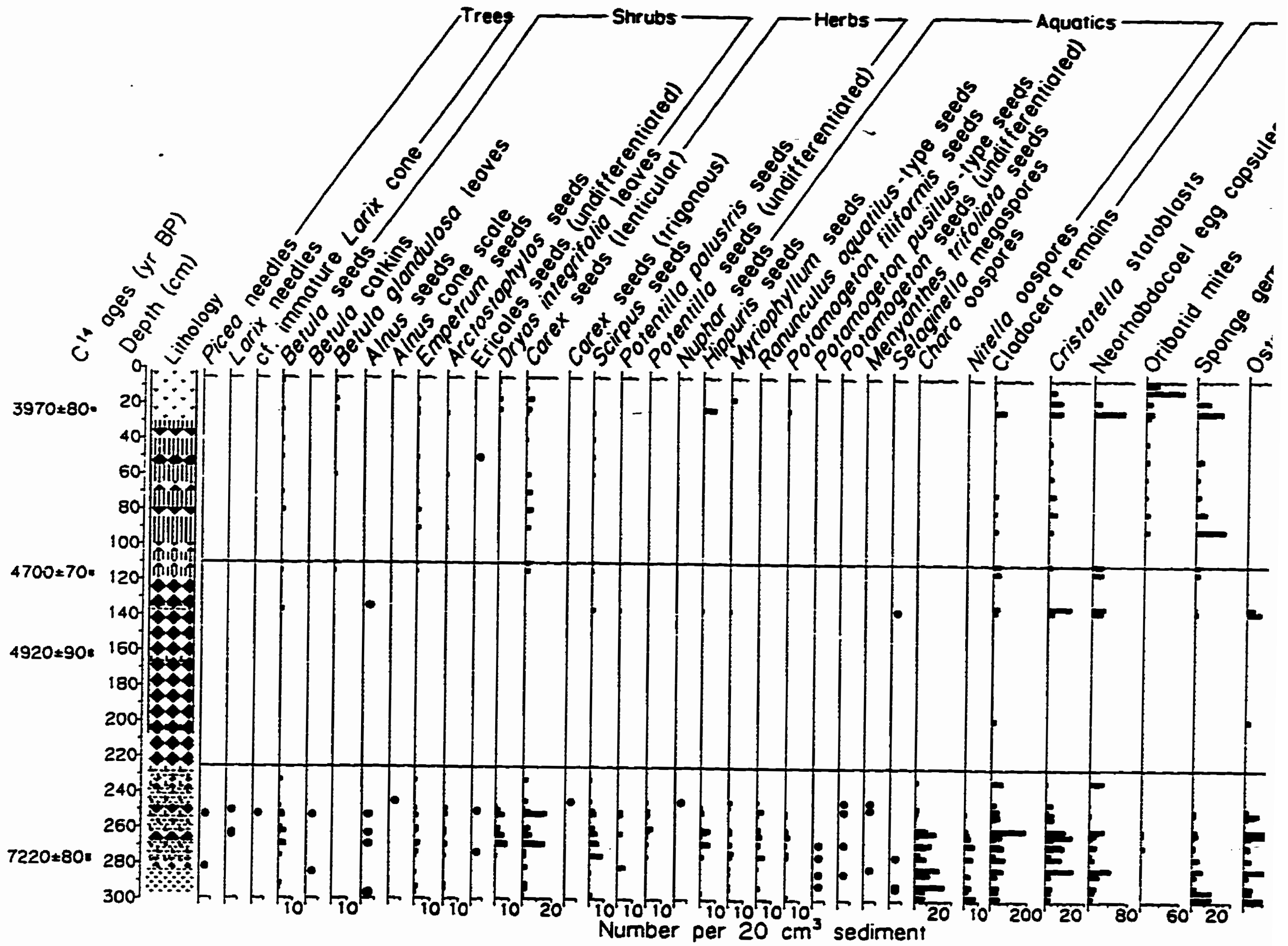
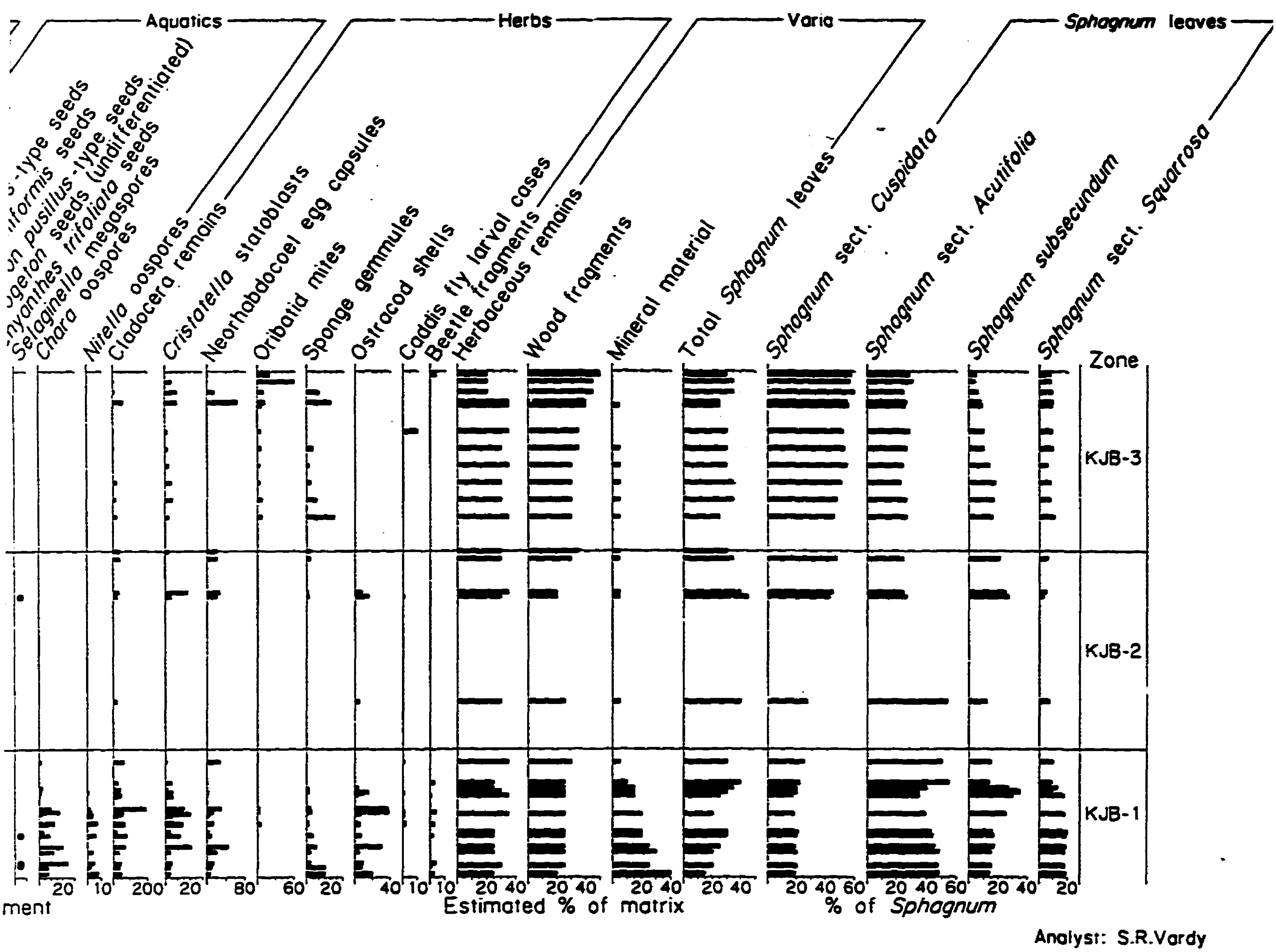


Figure 5.5. Macrofossil diagram for core KJ-B. "U which could not be identified to species; "•" indic sediment. See Figure 5.4 for lithology key.



'ofossil diagram for core KJ-B. "Undifferentiated" refers to seeds be identified to species; "•" indicates less than 3 per 20 cm³ of ;ure 5.4 for lithology key.

The peat in zone KJB-3 contains pollen of many of the shrub, dwarf shrub and herb taxa which occur on the surface of the peatland today. Pollen, plant macrofossils and remains of invertebrates indicate the presence of shallow open-water habitats, but probably more restricted than in zone KJB-1.

Core KJ-G (Figures 5.6 and 5.7): The same general trends recorded in core KJ-B are apparent in this core, which can be divided into two pollen zones. Zone KJG-1 contains high *Betula* and Cyperaceae pollen values and zone KJG-2 shows an increase in *Alnus*. The age of the transition between them is about 6300 BP, approximately the same age as the boundary between zones KJB-1 and KJB-2. In general, the rest of the pollen record is more or less uniform throughout the core, not unlike the pollen record in core KJ-B.

5.2.3 Stable Isotope Analysis

Hydrological changes that occurred during peatland development have been inferred from analysis of the stable isotopes oxygen-18 (^{18}O) and deuterium (^{2}H) in ice and water samples from core KJ-B (Vardy *et al.*, 1997). The isotopic composition of ground ice is a function of the sources of water for ice growth, which include snow melt, summer precipitation and thawed ice in the active layer. Isotopic fractionation during freezing could also have an effect, but this is not usually more than 2‰ for ^{18}O, and the similarity between the isotopic composition of the active layer and the mean annual precipitation suggests that

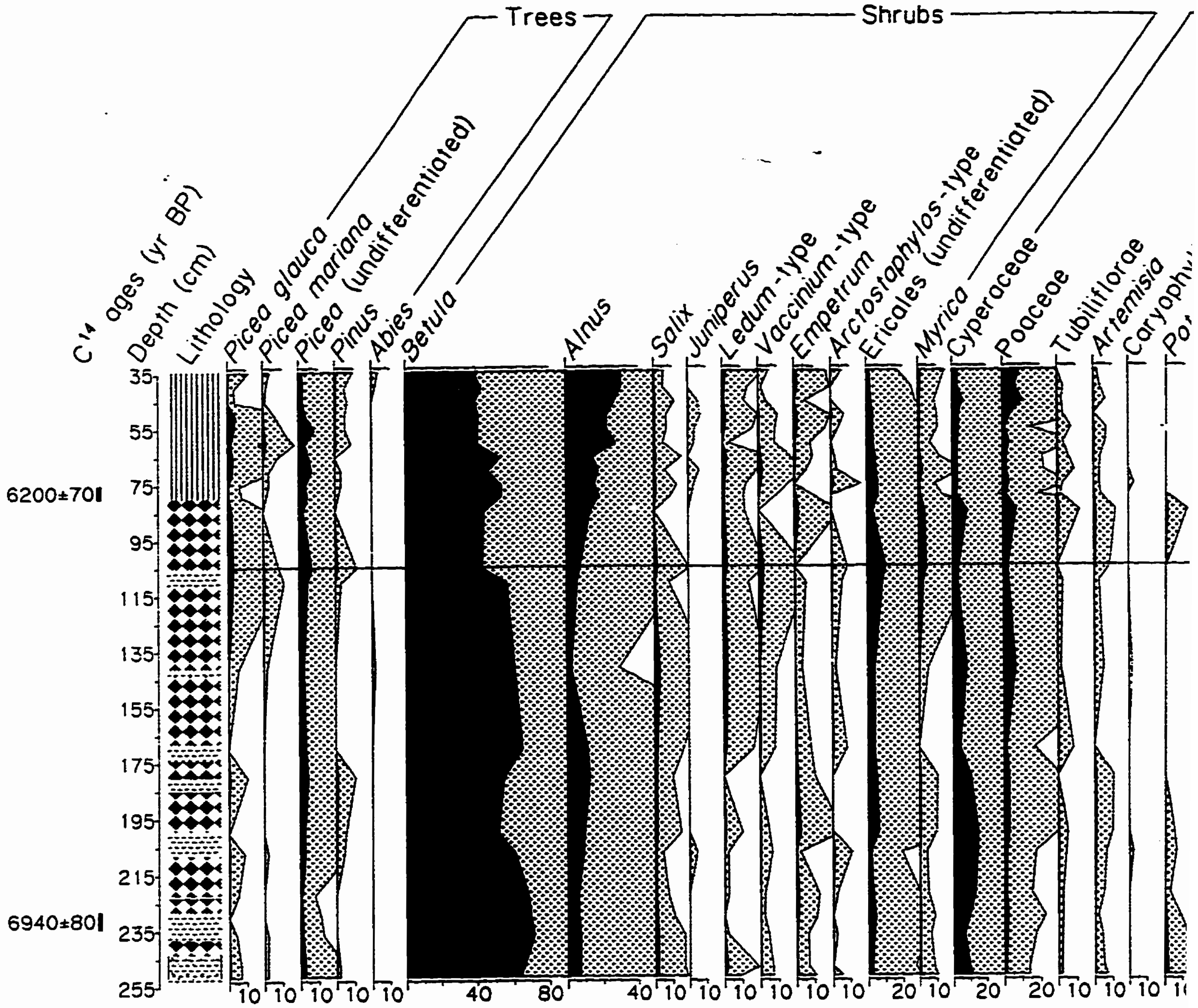
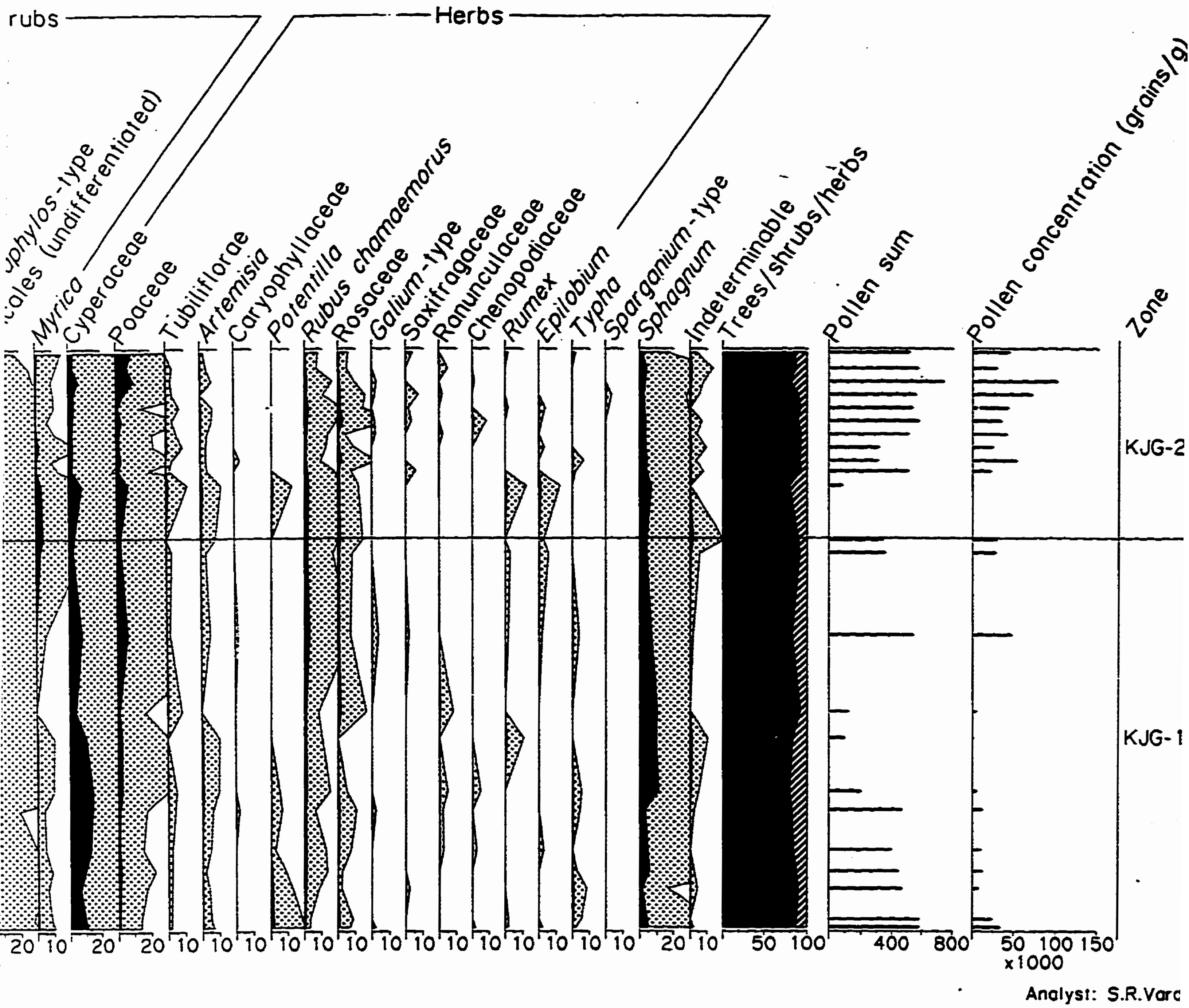


Figure 5.6. Pollen diagram for core KJ-G. See Figu



ram for core KJ-G. **See Figure 5.4 for lithology key.**

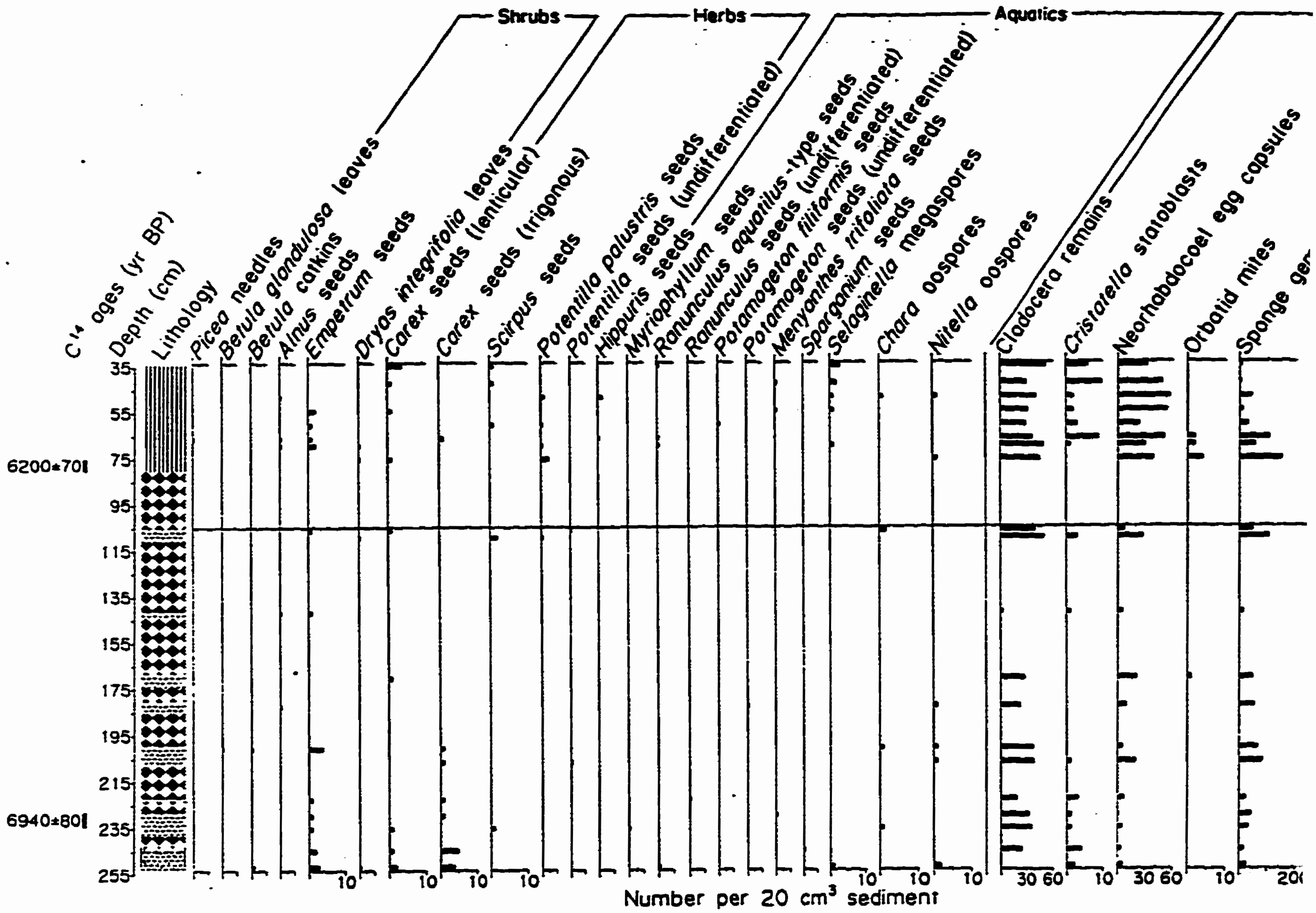
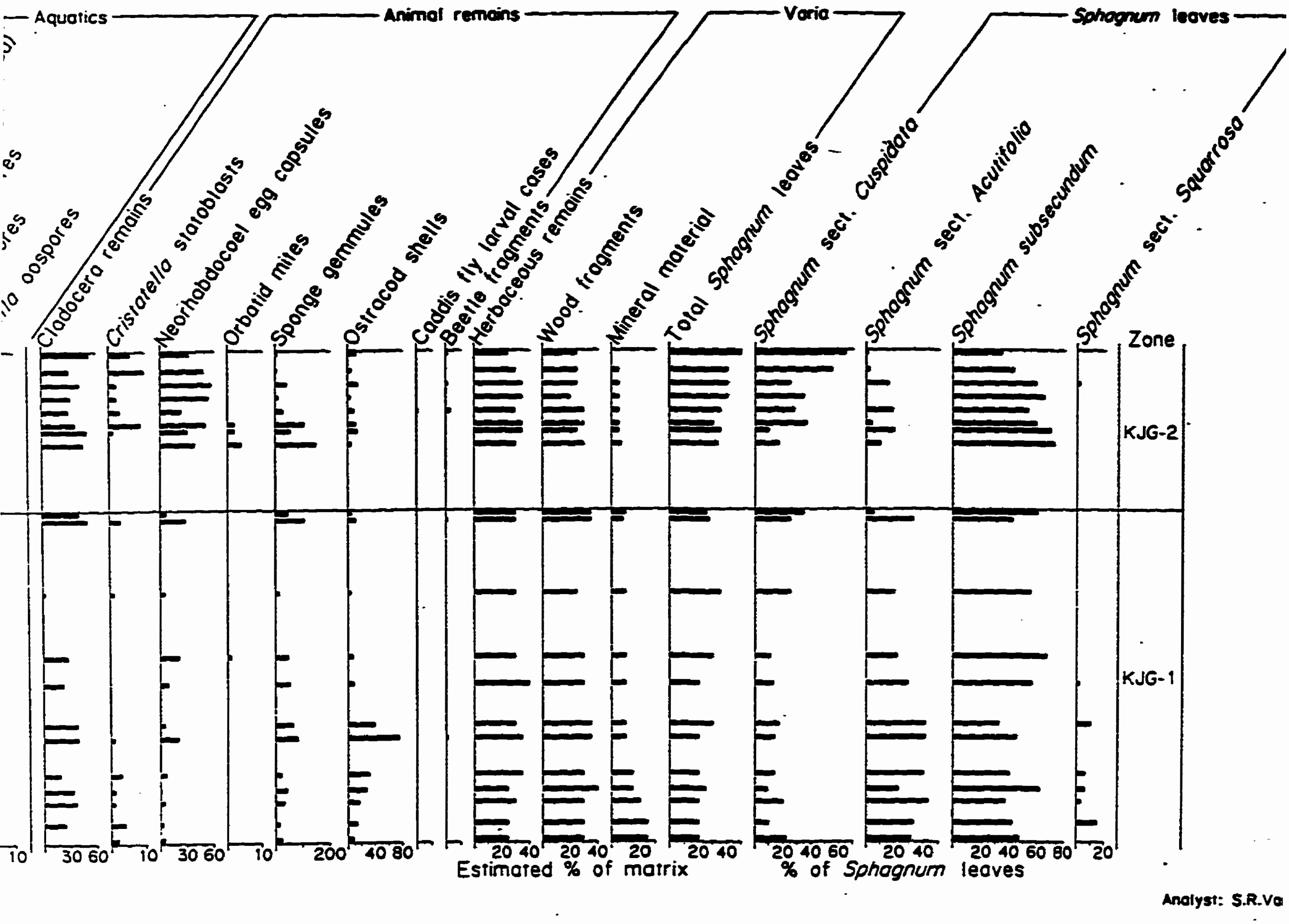


Figure 5.7. Macrofossil diagram for core KJ-G. Se



sil diagram for core KJ-G. See Figure 5.4 for lithology key.

this effect is not significant in the Tuktoyaktuk area (Michel *et al.*, 1989). Due to relatively dry climate in the study area, evaporation should play a role in the water balance of open water bodies by enriching the remaining water with the heavy isotopes, ^{18}O and ^{2}H, following a well-defined trend that is a function of relative humidity and temperature (Gibson *et al.*, 1994).

A significant isotopic range was observed in water from ice samples and peat pore water, varying between -18.5‰ and -9.8‰ for $\delta^{18}O$ and between -148‰ and-113‰ for $\delta^{2}H$ (Figure 5.8; Vardy *et al.*, 1997). $\delta^{18}O$ values were around -13‰ in the deepest part of the core, corresponding with the pollen zone KJB-1 (Figure 5.9). A trend to more enriched values continues up core, reaching a peak value of -9.8‰ at 215-209 cm depth, at the base of zone KJB-2, followed by a sharp $\delta^{18}O$ reversal. Isotopic values then become more progressively more depleted toward the top of the core, reaching a value of -18.6‰ at the surface of the peatland.

5.3 *DISCUSSION*

Paleoecological records covering the last 12,500 years from the Tuktoyaktuk Peninsula and adjacent areas show high levels of *Picea* pollen in the early Holocene, indicating that the treeline was 75-100 km north of its present limit,

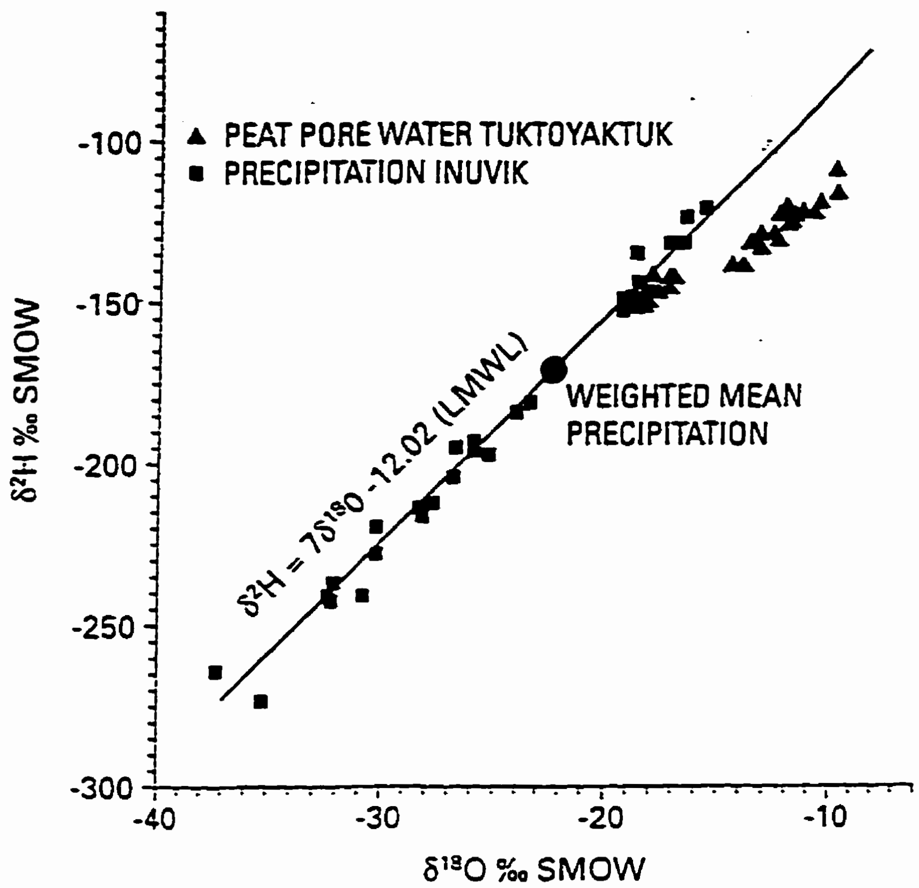


Figure 5.8 Isotopic composition of water from ice samples and peat for core KJ-B. Also shown is modern precipitation for Inuvik. (LMWL - Local meteoric water line). After Vardy *et al.* (1997).

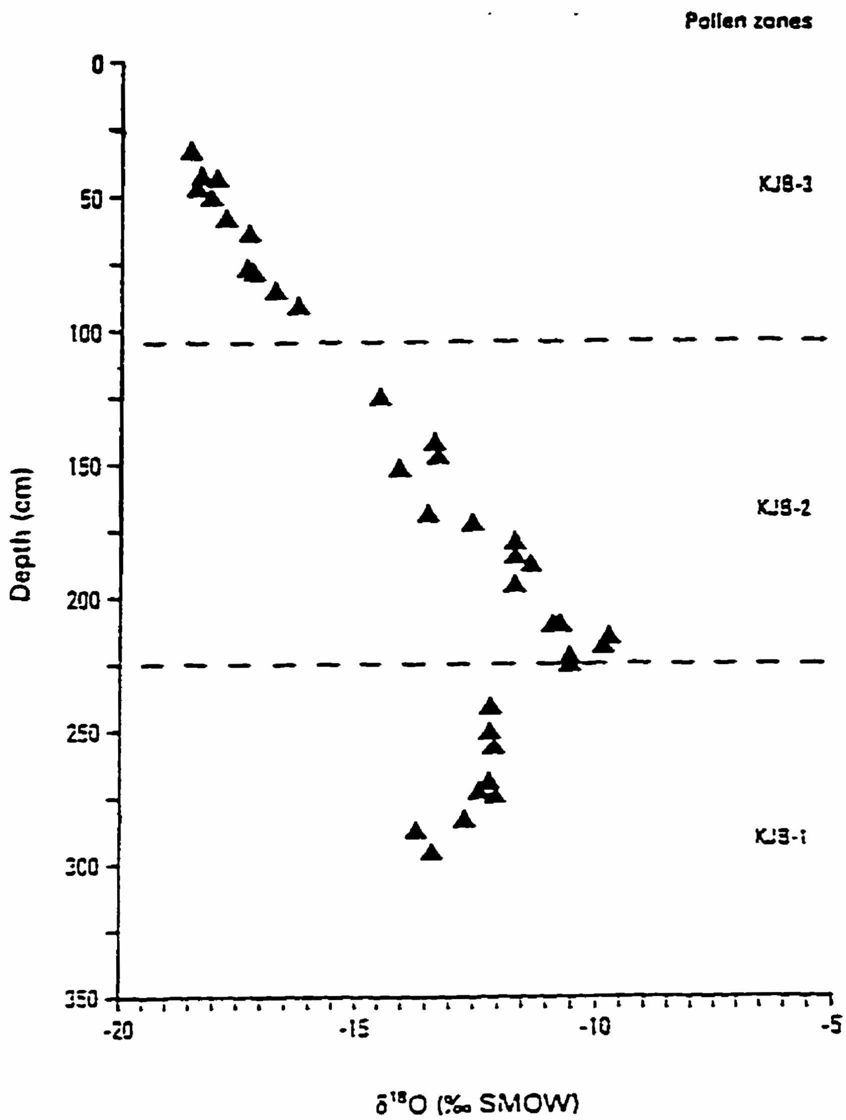


Figure 5.9. $\delta^{18}O$ depth profile for water from ice samples and peat for core KJ-B (After Vardy *et al.*, 1997). Pollen zones are shown for comparison with Figures 5.4 and 5.5.

with trees reaching the Tuktoyaktuk Peninsula between 10,000-5000 BP (Ritchie and Hare, 1971; Ritchie, 1972, 1984; Hyvärinen and Ritchie, 1975; Spear, 1983). The presence of spruce on the peninsula is confirmed by macrofossils found at a number of localities (Ritchie and Hare, 1971; Delorme *et al.* 1977; Spear, 1983), in addition to the needles found in the Kukjuk peatland. A *Larix laricina* tree stump from a peat deposit near the East Channel of the Mackenzie Delta, dated at 7510 ± 140 BP (BGS-472), provides evidence that the range of this species also extended 75-80 km beyond its present limit (Zoltai and Zalasky, 1979). As explained in Chapter 3, climate is believed to have been significantly warmer than present, possibly as a result of an early Holocene Milankovitch insolation maximum (Ritchie *et al.*, 1983; Ritchie, 1984), although greater distance from the Beaufort Sea coast as a result of lower sea level may also have been a factor (Hill *et al.*, 1985; Pelletier, 1987; Mackay, 1992). Mean July temperature is estimated to have been at least 3°C warmer than present (Ritchie, 1984). The paleoecological record from lake sites indicates a gradual climatic degradation began around 8000 BP, with a more rapid cooling around 4500-5000 BP (Ritchie, 1984).

The influence of geocryological processes complicates reconstruction of the developmental history of peatlands affected by permafrost. For example, the growth of ice wedges results in the deformation of sediments in the adjacent polygons (Mackay, 1980, 1992; French, 1976). Mackay (1980, 1993, 1995) has also shown that materials in the active layer and top of the permafrost in low centred polygons are prone to horizontal movement as a result of summer expansion following winter contraction, with a net movement of up to several millimetres a year, towards the ice-wedge troughs (Mackay, 1995). Adjacent to the ice wedges, the upper part of the permafrost may therefore move inward in the opposite direction to that of the summer active layer, so that some shearing may take place. Such deformation potentially poses problems for stratigraphic interpretation of samples from cores, which may not represent *in situ* vertical sequences. Core KJ-B was collected from close to the centre of an ice-wedge polygon, and core KJ-G from the unpatterned area of the peatland, so deformation of sediments by ice-wedge growth should be relatively minimal at both coring sites. While it is possible that thermal expansion has resulted in significant horizontal movement of sediments, the radiocarbon profiles do not indicate any vertical displacement of material. It may therefore be assumed that these two cores provide a long-term chronological record of changing conditions in the peatland, even if the sediments might have been susceptible to short-term geocryological disturbances.

High inorganic content in peat deposits is sometimes related to cryoturbation, but the presence of regular horizontal banding of ice and sediment throughout the cores suggest that this is not a significant factor in this peatland. The inorganic component is composed of fine-grained silt and clay, and may be the result of sediment inflow or deposition of windblown particles.

The Kukjuk basin began to receive organic sediments at around 7200 BP. There is no record of what happened at the site prior to this date, but it is possible that the basin originated by thermokarst collapse of ice-rich sediments during the early Holocene warm period. Thermokarst processes were particularly active in the region during this time, with the period of maximum thaw lake formation around 9000-8000 BP (Mackay, 1992).

A shallow open-water mineral wetland occupied the basin initially. *Carex, Scirpus, Typha, Menyanthes trifoliata* and *Sphagnum* probably grew around the edges of a shallow pond inhabited by submerged aquatic plants such as *Myriophyllum, Potamogeton filiformis, Isoetes, Hippuris,* and *Nuphar,* as well as populations of various aquatic invertebrates. The permafrost table was probably below the bottom of the basin, which allowed enough water to collect in the basin to support a shallow mineral wetland.

The average dispersal distances for *Typha* pollen have been found to be only 2-10 m, with maximum distances of up to 1 km in strong winds (Krattinger, 1975), so even isolated pollen grains indicate that the plant grew in or near the Kukjuk peatland, more than 500 kilometres north of its present limit (Porsild and Cody, 1980; Ritchie 1984). *Typha* pollen has previously been found at several other sites in the area from 11,500-5000 BP (Ritchie, 1972, 1984; Hyvärinen and Ritchie, 1975; Spear, 1983). *Typha* pollen in early Holocene sediments from this region therefore provides further evidence of a climate warmer than today, with mean temperatures up to 6°C warmer for the months of May through August (Ritchie, 1984).

Isotopic analysis on the ice from this time interval show $\delta^{18}O$ values between -13‰ and -9.8‰, most likely representing an open water body that was affected by evaporation (Figures 5.8 and 5.9; Vardy *et al.*, 1997). The effects of evaporation have been well established by $\delta^{18}O$ data from modern water bodies on the Tuktoyaktuk Peninsula which show a range of -19.1‰ to -8.1‰ (Mackay, 1983), values 3.6 to 14.6‰ more enriched than the mean weighted isotopic composition of the precipitation in the study area today (i.e. -22.7‰, Figure 5.8). Insufficient data precludes speculation on past changes in the mean weighted isotopic composition of the precipitation that may have accompanied changes in mean annual temperature. The permafrost table was probably below the bottom of the basin, which contained enough water to support a shallow mineral wetland.

Submerged and emergent aquatic macrophytes continued to invade the wetland. Beginning as early as 6300 BP and certainly by 5000 BP, transformation from open-water mineral wetland to a graminoid fen peatland was underway. *Ranunculus aquatilis*-type may have grown in wet mud along the edges of the basin or in shallow open water. Wet meadows with *Carex, Scirpus, Menyanthes trifoliata,* and *Sphagnum* and possibly shrubs such as *Myrica* and *Ledum* occupied parts of the wetland where more peat had accumulated.

Around 6300 BP, permafrost began to affect the wetland. $\delta^{18}O$ values show a marked reversal towards more depleted values (Figures 5.8 and 5.9), that is interpreted as a function of changing hydrological conditions in the wetland, probably due to the aggradation of permafrost (Vardy *et al.*, 1997). Material at the bottom of the wetland may not have totally thawed during the summer, resulting in progressively less atmospheric exposure to evaporation. The net result was a build up of ice and organic-rich ice, probably formed syngenetically by segregation during freezing. The vertically elongated air bubbles and horizontal layering of the sediments and ice lenses are characteristic of segregation ice (Washburn, 1979, p.46).

These changes are roughly coincident with a decline in *Picea* pollen in nearby lake sediments beginning by 5000-6000 BP, which signals the southward retreat of treeline from the peninsula and the end of the early Holocene Milankovitch insolation maximum (Ritchie, 1984).

A peatland largely influenced by permafrost was well established by 4700 BP. Studies at numerous sites in the region indicate that ice-wedge growth, which was uncommon during the early Holocene warm period, was underway around 4500 BP (Mackay, 1992). Though pollen and plant macrofossils show no major changes during this time, it is possible that low-centred polygons were beginning to develop in the wet sedge-meadow, later evolving into the high-centred polygons found on the south-western edge of the basin today. The formation of ice-wedges, such as those separating the high-centred polygons in the peatland, begins with thermal contraction in frozen ground, forming cracks which expand over a number of years through infilling by spring meltwater and downward percolation (Mackay, 1974). In an area of peat accumulation such as this, ice wedges are typically syngenetic, growing more or less simultaneously with the deposition of material (Mackay, 1990). Core KJ-D was collected in one of the ice-wedge trenches, and reveals characteristic *Sphagnum* peat overlying the ice wedge. An age of 1590 ± 80 BP (WAT-2767) shows that peat began to accumulate in the trough by this time.

By 4000 BP, the peatland had differentiated into its present-day habitats. Peat accumulation coupled with continued ice growth probably raised the peatland so that earlier wet meadows and possibly low-centred polygons were converted to the high-centred polygons which are characteristic of the south-western area of the peatland today. Peat accumulation rates declined as a result, with only a 25 cm increase in peat thickness in the last 4000 years in the centre of one of the high-centred polygons (core KJ-B).

The isotope data from core KJ-B show a trend towards more depleted $\delta^{18}O$ values, changing from -17‰ to -18.8‰ around this period (Figs. 5.8 and 5.9; Vardy *et al.*, 1997). A comparison of the isotopic composition of peat pore water and weighted mean isotopic composition of the precipitation indicates that summer precipitation is the main source of water within the active layer of the permafrost. The $\delta^{18}O$ peat pore water values are within the -18‰ to -20‰ range reported for waters in the modern active layer on the Tuktoyaktuk Peninsula (Fujino and Kato, 1978; Michel and Fritz, 1980, 1981). The peat pore water $\delta^{18}O$ data plot close to the local meteoric water line, indicating that these waters have not been affected by evaporation (Figure 5.8), similar to other active layer isotopic studies in the Yukon Territory (Harris *et al.* 1992).

The record from the northeastern area of the peatland indicates that the meadow and tussock communities have not changed markedly since inception at around 7000 BP, except for a possible decrease in extent or depth of open water. Given that the dry high-centred polygon development is on the south-western upslope side of the basin, water running off the surface would be directed around the high-centred polygons and downslope to the northeastern side. This supply of water would maintain the wetter sedge meadows and even allow for a slow expansion downslope which has been underway since at least 350 BP.

The results illustrate the significant effect permafrost has on peatlands, by elevating the surface and drastically altering the hydrology. A previously water-saturated surface becomes drier, allowing an entirely different vegetation to grow. Elevation of the peat surface by upward permafrost aggradation may eventually lead to conditions too dry for peat accumulation, as evidenced by reports of old radiocarbon ages (i.e. 1145 to 3000 BP) on near-surface peats in arctic Canada (Zoltai and Tarnocai, 1975; Ovenden, 1982; MacDonald, 1983;

Zoltai, 1995), and Russia (Smirnov, 1992; A. Andreev and B. G. Warner, personal communication). The results from this site also confirms that peatland growth in permafrost regions is not uniform within a single basin. Some parts may be actively accumulating peat while other parts may have negligible accumulation.

5.4 CONCLUSIONS

The results from the Kukjuk peatland suggest that the timing of detectable vegetational and hydrological changes in the peatland are coincident with the timing of movements in the position of treeline and regional climate changes. The basin with a mineral wetland in it formed during the early Holocene insolation maximum. The wetland responded to the subsequent deterioration of climate, with permafrost formation around 6000-6300 BP and gradual transformation to a peatland by about 4500-5000 BP.

Climate is often used to account for stratigraphic and paleoecological changes in peatlands. The climate signal can be difficult to recognise because changes in the peat record brought about by autogenic succession, or by any of a variety of local factors which might affect the hydrology, could cause comparable changes in the record. Generally, it is difficult to recognise internally driven successional changes from those driven by extrinsic factors, such as climate, without an independent climate record for comparison. While the transition from sedge dominated to *Sphagnum* peatland may be the result of autogenic hydroseral succession, a comparison of the results from this site with the independent regional record suggests that the change was associated with a change in climate.

6. CAMPBELL CREEK PEATLAND

6.1 SITE DESCRIPTION

Campbell Creek Peatland (68°17.3'N, 133°15' W; Figure 1.1) is located 21 km ESE of the community of Inuvik, immediately west of Campbell Creek and north of the Dempster Highway. It is a 6.5 ha peat plateau complex surrounding a small (70 m diameter) pond (Figure 6.1), situated in the High Continental Subarctic Wetland subregion (National Wetlands Working Group, 1988). The microtopography varies across the peatland and there are several distinct plant communities, including *Sphagnum fuscum* hummocks with abundant *Rubus chamaemorus* and ericaceous shrubs; wet *Sphagnum cuspidata* filled hollows, flat surfaces covered with lichens including *Cladina rangiferina*, *C. alpestris* and *Cetraria nivalis*; and a wet low-lying area near the southeastern end of the pond with *Carex* spp., *Alnus crispa*, *Myrica gale*, *Vaccinium vitis-idaea* and *V. uliginosum*. *Betula glandulosa* is abundant across the entire peatland (Figure 6.2).

Scattered *Picea mariana* and a few *Larix laricina* grow on the peatland. *Betula papyrifera*, mostly dead, are found scattered throughout the peatland and concentrated around the perimeter (Figure 6.2). The peatland is surrounded to the north, east and west by denser forest including *P. glauca* as well as *P. mariana*,

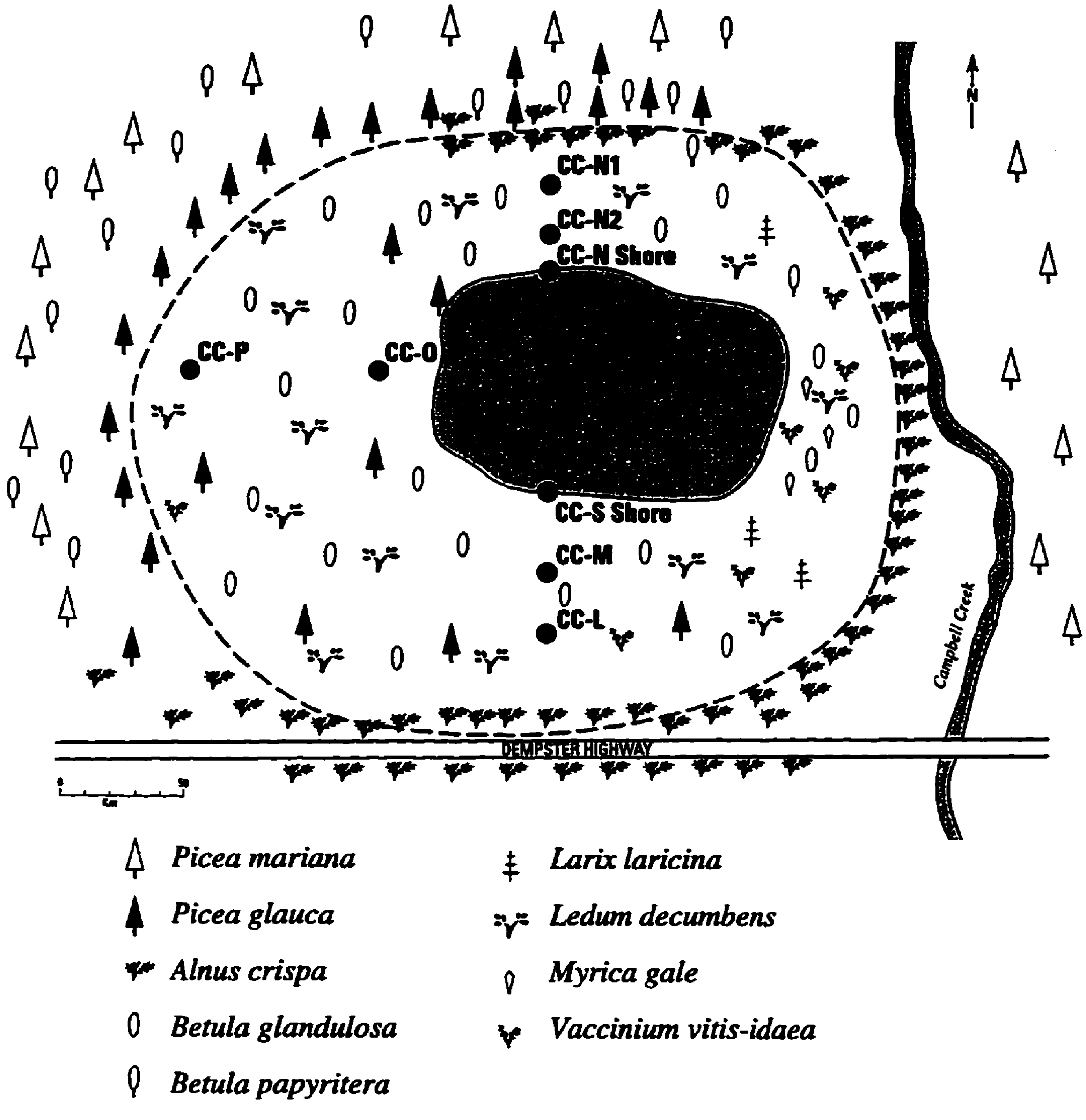


Figure 6.1. Sketch map of Campbell Creek peatland, illustrating dominant vegetation types and positions of coring sites. The dashed line delineates the peatland.

Figure 6.2. Campbell Creek peatland. a) *Larix laricina* on the right; stunted *Picea mariana* in the left foreground. b) The raised western margin of the pond. *Betula glandulosa* and *Ledum decumbens* are abundant across most of the peatland. Taller *P. mariana* like those near the pond are often tilted as a result of being rooted in the shallow active layer. The denser *Picea glauca* surrounding the peatland is visible in the distance.

a

b

Larix laricina, and *Betula papyrifera* (also mostly dead). Thickets of *Alnus crispa* line the south-southwestern edge of the peatland, along the Dempster Highway.

The peatland lies within Mackay's (1963) Campbell Lake Hills Physiographic Region, which is a mainly upland area where bedrock is close to the surface and outcrops in escarpments. Ritchie (1977, 1984) has published detailed pollen records from two lakes in these uplands, Maria Lake (M-Lake) and Twin Tamarack Lake (TT-Lake) (Figure 3. 3). The peatland is within the Campbell-Sitidgi Lake Lowland section of this region, a low, flat area which is part of the course of a former river which flowed through the Campbell-Sitidgi Lake depression. The surficial deposits under the peatland area are Late Wisconsin lacustrine deposits, including silt, sand and some gravel, deposited mainly through thermokarst during a high water phase of the Eskimo Lakes (Rampton, 1988).

6.2 RESULTS

6.2.1 Basin stratigraphy and chronology

A total of six cores were collected and two sections were sampled from this peatland. In August of 1992, two sections at the edge of the pond (CC-N Shore and CC-S Shore) were described and sampled, and six cores were collected in August of 1993, four of which formed a north-south transect including the two shoreline sections (Figure 6.1). The remaining two cores (CC-

O and CC-P) were collected 25 and 100 m, respectively, west of the pond. With the exception of the southeastern area and some wet hollows, the surface was dry at the time of sampling. Active layer depths varied from 30 to 60 cm, being deepest in wet areas of the peatland. Unlike at the Kukjuk peatland, a water table was present in the active layer at most coring sites, at depths of 10 to 45 cm (Figure 6.3). Ground ice is restricted mainly to thin (1-3 cm) lenses, mostly in the lower parts of the cores.

The stratigraphy of the cores is illustrated in Figure 6.3. Clay was found at the base of all cores, overlain by a fine, amorphous sediment with approximately 40% organic content in most cores (Figure 6.4). Mollusc shells are present at the base of this layer in some cores (CC-L, CC-O, CC-P; Figure 6.3). This grades upwards into more organic material, which varies among the cores: *Drepanocladus* peat overlain by fine herbaceous-sedge peat in CC-L, highly humified black peat in CC-M, CC-S Shore, CC-N Shore, CC-N1, CC-N2 and CC-O; and in core CC-P, humified black peat overlain by fine herbaceous-sedge peat and then *Drepanocladus* peat. In all cores, there is a subsequent layer of *Sphagnum* peat, which extends to the present surface in cores CC-M, CC-N1 and CC-P. In CC-L, CC-S Shore, CC-N Shore, CC-N2 and CC-O a woody peat forms the uppermost layer.

CC-P, 25 m from the western edge of the peatland, contains the deepest accumulation of sediment with high organic content (268 cm) and was chosen for

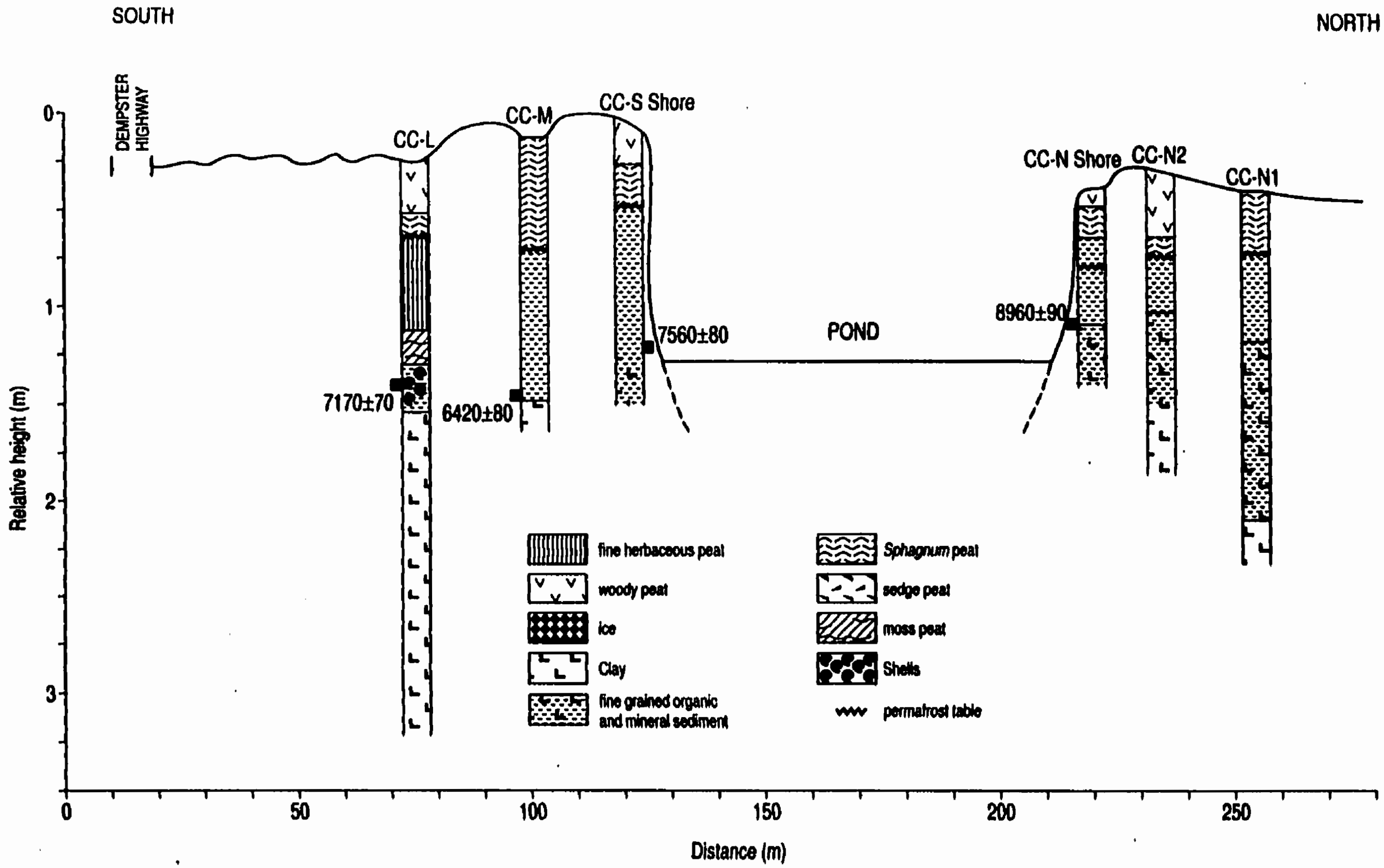


Figure 6.3. Stratigraphic profiles of Campbell Creek peatland: a) north-south; b) east-west (next page).

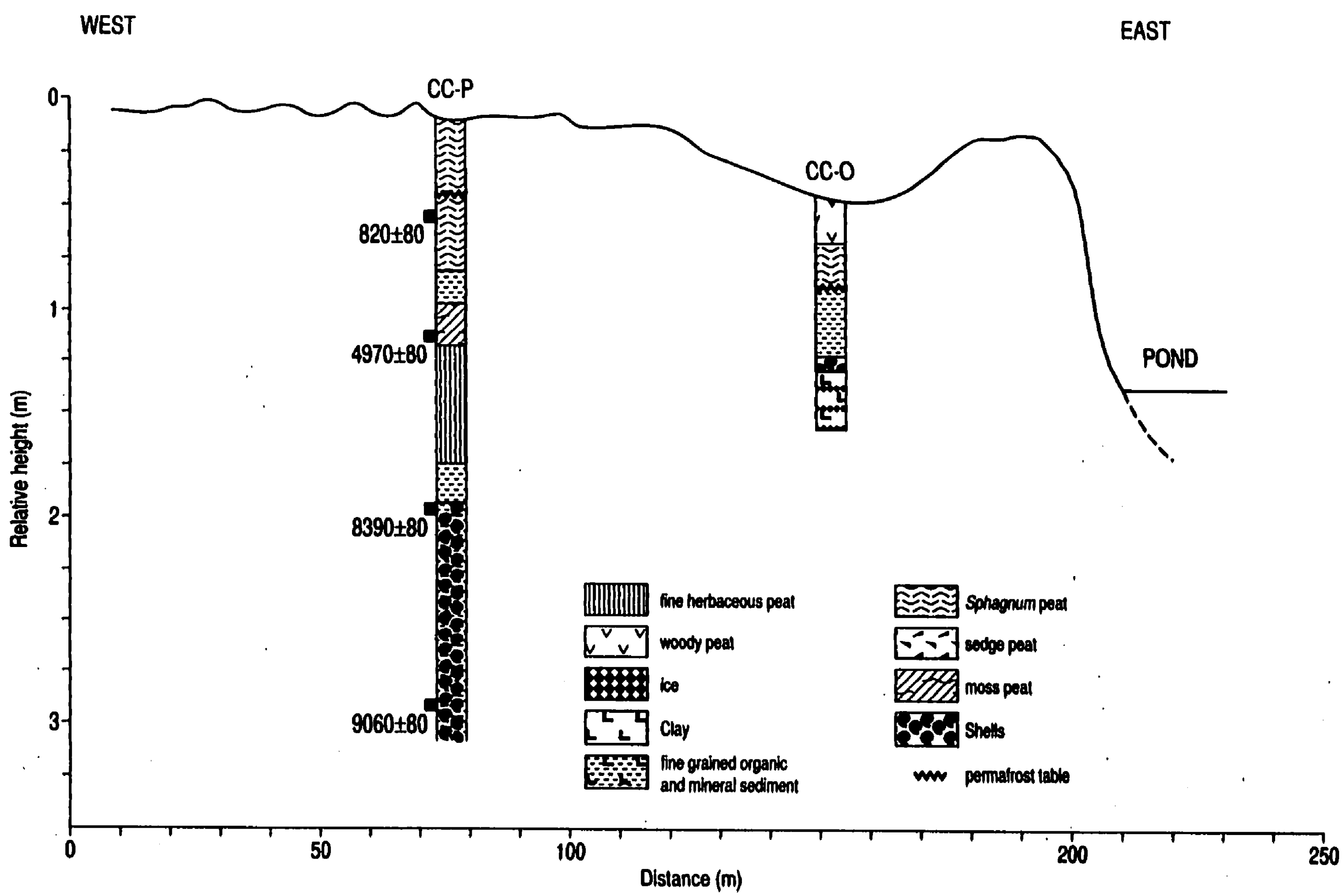


Figure 6.3 (b). East-west stratigraphic profile, Campbell Creek peatland.

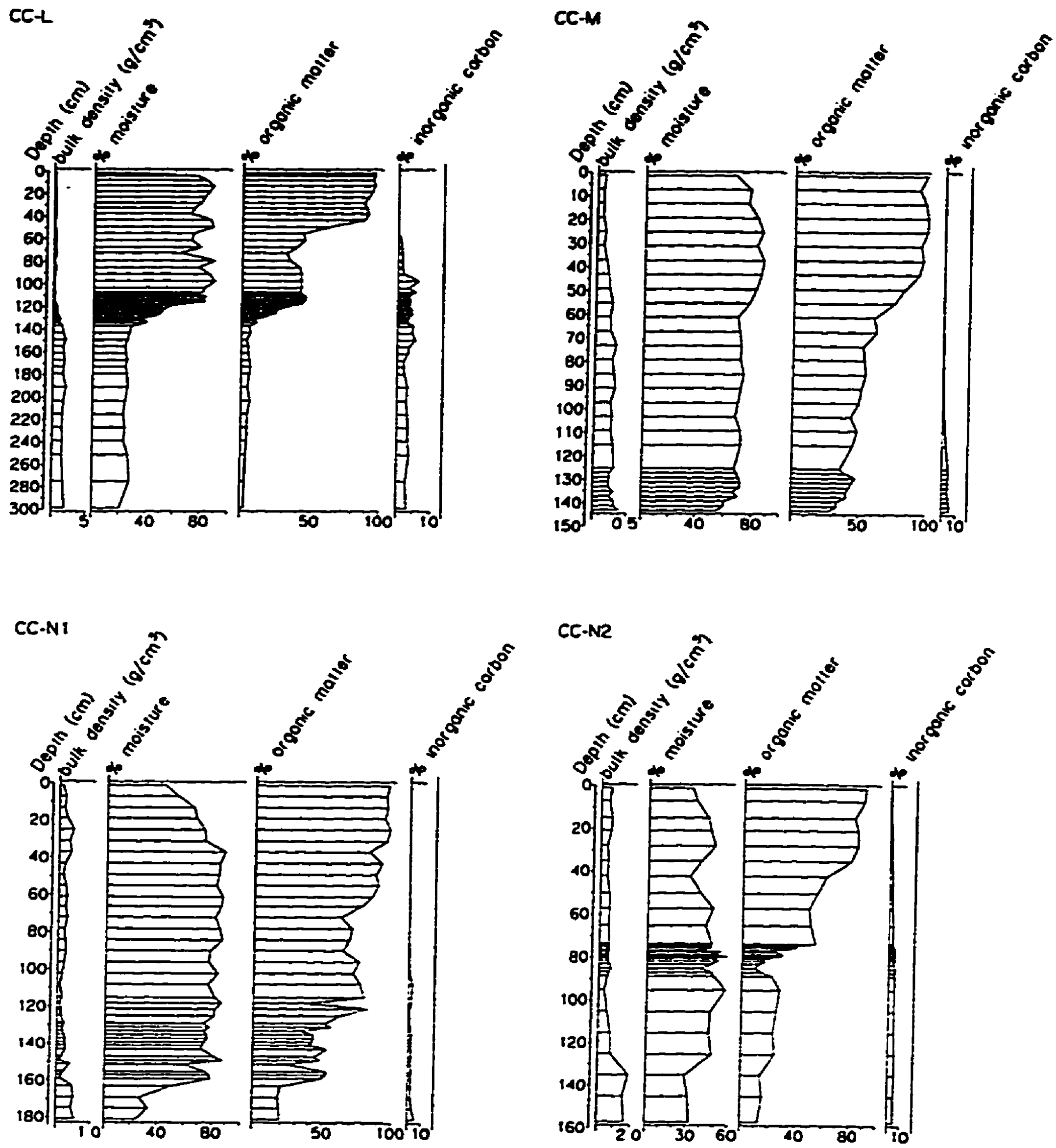


Figure 6.4. Profiles of the six cores and two sections showing bulk density, moisture (% of fresh weight), organic matter (% of dry weight) and inorganic carbon (% of dry weight). Zero point on each diagram represents the peat surface at the coring site. Continued on next page.

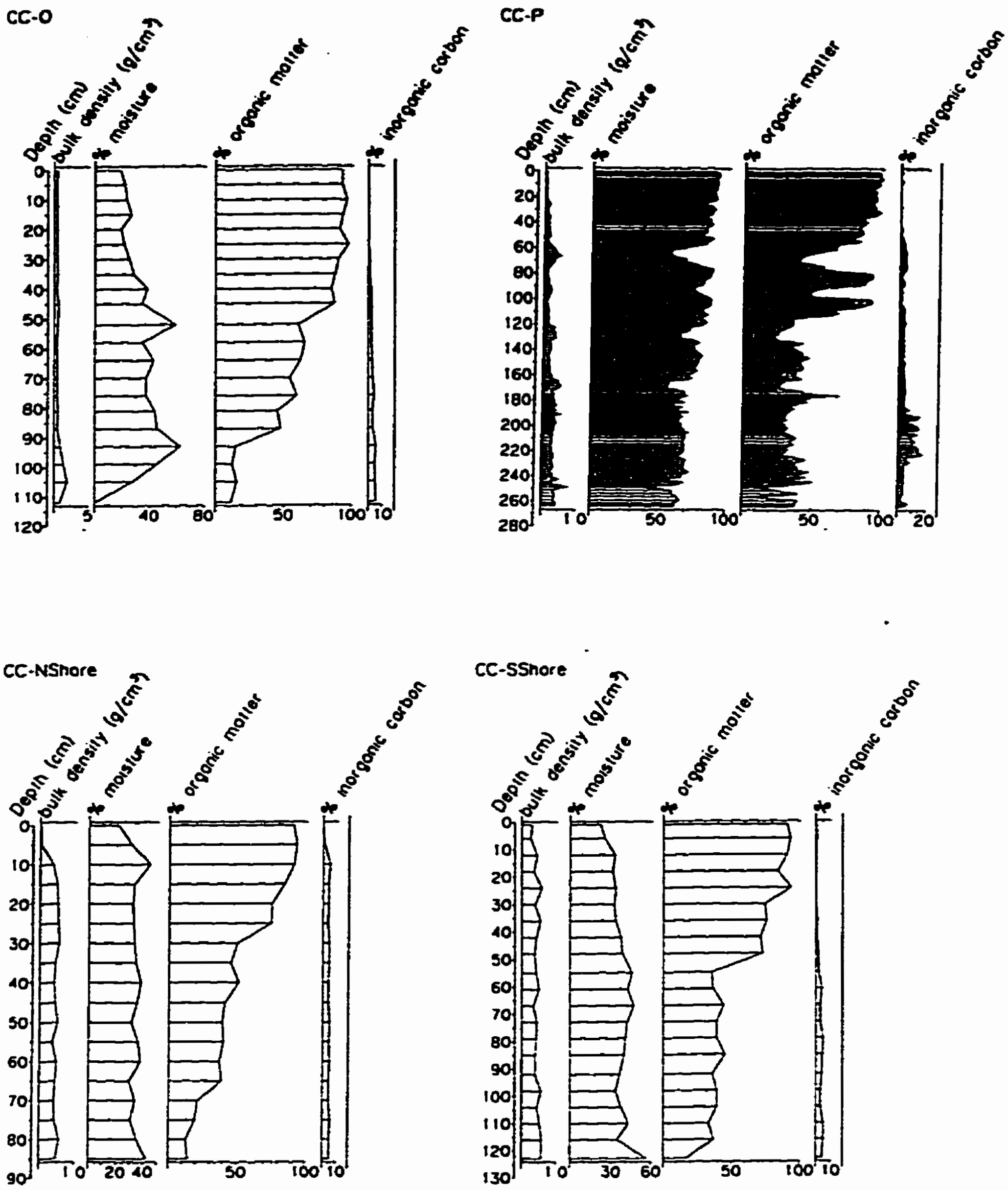


Figure 6.4. (Continued from previous page)

detailed pollen and macrofossil analysis. CC-L, 50 m from the southern shore of the pond, provided 135 cm of material with high organic content, compared with less than 1 m in most other cores, and was also analyzed for pollen and macrofossils.

Radiocarbon dates from this peatland are listed in Table 6.1. The oldest age reported is 9060 ± 80 (WAT-3027) from the base of CC-P. Other basal dates range from 4370 ± 70 (WAT-2812) from CC-N1, to 8690 ± 90 (WAT-2690) from CC-N Shore. Peat accumulation rates have obviously varied considerably among coring sites, since at CC-N Shore only 80 cm has accumulated in the past 8700 years, while at CC-P 50 mc has accumulated in 820 yrs (WAT-3028).

6.2.2 Pollen, spores and plant macrofossils

Core CC-P (Figures 6.5 and 6.6): This core was collected from a *Sphagnum fuscum* hummock near the western edge of the peatland. *Rubus chamaemorus, Betula glandulosa, Vaccinium vitis-idaea, Dryas integrifolia* and *Empetrum nigrum* are abundant around the coring site. The pollen stratigraphy from this core can be divided into 3 zones based on a dendrogram produced in the CONISS program in TILIA, using a stratigraphically constrained cluster analysis.

Table 6.1. Radiocarbon ages from the Campbell Creek peatland.

Core	Depth (cm)	Radiocarbon age[a] (yr B.P.)	Laboratory number[b]	Dated material
CC-N Shore	75-80	8690 ± 90	WAT-2690	Organic/mineral sediment
CC-S Shore	115-120	7560 ± 80	WAT-2691	Organic/mineral sediment
CC-L	112-117	7170 ± 70	WAT-2810	Organic/mineral sediment
CC-M	132-137	6420 ± 80	WAT-2811	Peat
CC-N1	155-160	4370 ± 70	WAT -2812	Peat
CC-P	48-52	820 ± 70	WAT-3028	Peat
	103-108	4970 ± 80	WAT-2813	Peat
	186-192	8390 ± 80	WAT-3026	Organic/mineral sediment
	258-265	9060 ± 80	WAT-3027	Organic/mineral sediment

[a]Ages are corrected for isotopic fractionation and are reported with 1σ.

[b]Laboratory designation: University of Waterloo Radiocarbon Laboratory.

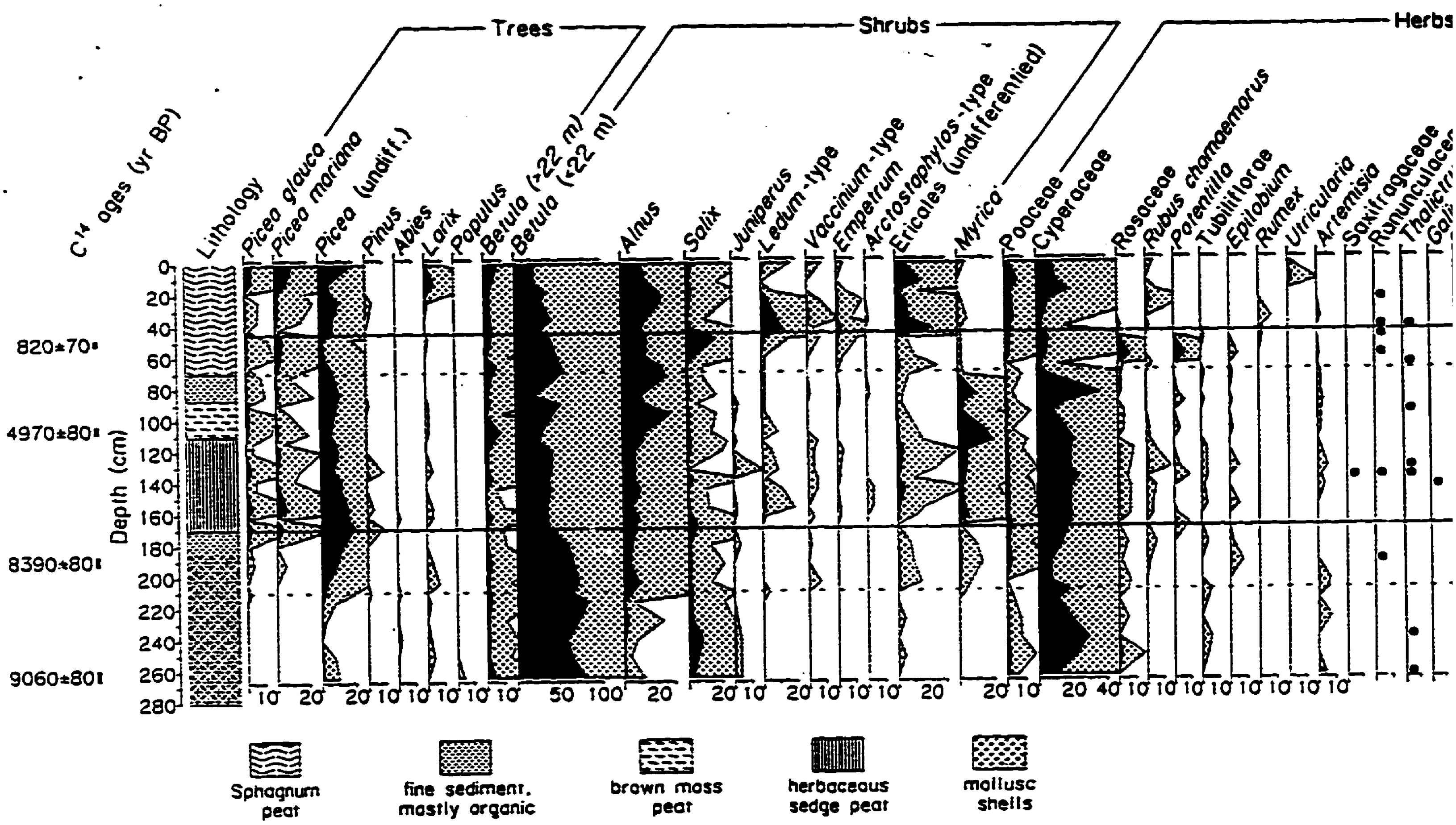
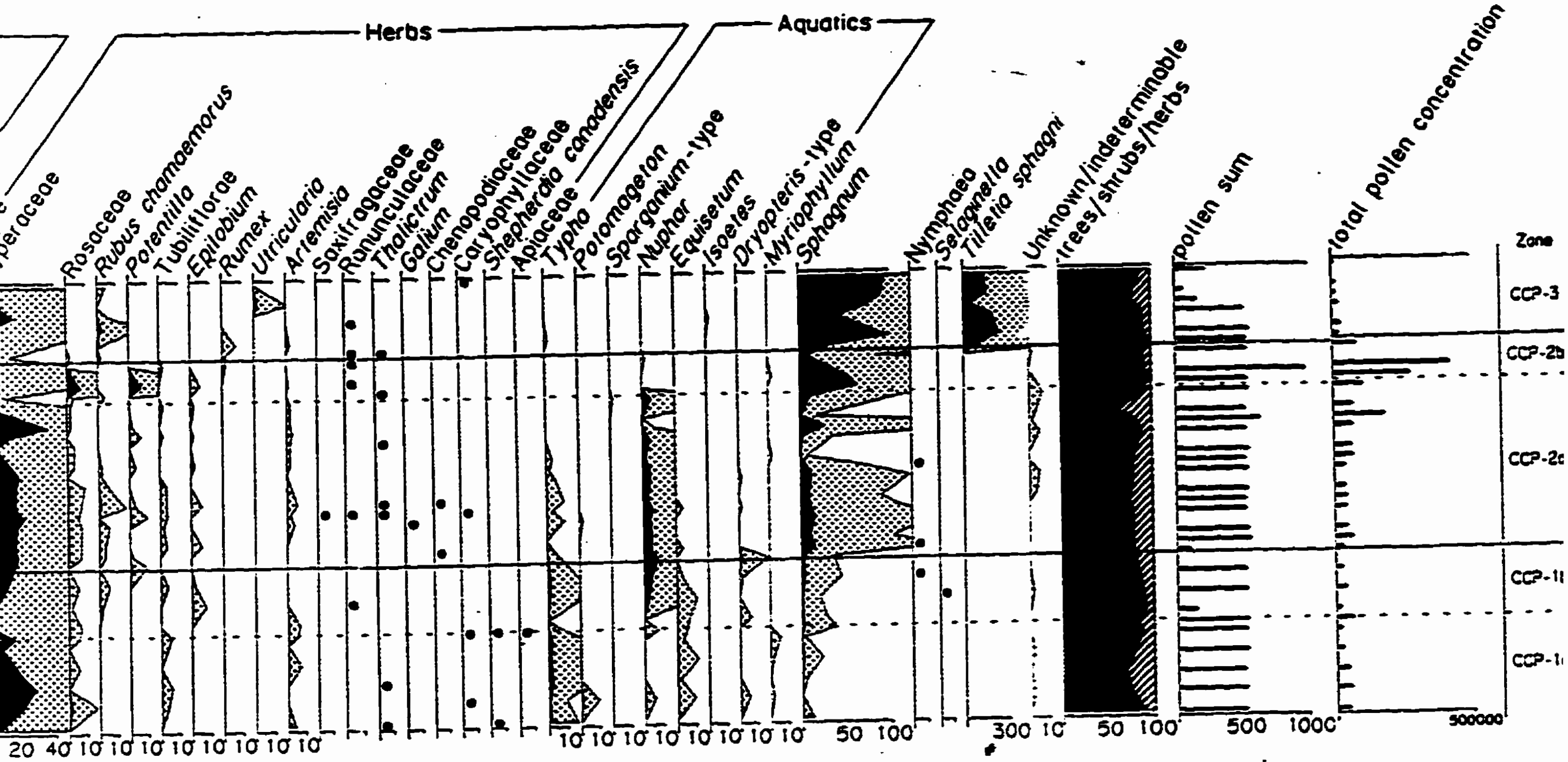


Figure 6.5. Pollen diagram for core CC-P. "Undiff which could not be identified to further taxonomic less.



...iagram for core CC-P. "Undifferentiated" refers to grains ...identified to further taxonomic levels; "•" indicates 2% or

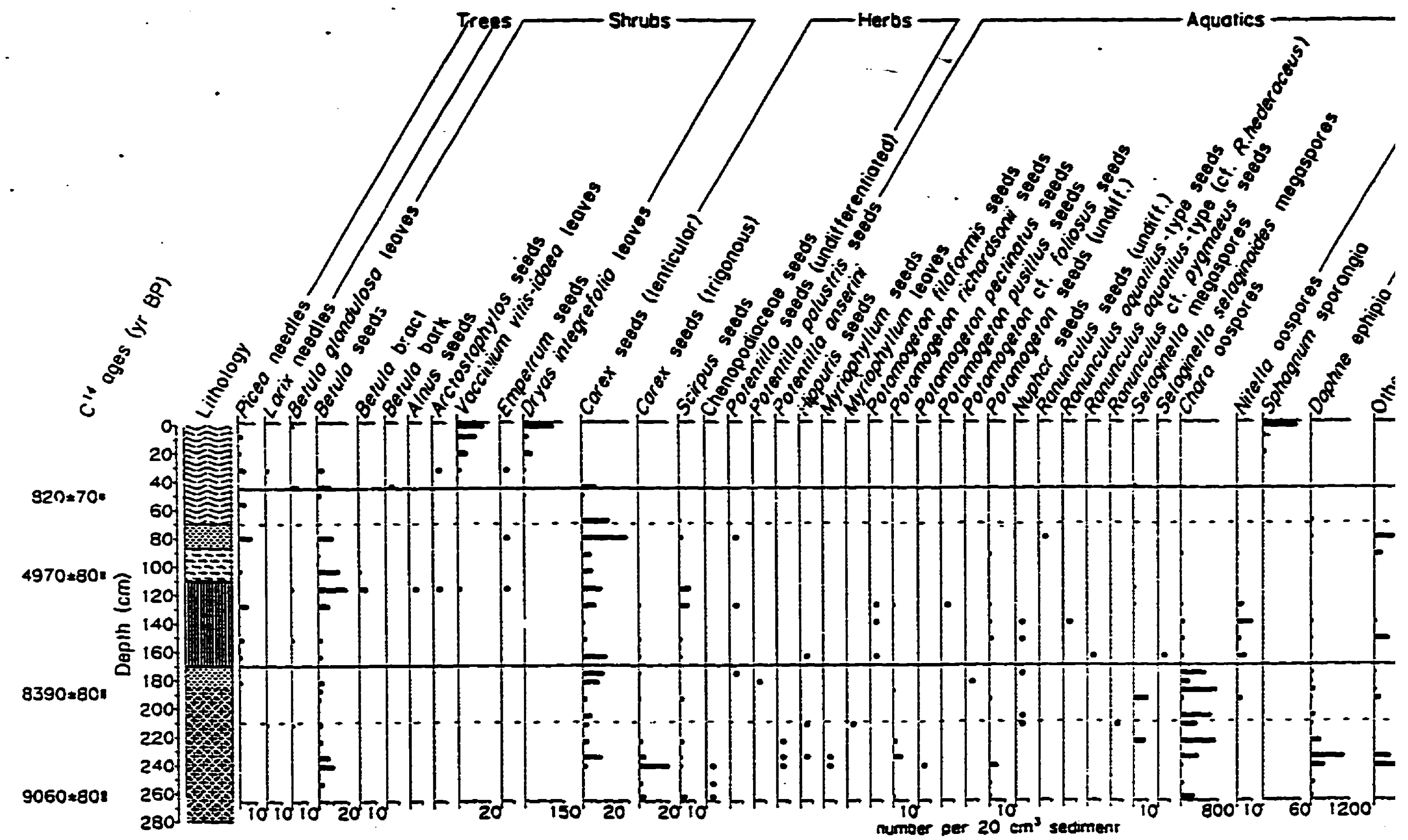
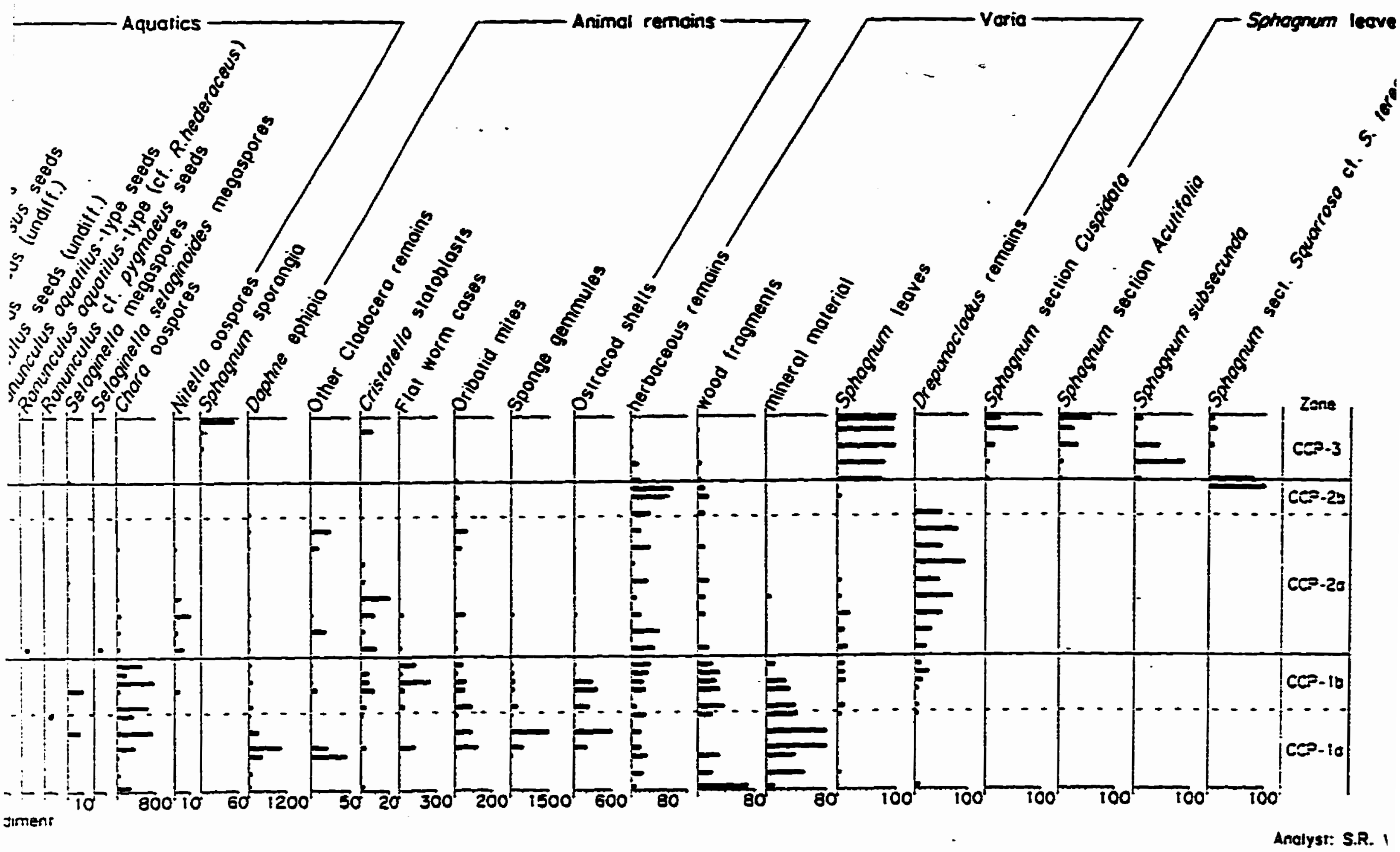


Figure 6.6. Macrofossil diagram for core CC
seeds that could not be identified to species;
of sediment. See Figure 6.5 for lithology key



fossil diagram for core CC-P. "Undifferentiated" refers to
ot be identified to species; "•" indicates less than 3 per 20 cm^3
Figure 6.5 for lithology key.

Zone CCP-1 (268-170 cm) is dominated by *Betula* (up to 80%) and Cyperaceae (up to 20%). *Typha* and *Equisetum* pollen are present throughout the zone, as are oospores of *Chara* in the macrofossils samples. Two subzones are recognized based on changes in the values for other key taxa. Zone CCP-1a (268-210) features peaks in the percentages of *Salix, Potamogeton,* and *Myriophyllum* pollen. *Picea* and *Alnus* each make up less than 3% of the total pollen.

Macrofossil samples from this part of the core contain seeds of *Betula* and *Carex* as well as those of a variety of aquatic plants, including *Hippuris* cf. *vulgaris, Myriophylum, Potamogeton richardsonii, P. pectinatus* and other *Potamogeton* seeds which could not be identified to species. The remains of a number of invertebrates typical of fresh open water, including *Daphnia* and other cladocera, oribitid mites, freshwater sponges, ostracods and molluscs, are common throughout zone CCP-1, and are especially abundant in this subzone.

The transition to subzone CCP-1b is marked by the first presence of *Myrica*, the beginning of the *Picea, Alnus* and *Nuphar* increases, and a slight rise in Ericales pollen. *Nuphar* seeds are found in several samples from this zone, while *Chara* oospores remain abundant, and the remains of aquatic invertebrates are still strongly represented but in fewer numbers.

The sediment matrix is composed mostly of mineral material in subzone CCP-1a, with wood and herbaceous remains becoming more dominant in subzone CCP-1b (210-170 cm). Organic carbon content is 30-45% throughout most of the zone, while inorganic carbonate values are relatively high, peaking at 15% between 220-185 cm, where mollusc shells are most abundant.

Zone CCP-2 (170-45 cm) is marked by a rise in *Picea* pollen from values of less than 3% to 10-20%. *Picea* needles are also present throughout the zone. *Betula* pollen percentages decline to 35-40%, while *Myrica* increases to a peak of 18%, and *Alnus* continues to increase, reaching peaks of up to 25% at 110 and 75 cm. This zone is also divided into two subzones. In CCP-2a, Cyperaceae pollen values remain high, *Nuphar* pollen is present at values of 5-6%, and *Typha* is present up to 90 cm, though at slightly lower values than in the previous zone. Ericales pollen increases, but the total of all types is never more than 6%. *Sphagnum* spores reach values of 10-13%.

Carex and *Betula* are still the dominant seed types found in the macrofossil samples, while seeds of *Scirpus* and various species of *Potamogeton* are also common. *Chara* oospores and invertebrate remains are present in most macrofossil samples from this subzone, but less abundant than in CCP-1. Leaves of *Drepanocladus* mosses (mostly *D. aduncus* with some *D. fluitans*) make up most of the sediment matrix.

Zone CCP-2b (70-45 cm) is distinguished by brief peaks in *Salix* (15%), *Potentilla* and other Rosaceae pollen, and sharp declines in the percentages of *Myrica* and *Nuphar* pollen. There is a switch in the matrix composition as well, from *Drepanocladus* peat to a dominance of sedge and other herbaceous remains.

Zone CCP-3 (45-0 cm) is characterized by peaks in percentages of *Ledum*-type, *Vaccinium*-type, *Empetrum* and undifferentiated Ericales pollen, as well as *Larix, Rubus chamaemorus* and Poaceae. There is a dramatic increase in *Sphagnum* spores (up to 95% of pollen + spores). *Tilletia sphagni,* a fungus associated with *Sphagnum,* is also abundant. *Dryas integrifolia* and *Vaccinium vitis-idaea* leaves and *Sphagnum* sporangia dominate the macrofossils, while the matrix is composed mainly of *Sphagnum* leaves with some herbaceous remains and wood fragments. Within this zone there is an apparent succession of *Sphagnum* species, from *Sphagnum* cf. *teres* to *S. subsecunda,* and then species of *Sphagnum* sections *Cuspidata* and *Acutifolia.* The organic carbon content rises to 90-100% in this zone, while inorganic carbon is no longer present (Figure 6.4).

Core CC-L (Figures 6.7 and 6.8): This core is also divided into three zones based on the CONISS results. In zone CCL-1 (140-130 cm) pollen concentrations are very low, and dominated by Poaceae, with increasing *Betula* and a peak in Tubilflorae. The *Larix* pollen peak at the base of this zone is misleading because it represents only a few grains. The only macrofossils found in this zone were a few *Chara* oospores in one level. The matrix is made up mostly of mineral detritus, with some herbaceous material. The organic matter content is less than 10% by weight.

Zone CCL-2 (130-55 cm) is dominated by *Betula* (mostly < 22 µm), and the highest values of Cyperaceae found in this core. *Picea,* which is present from the beginning of CCL-1, increases gradually throughout this zone, reaching total

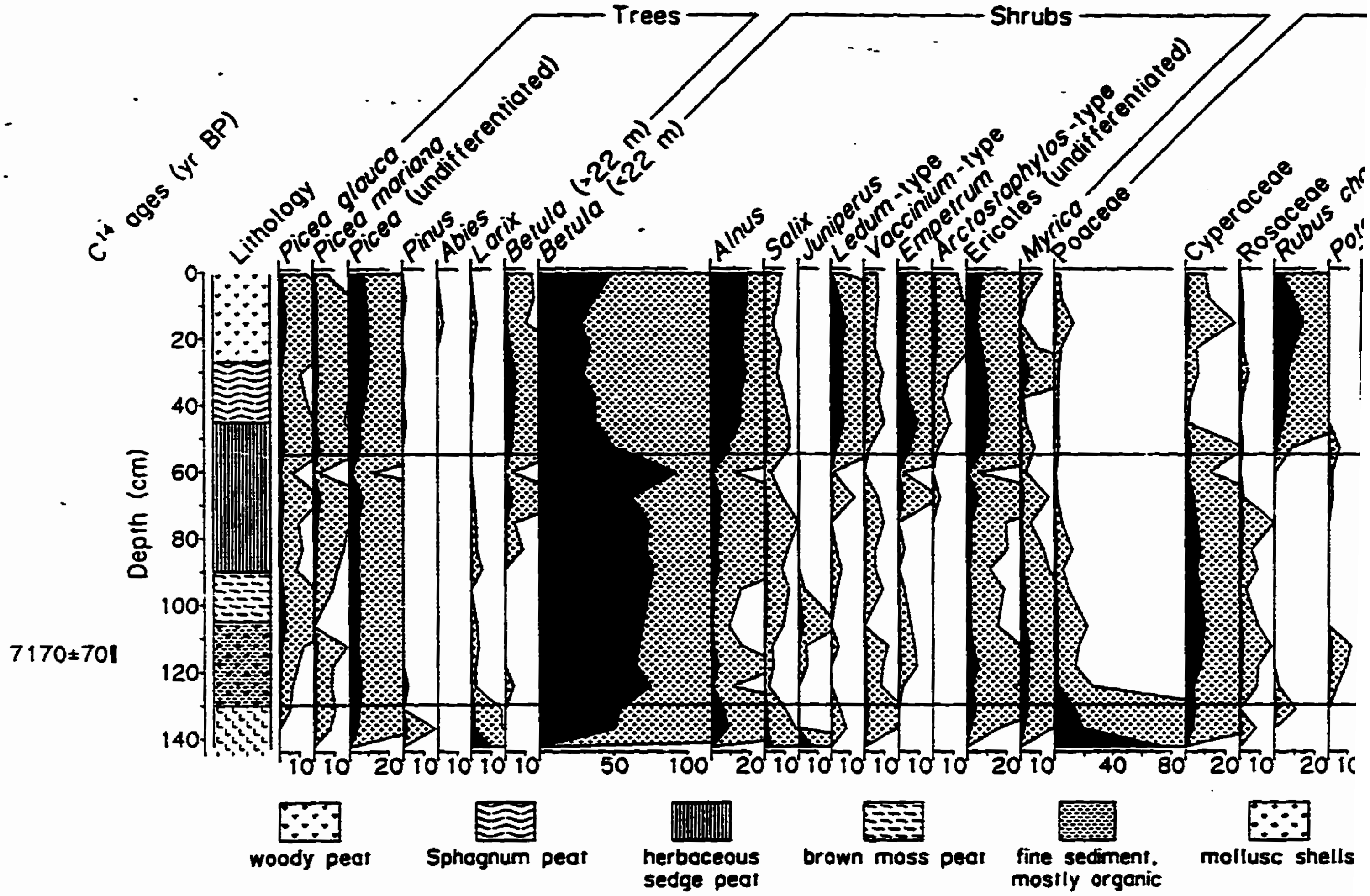
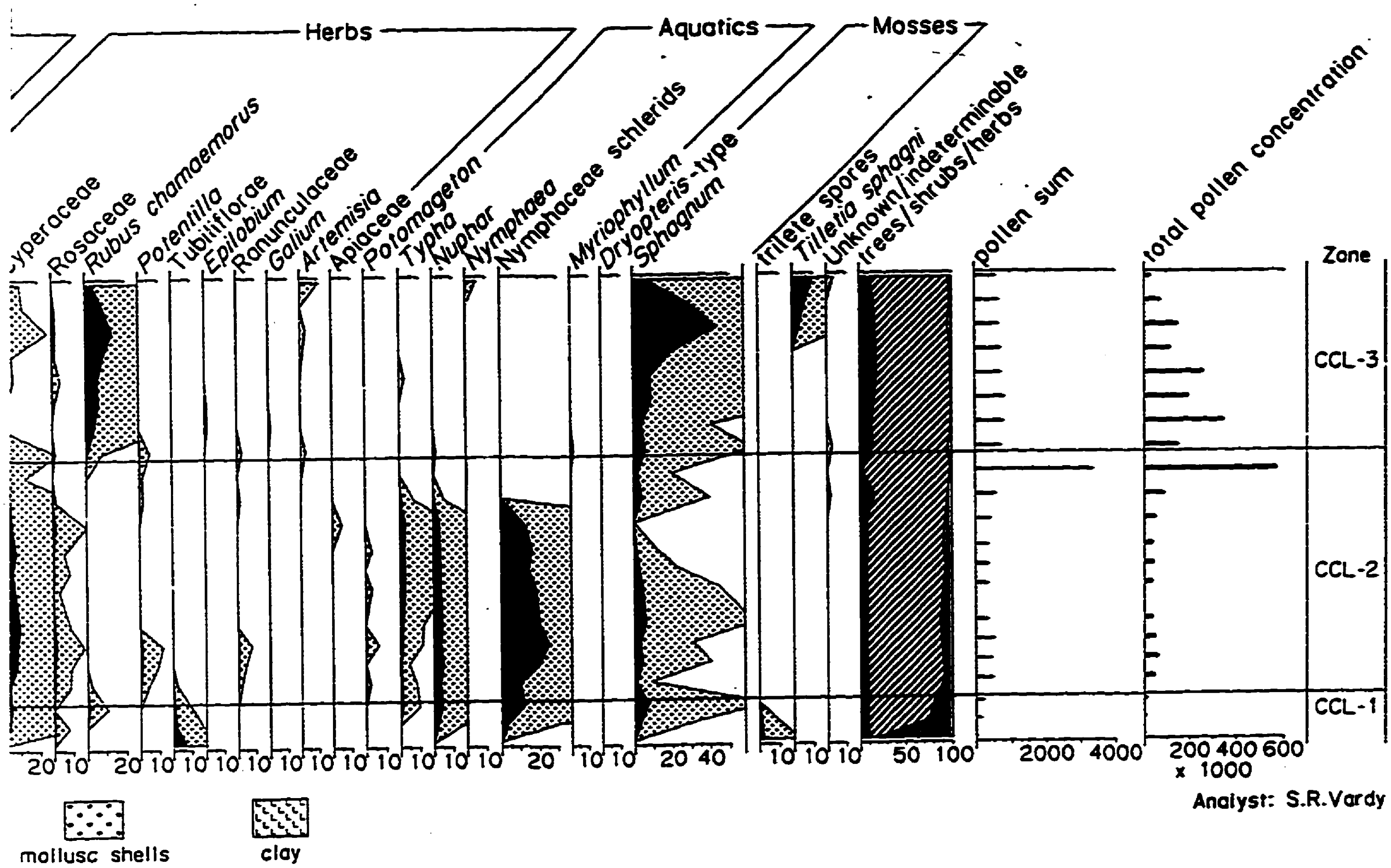


Figure 6.7. Pollen diagram for core



am for core CC-L.

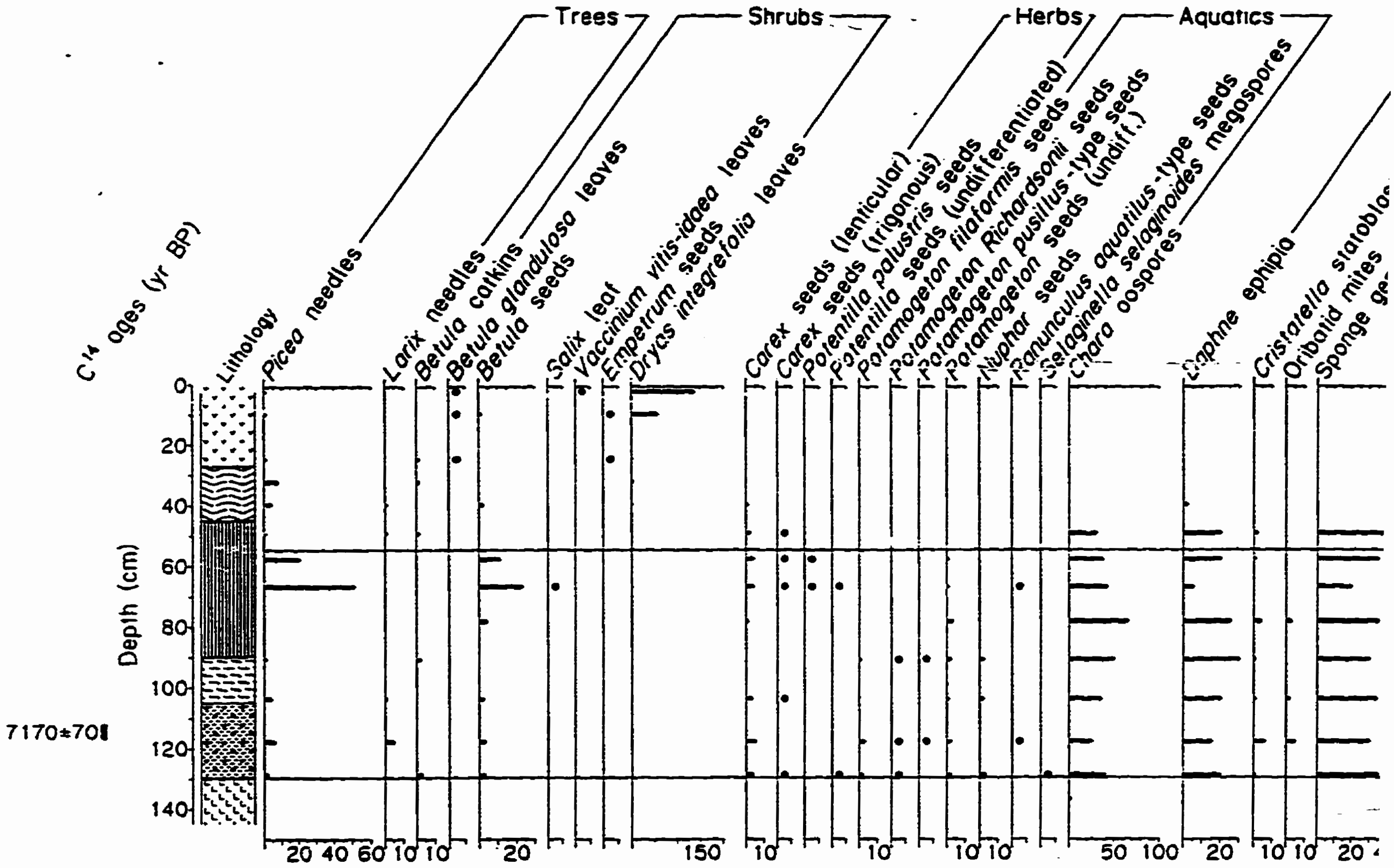
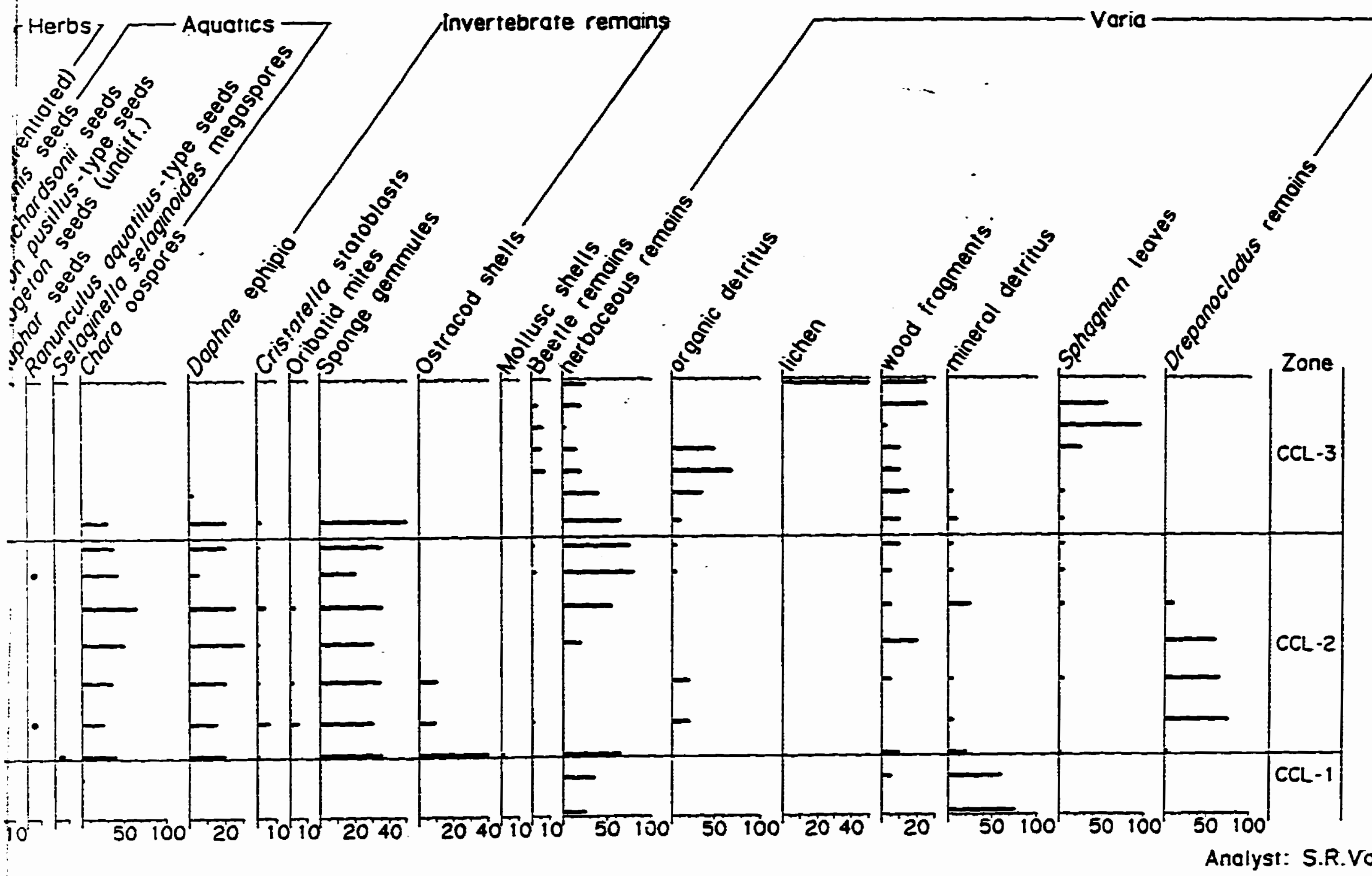


Figure 6.8. Macrofossil diagram for core CC-L. See I



l diagram for core CC-L. **See Figure 6.7 for lithology key.**

values of 10-12% for *P. glauca, P. mariana* and undifferentiated *Picea* grains. *Typha* and *Nuphar* pollen are found in samples up to 65 cm, and *Potamogeton* pollen is present in small numbers up to 75 cm. Nymphaceae schlerids are abundant in the pollen samples. *Betula, Carex, Potamogeton* and *Nuphar* seeds are common, as well as *Picea* needles, *Chara* oospores and the remains of aquatic invertebrates. *Larix* needles in the lower part of the zone confirm the local presence of larch at around 7200 BP. *Drepanocladus* peat (mostly *D. aduncus*) makes up most of the matrix for the lower half of this zone, but is replaced by herbaceous remains above 75 cm. Organic matter content increases sharply at the beginning of the zone and remains around 40% throughout this zone.

Zone CCL-3 (55-0 cm) is marked by a decline in *Betula* and *Carex* pollen, and increases in *Alnus, Ledum, Empetrum, Arctostaphylos* and undifferentiated Ericales. *Rubus chamaemorus* appears at the beginning of the zone and reaches a peak of 14% at about 18 cm depth. *Sphagnum* increases to a peak at the same level. *Nuphar* pollen and seeds and Nymphaceae schlerids are no longer found. *Typha* pollen is absent except for a single tetrad. Above 50 cm there are no *Chara* oospores or aquatic invertebrate remains. A few *Carex* seeds were found up to 40 cm, but otherwise the only plant macrofossils in this zone are *Picea* and *Larix* needles, *Betula* remains, *Empetrum* seeds and the leaves of *Vaccinium vitis-idaea* and abundant *Dryas integrifolia*, the latter being particularly abundant in the top 10 cm. The matrix in the bottom part of this zone consisted of fine, unidentifiable herbaceous remains and organic detritus, with *Sphagnum* leaves dominating from 25-5 cm, and the surface peat composed mostly of lichens and wood.

6.3 DISCUSSION

6.3.1 Peatland history

The paleoecological record from CC-P is interpreted as representing a succession beginning with an initial shallow open water, mineral wetland habitat with aquatic plants such as *Potamogeton, Equisetum, Myriophyllum, Typha* and *Chara. Betula, Salix,* and various herbs probably grew nearby. Submerged mats of *Drepanocladus* later appeared, followed by the invasion of sedges and *Myrica gale,* leading to the development of a fen environment near the coring site, while open water conditions still existed nearby, according to the high values of *Nuphar* pollen and other aquatic indicators. *Picea* was established locally early in this stage. The CCP-2/CCP-3 boundary represents a transition to a *Sphagnum* dominated, ombrotrophic peatland, with a variety of heaths and *Rubus chamaemorus,* similar to the present vegetation at this coring site.

The paleoecological record from core CC-L also suggests an early minerotrophic environment with open water, with early assemblages dominated by Cyperaceae and aquatic species. By at least 7200 BP, *Drepanocladus* mats had developed in an open water system where *Nuphar, Potamogeton* and *Chara* thrived, and *Carex* grew around the edges. *Picea* and *Larix* were both present locally by that time as indicated by needles of both taxa. As in CC-P, this is followed by a decline in aquatic indicators and increases in *Alnus*, ericaceous shrubs, *Rubus chamaemorus* and *Sphagnum* spores, although the transition does not appear as sharp here. A transition from *Sphagnum* peat to the lichen-*Dryas integrifolia* vegetation that dominates the surface at the site today is apparent at the top of the macrofossil diagram, although in the pollen core it is indicated only by a decrease in *Sphagnum* spore percentages.

The stratigraphies of other cores from this peatland indicate that a successional sequence from minerotrophic conditions with open water to *Sphagnum*-dominated ombrotrophic peatland occurred across most of the basin. However, there were local variations in plant communities, and the *Drepanocladus* stage is not represented in all cores.

The origin of the pond in the middle of Campbell Creek is difficult to determine. The pond appears to be on the order of 2-3 m deep, since emergent aquatics including *Potamogeton* spp. and *Nuphar* grow in the middle of it. The wetland surface slopes very gently to the southeast, with the wet southeastern area being closest to the water table. While it is possible that the pond is a thermokarst feature, it might also be a remnant of the original open water body that occupied the entire basin early in the Holocene. It seems to be still undergoing infilling at the southeastern edge, while around the rest of its perimeter it is bordered by bluffs of peat up to 1 m above the water surface.

6.3.2 Comparison with the regional record

Picea pollen in nearby lake sediments (M-Lake and TT-Lake) reaches the 10% threshold estimated to indicate local presence of spruce trees at 8200-8700 BP (Ritchie, 1977,1984a). Ages of this spruce threshold elsewhere in northwestern North America range from 7900 to 10900 BP, with no obvious directional trend. *Picea glauca* and *P. mariana* increase simultaneously in core CC-P, with the 10% threshold of total *Picea* pollen reached at approximately 8400 BP. Organic accumulation at this site began sometime before 9060 BP. *Populus* values are high from 10300-8900 BP at M-Lake, and from 10300 to 8000 BP at TT-Lake, peaking between 10300 and ~9500 BP (Ritchie, 1977, 1984). *Populus* is present only at the base of core CC-P, disappearing soon after 9000 BP.

Myrica arrives at about 8000 BP at TT-Lake, and declines to very low percentages after ~5500 BP. In both CC-P and CC-L, it arrives at about the same time as *Picea* (about 8400 BP in CC-P). It peaks at around 5000 BP in zone CCP-2 and subsequently declines, whereas it never reaches values more than about 5% in CC-L.

Alnus first appears in pollen diagram of M-Lake at 8000 BP and at TT-Lake at ~7500 BP, rises sharply to >40% at ~6700 BP and ~5600 BP, respectively

(Ritchie, 1977, 1984). In CC-P it is present at less than 2% below 220 cm, then increases to about 10% until 120 cm (ca. 5000-5500 BP). Above this level it increases to values of up to 25%. It is present at low values in CC-L until 60 cm, when it increases to 18-20%. *Juniperus,* which peaks in M-Lake and TT-Lake between 10000 - 6000 BP, is present in CC-P in most levels before approximately 5000 BP, but in lower percentages than in lake sediments. It is recorded in CC-L only between 140 and 90 cm.

Typha is rarely frequent enough in sediments to produce continuous percentage curves, as it does in CC-P from 268-100 cm (until ca. 5000 BP), and in CC-L from 135-60 cm. Its northern limit is now several hundred kilometers south, near Norman Wells, but it occurred in NW Canada (Hanging Lake, Yukon) as early as 12,000 BP (Cwynar 1982). It persisted at several sites in The Mackenzie Delta area until 5000 BP, but was recorded at M-Lake and TT-Lake only from 10600 - 8900 and 9900-7800 BP, respectively. The record from CC-P therefore extends the record of its occurrence in this immediate area by another 2800 yr. It is also present in core CC-L until the time of the transition to *Sphagnum*-Ericad dominated assemblages. *Typha* is considered a strong indicator of conditions possibly warmer than present. Ritchie (1984) suggests it might require summer temperatures warmer by up to 6°C (May-August) and 350 more degree days above 5°C.

6.4 CONCLUSIONS

Organic sediments began to accumulate at the CC-P coring site by at least 9100 BP. A hydroseral succession occurred starting with open water conditions with aquatic plants and animals, followed by a transition to a fen with brown moss and sedges, and a change to a *Sphagnum* dominated peatland similar to present, with abundant heaths, *Betula* and other shrubs, and *Rubus chamaemorus*. The beginning of this change in the peatland seems to coincide with the larger-scale regional vegetation changes interpreted by Ritchie, Spear and others as representing a deterioration of climate. It may therefore have been triggered, at least indirectly, by the change in climate. Aggradation of permafrost at this time, initiated by the climatic cooling or the development of a thin *Sphagnum* cover, or both, may have initiated up the transition, by raising the peat surface. The more sudden change at 800 BP does not coincide with any known climate event recorded in the pollen stratigraphy form lake sediments. This probably represents a further raising of the peat surface above the water table by the combined effects of permafrost aggradation, ground ice build-up and peat accumulation, creating conditions more suitable to a *Sphagnum*-heath community than the previous *Drepanocladus* - sedge community.

7. DISCUSSION

7.1 COMPARISON OF RESULTS WITH REGIONAL VEGETATION AND CLIMATE HISTORY

One of the objectives of this thesis is to compare the reconstructed histories of peatland development with the regional paleoenvironmental history known from lake sediment studies, in order to understand the response of the peatlands to climate change. To facilitate this comparison, the most significant events in the regional and peatland records are summarized in Table 7.1. The degree to which the regional vegetation history is detected directly in the pollen and macrofossil records from the peatlands will also be considered.

In the Kukjuk peatland cores, changes in the regional vegetation are not well represented in the pollen record, since there is little or no fluctuation in the pollen curves of trees and non-ericoid shrubs. The exception is an increase in *Alnus* pollen to about 30% at 220 cm, approximately 6300 BP. *Picea* pollen totals are consistently less than 10%, and *Larix* pollen is recorded only in small numbers at a few levels. Peaks in tree pollen during the period of suggested treeline advance are also absent in the 9000 yr record from the nearby Bluffers Pingo site (Spear, 1993), so it is possible that treeline never advanced this far northeast on the Tuktoyaktuk Peninsula. However, macrofossils of *Picea* and *Larix* in core KJ-B confirm the local presence of these trees. The failure of the

Table 7.1. Comparison of events identified in the Kukjuk and Campbell Creek peatlands with the regional pollen record and inferred Holocene climate history.

yr. BP	REGIONAL RECORD FROM LAKE SEDIMENTS		KUKJUK PEATLAND		CAMPBELL CREEK PEATLAND*	
	Vegetation	Inferred climate	Peatland	Regional	Changes in peatland	Regional signal
0-4500	No evidence of significant vegetation change	Only minor fluctuations	Very low peat accumulation rates since 4000 BP	No change in regional pollen spectra	Development of local, dry lichen surfaces in parts of peatland	No change in regional pollen spectra
4500-5000	Treeline retreated to present position; decline in *Picea* pollen north of treeline	Rapid cooling to conditions similar to present	Surface raised by permafrost & ice build-up; development of high-centre polygons and ombrotrophic conditions	*Picea* and *Larix* needles and *Typha* pollen disappear	Rapid change to *Sphagnum* dominated peatland ~5000; aggradation of permafrost	
5000-9000	-Continuous boreal forest extends further north than today. -*Typha* at some sites -*Myrica* peaks 9000-6500 BP -*Alnus* rise 5700-7800 BP	-Climate warmer and more humid than present (see text for estimated values) -gradual cooling after 8000 BP.	Organic sedimentation began at ~7200 BP. -transformation from open-water to fen ~6300 BP -first affected by permafrost after 6300 BP, when low-centre polygons developed	-*Picea* and *Larix* needles confirm presence of these trees locally. - *Typha* present locally - *Alnus* increase around 6500 BP	Transformation to fen with *Drepanocladus* and sedges.	-*Alnus* increase approx. 5000-5500 BP - *Typha* present until ~ 5000 BP
9000	-*Picea* reaches Sleet Lake and possibly Beaufort Sea coast. -*Larix* arrival					*Picea* increase - 8500-9000 BP
9000-10900	-*Picea* arrival -*Populus* peak -*Typha* common, 9000-10000 BP	-estimated summer temp. 6°C higher than present			Shallow open-water with aquatic flora and fauna	*Populus* pollen at base of core ->9000 BP
11000-11500	-Sharp increase in *Betula* pollen influx -increase in ericad pollen, *Sphagnum* spores and *Rubus chamaemorus*	-Rapid climate warming begins - greater summer precipitation as well as temperatures				
11500-15000	herb tundra	Deglaciation and gradual warming of climate				

pollen record from the peatland to reflect this may be due in part to the over-representation of *Betula* pollen, which effectively drowns out changes in other taxa. There is no evidence here to indicate that trees were abundant in the vicinity of Kukjuk peatland within the past 7200 years; the macrofossils alone indicate the presence of at least scattered trees.

The main pollen evidence in the Kukjuk cores to support the suggested early Holocene insolation maximum is the presence of *Typha* pollen, hundreds of kilometers north of its present range, until approximately 4500 BP. *Picea* and *Larix* needles provide evidence that at least a few trees grew in the vicinity during this period. The timing of the most significant vegetation change in the peatland itself, from minerotrophic wetland to ombrotrophic peatland, also corresponds to inferred regional climatic deterioration at 4500-5000 BP, and may have been triggered by the aggradation of permafrost in response to climate change, as discussed in the following section.

Regional vegetation changes are more apparent in the Campbell Creek pollen records, with changes observed in the percentages of *Picea, Larix, Alnus, Myrica, Juniperus* and *Populus* as well as *Typha*, all of which are comparable to regional changes in the lake sediment reconstructions of Ritchie (1977, 1984, also Ritchie and Hare, 1971). The initial arrival of *Picea* in the area is represented by the increase in *Picea* pollen at the expense of *Betula*. Changes in the abundance of

Alnus, Myrica and *Juniperus* are also apparent, as well as the presence of *Typha* pollen almost continuously until after 5000 BP.

The timing of the switch from fen to ombrotrophic peatalnd conditions in CC-P is less clearly related to the regional climate change. Although the transition does appear to have begun gradually soon after 4500 BP, it happened much more quickly after 800 BP, signalling a change in hydrological conditions that cannot be explained by any known climatic event. Instead, it may have been due at least partly to permafrost aggradation and expansion of freezing ground water, perhaps triggered by the presence of a thin insulating layer of *Sphagnum* peat.

7.2 THE EFFECTS OF PERMAFROST ON PEATLAND DEVELOPMENT

The second main objective of this research is to investigate the general processes of peat accumulation in permafrost peatlands, to determine how those processes are affected by permafrost and ground ice.

Permafrost seems to have played an important role in the development of the Kukjuk peatland. The aggradation of the permafrost and the build-up of ground ice are largely responsible for the surface of the peat being raised above the local water table. The hydrology of the early mineral wetland was thus altered, creating conditions suitable for a transition to *Sphagnum* peatland.

Zoltai and Tarnocai (1975) hypothesized that the sedge-*Sphagnum* transition found in many permafrost peatlands south of my study area was often associated with a widespread climatic cooling 3000-4000 BP, and the initial invasion of the peatlands by permafrost in discontinuous permafrost areas.

Ice wedge polygons formed in the peatland sometime after permafrost aggraded closer to the peat surface, and eventually evolved into high-centre polygon forms. As the peat surface of the polygon centres was raised further still, due to a combination of ground ice build-up and peat accumulation, conditions over much of the peatland became too dry for *Sphagnum* growth, and the ground cover changed to lichen and dwarf shrubs. Peat accumulation rates in the centre of at least one polygon (KJ-B) have been very low for the past 4000 years or so. Meanwhile, peat is still actively accumulating in other parts of the peatland, especially in the ice-wedge trenches and in the wetter moat surrounding the peatland.

Similar reductions or cessations in peat accumulation rates have been reported from other permafrost peatlands. Zoltai *et al.* (1988) noted that a period of a high accumulation rate was found in the lower halves of several subarctic peatlands, with a marked reduction in accumulation rates in higher levels that may signal the establishment of surface-dry peatlands elevated by permafrost.

Permafrost probably played a significant role in the developmental history of the Campbell Creek peatland as well, although there is no polygonal patterning at this site and ground ice is restricted to thin (1-3 cm) lenses in most cores. The transition from sedge-*Drepanocladus* fen to *Sphagnum* peatland, apparent in both CC-P and CC-L, might be expected to occur with or without the presence of permafrost. However, the transition seems to have occurred quickly in both cores. Assuming this transition is synchronous in these two cores and across the peatland, where it is represented by transitions in the stratigraphy of other cores, then it seems likely that it is related to an aggradation of permafrost in the peatland. A positive feedback mechanism may operate in such circumstances: permafrost raises the surface of the peatland above the water table enough for ombrotrophic vegetation, including various *Sphagnum* spp., to become established. As a layer of peat and living moss accumulates, it provides a layer of insulation which promotes further aggradation of permafrost, which in turn raises the peat surface even higher as the water contained in the peat freezes and expands.

The apparent effects of the aggradation of permafrost and development of ground ice in these peatlands is generalized in a conceptual model of the processes of peat accumulation by terrestrialization under climate conditions like those inferred for this region during the Holocene.

The model (Figure 7.1) begins in the early Holocene with a shallow basin (2-3 m deep), which may be topographically confined, or of thermokarst origin (A1). An open water body inhabited by submergent and emergent aquatic

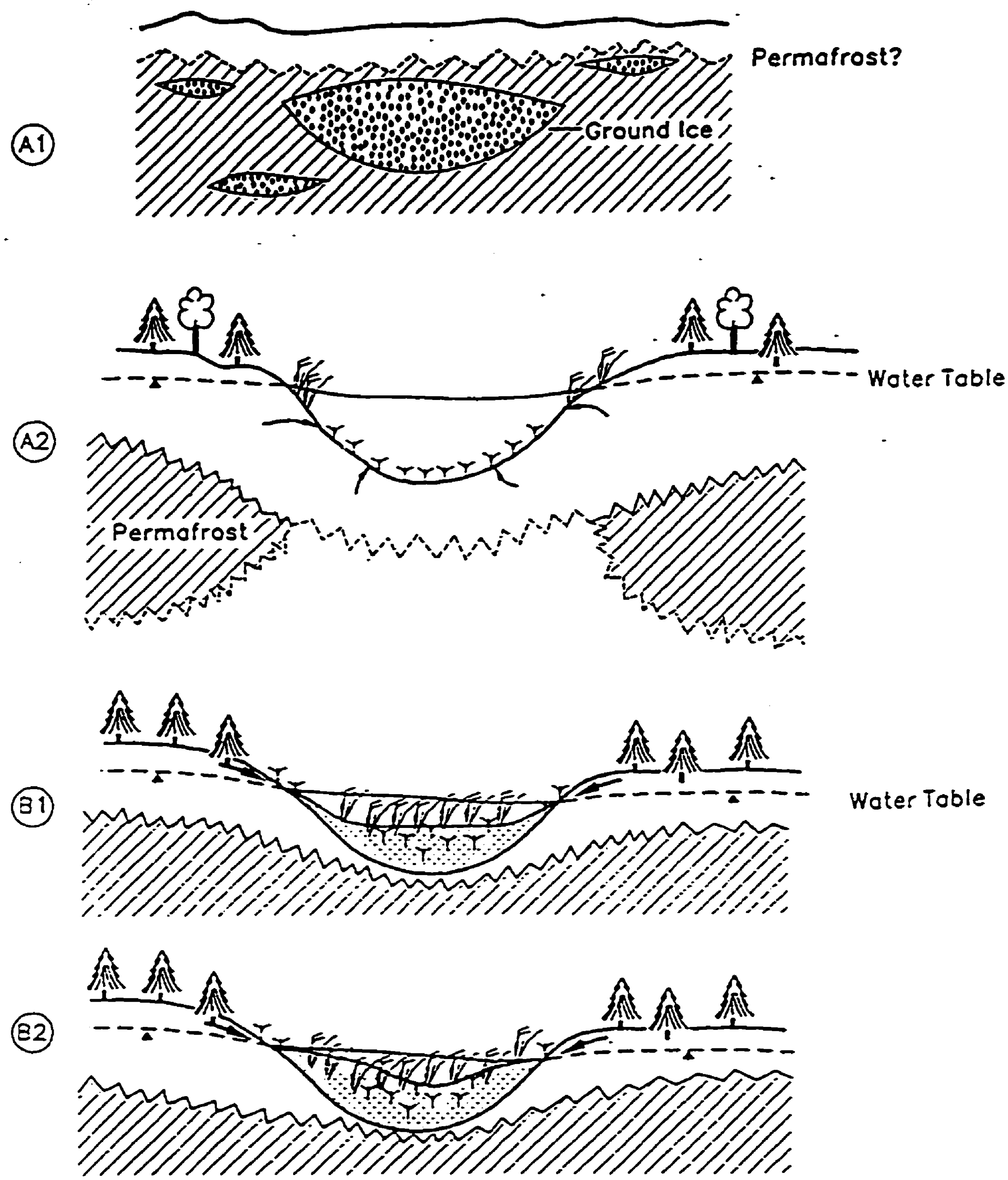


Figure 7.1. Conceptual model of peatland development under a warming, then cooling climate in an area of continuous permafrost. (Continued on following page).

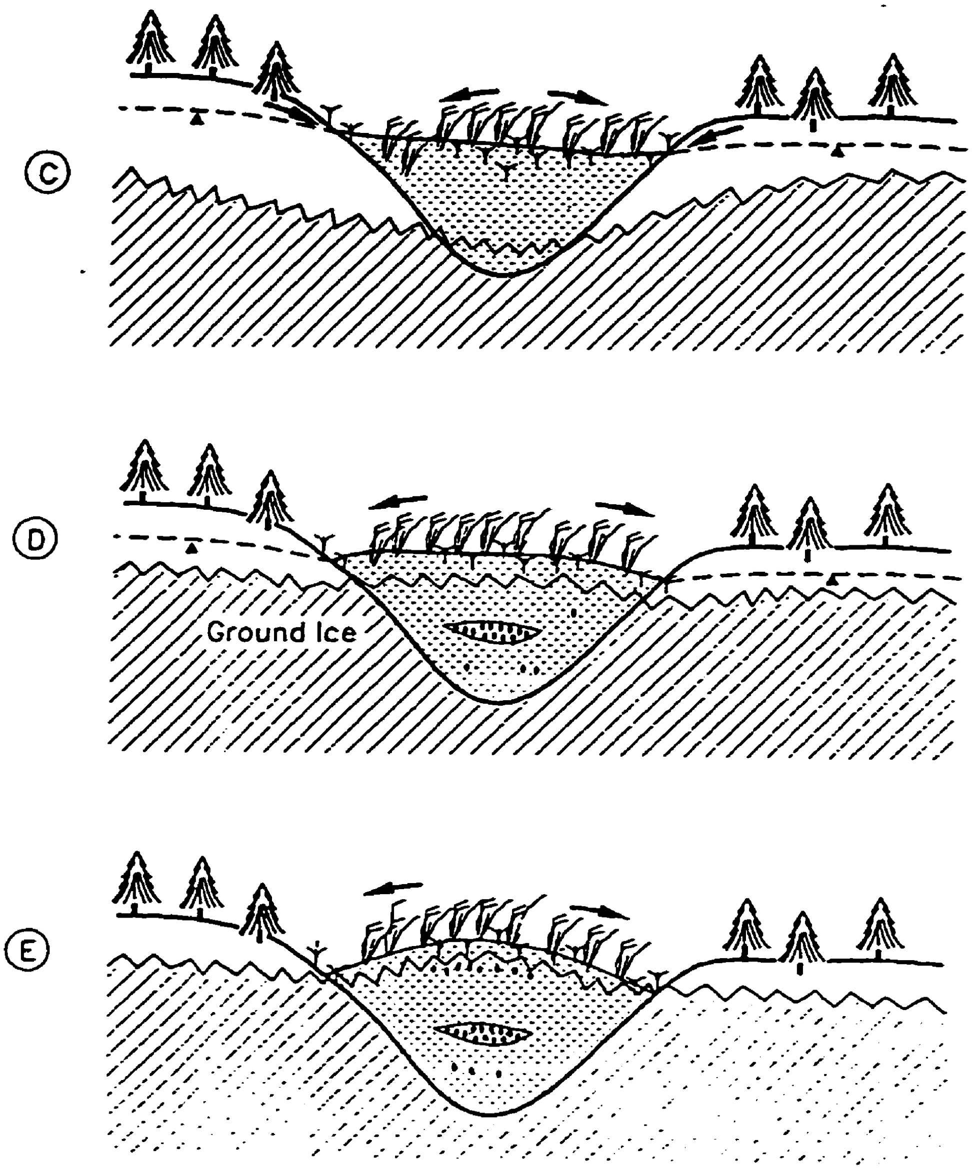


Figure 7.1. (Continued from previous page)

vegetation and aquatic invertebrates occupies the basin originally, while sedges and perhaps *Typha* grow around its perimeter (A.2). Under the warming postglacial climate, permafrost may or may not be present in the vicinity. If it is present, it does not affect the basin at this stage.

As climate gradually cools, organic sediment accumulates in the bottom of the basin, and vegetation invades around the margins, creating a minerotrophic wetland (B.1). There is still some open water, with vegetation such as *Nuphar* and *Chara*, and populations of aquatic invertebrates. As sediment accumulation and infilling continue, a minerotrophic peatland (fen) develops (B.2). Permafrost continues to aggrade until it eventually reaches the lower sediments of the basin (C). The peatland surface is raised by the expansion of freezing pore water and development of segregation ice, until it is at or above the water table. *Sphagnum* spp. become more dominant in the surface vegetation.

The combined effects of a more rapid cooling such as that experienced in this region at about 5000 BP, and the development of an insulating *Sphagnum* layer lead to further aggradation of permafrost, with the permafrost table becoming increasing close to the surface (D). The resulting increased ground ice content and expansion further raises the surface of the peatland until it is above the local water table, and ombrotrophic conditions are established.

After this stage, further elevation of the peat surface by the combined effects of peat accumulation, ground ice and permafrost aggradation causes the surface to become too dry for further accumulation of peat. The *Sphagnum* ground cover is replaced by lichens and heath (E).

This general sequence is found in the paleoecological records of both Kukjuk and Campbell Creek peatlands, and might be expected in other shallow basins under similar conditions, with some variation. With the exception of the influence of permafrost in later stages, the succession fits the general pattern of terrestrialization first described by Weber (1908), whereby peatlands develop autogenically through infilling as a result of hydroseral succession, with accompanying vegetation changes ultimately leading to a stable "climax" vegetation. A shallow open water system is gradually infilled with organic remains until the peat surface is raised above the influence of the local groundwater table, and the precipitation provides the only source of water and nutrients for the peatland (Heathwaite *et al.*, 1993). I propose the term "cryoterrestrialization" to describe the process illustrated in Figure 7.1, whereby an open water body is transformed into a fen and then an ombrotrophic peatland through the combined effects of infilling by organic matter and the influences of permafrost aggradation and ice buildup. During the later stages, the surface may become too dry for peat accumulation to continue. The peatland may eventually be invaded by vegetation communities typical of local upland environments, including forest where climatic conditions are suitable.

How applicable is this model of "cryoterrestrialization" to the other permafrost peatlands? The peatlands studied here can by no means be considered representative of all perennially frozen peatlands, although the successional sequences recorded in them are similar to those reported from many others in similar environments. The sequence most commonly reported from peatlands in the western Northwest Territories and east-central Yukon begins with aquatic detritus at the base, overlain by peat consisting predominantly of sedge and brown moss, with only a thin cap of *Sphagnum* peat (Tarnocai and Hughes, 1972; Zoltai and Tarnocai, 1975; Ovenden, 1982). It is believed that thin layers of *Sphagnum* peat provide enough insulation to promote local permafrost development even in areas of sporadic discontinuous permafrost (Zoltai and Tarnocai, 1975; Tarnocai and Zoltai, 1988). The same mechanism may be expected to lead to permafrost aggradation in peatlands in the continuous permafrost zone. However, in this case the cause and effect relationship is less clear. If permafrost is already present locally, it may aggrade into the accumulating peat due to climatic cooling and raise the surface of the peatland, altering its hydrology enough to allow the establishment of a *Sphagnum* cover, which in turn may promote further permafrost aggradation. Whichever comes first, the combined effects of *Sphagnum* cover and permafrost aggradation is likely to lead to changes in peatland hydrology, form and vegetation.

The resulting stratigraphic sequence and peatland characteristics will vary among basins depending on the local topography, size and shape of basin, substrate, age of the peatland, climate and other factors influencing local hydrology. This is illustrated by the differences between the Kukjuk and Campbell Creek peatlands.

The presence of permafrost in Campbell Creek peatland is not immediately apparent from its surface, except for the occurrence of occasional *Sphagnum* filled trenches, 3-15 m in length and 1-3 m in width, which are believed to represent thermal cracks in the permafrost. One such trench was cored, but no ice wedge was found. In contrast, it is immediately evident from the presence of the high-centre polygons that Kukjuk peatland is strongly influenced by permafrost and ground ice.

The differences between the two peatlands in the nature and form of ground ice might be explained by differences in local hydrology and climate. Most of the Tuktoyaktuk Peninsula, including the Kukjuk peatland area, is very poorly drained, and there is a great deal of surface water in the form of lakes, small ponds filling thermokarst depressions, and wetlands (mostly low-centre polygons). The graben in which the Campbell Creek peatland sits, on the other hand, is better drained, so there is less water available for the formation of segregation ice.

One large high-centred polygonal peat plateau complex exists approximately 1 km west of the Campbell Creek peatland, but polygon development is much less common in the Inuvik area than on the Tuktoyaktuk Peninsula. This is probably a function of climatic differences between the two areas, since ice-wedge growth is commonly associated with extreme low temperatures and limited winter snow cover (Mackay, 1993).

It is important to note that not all permafrost peatlands have evolved through terrestrialization processes. In Keewatin and northern Manitoba, paludification, which involves peat formation due to swamping of previously well drained mineral soil, is apparently more common (Ritchie, 1960; Nichols, 1972; 1975). Payette (1988) analyzed plant macrofossils from a transect of seven peat profiles from an island in Clearwater Lake, northern Quebec, which is covered by ombrotrophic peat. He concluded that a *Sphagnum* dominated peatland developed 5500-4000 BP directly on the underlying well-drained mineral soil, during a relatively humid climatic episode.

7.3 POTENTIAL EFFECTS OF CLIMATE WARMING ON PERMAFROST PEATLANDS

Various authors have discussed the possible effects of future climate warming on arctic and subarctic peatlands. Permafrost is not necessarily a permanent condition, and changes in the environment beyond certain thresholds may trigger melting, with potentially significant consequences for peatlands in which permafrost occurs. Thawing of permafrost and ground ice is likely to cause considerable thermokarst erosion of peatlands, which may cause rapid lowering of the water table as new drainage channels are created (Billings et al., 1982, 1983; Billings and Peterson, 1990). However, it will also create many shallow thaw lakes which will eventually be colonized by aquatic plants and a succession of vegetation communities until they gradually evolve, through terrestrialization, into new peatlands.

It is thought that climatic warming may renew peat accumulation in some arctic and subarctic areas (Gorham, 1988, 1990). The degradation of permafrost would lead to melting of ground ice and subsidence of presently dry peatland surfaces. Unless the basins are drained, by melting of surrounding permafrost or other mechanisms, wetter conditions would be created and groundwater may begin to affect the peatlands again, leading to the re-establishment of fen vegetation. Miller (1981) suggested that higher temperatures might effectively shift the zone of peat formation northward. Zoltai and Pettapiece (1983) note that ~ 8500-9000 BP, when climate was warmer than present, peat formation was common in some parts of the arctic where accumulation has since been negligible as a result of climatic cooling and permafrost development.

In the event of climate warming beyond the threshold levels at which widespread thermokarst occurs, peatlands such as Kukjuk, with high volumes of ground ice, are likely to be most drastically affected. The excess ice in Kukjuk makes it "thaw sensitive". If the entire thickness of ground ice were to melt, the surface of the centre of the peatland would be lowered by a metre or more. Campbell Creek, on the other hand, is likely to be less sensitive to thermokarst, since it contains much less excess ground ice. Most of the water now present as ice would be contained in the peat upon thawing, and the surface of the peatland would probably not experience any drastic change as a direct effect of thawing in the peatland.

It is difficult to speculate on what magnitude of climate or other environmental change would be required to instigate widespread thawing of permafrost in peatlands within the present continuous permafrost zone. Observations of subarctic and arctic peatlands indicate that, at least under recent climatic conditions, extensive degradation of permafrost does not occur except under extreme circumstances. Even intense disturbances such as fires do not usually trigger large scale thawing except where the peat layer is thin and is completely consumed by the fire (Zoltai *et al.*, 1988). Zoltai *et al.* note that collapse seems to be more often initiated by persistent rising of the water table due to construction of roads or pipelines, or direct damage to the surface of frozen peatlands.

Paleoecological investigations such as this one provide important background information for understanding the ecological implications of global change. Proper resource management requires long-term environmental data which ecologists rarely have the opportunity to collect, but plaeoecological data provide a valuable proxy by providing insight into ecosystem response to past environmental change (Gorham and Janssens, 1992).

8. CONCLUSIONS

Permafrost aggradation and the accumulation of ground ice had a significant influence on the development of the peatlands studied. Both peatlands seem to have undergone significant changes coincident with the deterioration of climate after 5000 BP, which are interpreted as being related to the climate changes and associated aggradation of permafrost. However, it is difficult to interpret peat stratigraphies in terms of the influence of climate changes, since many more local factors may also have significant effects on peat stratigraphy, including local hydrological changes and autogenic hydroseral succession.

The results illustrate the advantages of multiproxy research methods and the examination of multiple cores in studies of peatland ecosystem processes. In both peatlands, significant information was gained by combining pollen analysis with analysis of plant macrofossils and the physical characteristics of several cores from different parts of the peatland. A significant example is the discovery of *Picea* and *Larix* needles at the base of core KJ-B, providing strong evidence of the local presence of trees during the early Holocene, even though there was no associated change in the pollen record. Examination of multiple cores made it clear that the entire histories of the peatlands could not have been obtained from any one core, since different parts of the peatlands did not have the same stratigraphy or history of peat accumulation. Even with the use of several cores, it is impossible to fully reconstruct the record of changes throughout a system as small as Kukjuk peatland; the larger and more diverse the peatland, the greater the need for analysis of multiple cores.

The results also demonstrate how pollen and plant macrofossil analyses complement each other in revealing the history of vegetation change in and around a peatland. While it may not be feasible to conduct detailed paleoecological analysis on many cores from a single peatland, collection of several cores is recommended wherever possible. Analysis of the general stratigraphy and physical characteristics of the peat from different parts of the basin can provide details that cannot be obtained from even the most detailed analysis of a single core. Combining data from paleoecological studies with the results of other methods offers our best chance at understanding the relationship between peat accumulation processes and environmental factors, in both permafrost and non-frozen peatlands. Stable isotope data in particular have been shown to be helpful in deciphering past hydrological changes (e.g. Vardy *et al.*, 1997).

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