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Postglacial vegetation and climate

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Abstract: The past climate and vegetation of the Mackenzie valley has been reconstructed for the past 13 000 radiocarbon years using fossil-pollen records and studies of tree rings. The earliest vegetation cover in much of the valley was sparse and dominated by small shrubs and herbs. It is possible that aspen and/or poplar trees were also present. By about 10 000 BP birch tundra was dominant throughout the valley. Between 10 000 and 9000 forests of white and black spruce were established as far north as the Tuktoyaktuk Peninsula. The forest retreated slightly in the Mackenzie Delta region and in the mountains after 5000 years ago. Tree-ring records show that the during portions of the nineteenth century, summer temperatures in the valley were perhaps 1 to 2°C colder than they are today. In the twentieth century there has been some increased establishment of spruce at higher elevations in the mountains.

Résumé : On a reconstitué le climat et la végétation de la vallée du Mackenzie au cours des 13 000 dernières années (âges au radiocarbone) en utilisant les renseignements fournis par le pollen fossile et les cernes de croissance des arbres. La première végétation à pousser dans la grande partie de la vallée était éparse et surtout composée de petits arbustes et de plantes herbacées. Des trembles ou des peupliers y poussaient peut-être également. Aux environs de 10 000 ans, la végétation dominante de la vallée était celle d'une toundra à bouleau. Entre 10 000 et 9 000 ans, des forêts à épinette blanche et à épinette noire se sont implantées aussi loin vers le nord que la péninsule de Tuktoyaktuk. Après 5 000 ans, les forêts ont reculé légèrement jusque dans le delta du Mackenzie et dans les montagnes. Les cernes de croissance des arbres indiquent que durant certaines périodes du 19^{e} siècle, les températures estivales étaient peut-être plus froides de 1 à 2 °C que celles d'aujourd'hui. Au 20^{e} siècle, on a observé une certaine augmentation des populations d'épinettes à plus haute altitude dans les montagnes.

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CLIMATE AND VEGETATION

The vegetation and climate of the Mackenzie valley have changed in the past and will change in the future. For example, approximately 18 000 BP (all data given here when discussing pollen records are radiocarbon years before 1950. Such ages do not correspond directly to calendar years) the entire Mackenzie valley was covered by glacial ice. Some of the ice came from valley glaciers and ice fields in the Mackenzie Mountains and some was part of the great Laurentide Ice Sheet that was centred to the east (*see* Dyke and Prest, 1987; Duk-Rodkin and Hughes, 1991; Duk-Rodkin and Lemmen, 2000). The glacial climate was extremely cold and vegetation was absent from the valley. By 13 000 years ago, the ice had begun to melt and the landscape of the Mackenzie valley became available for colonization by plants. A warming in climate had caused the melting of the ice. This warming and other climatic variations have caused changes in the vegetation of the valley over the past 13 000 years. For instance, 6000 years ago, summers were warmer and forests grew on the Tuktoyaktuk Peninsula, where only tundra exists today. In the near future, warming of the climate due to the 'greenhouse effect' may cause significant changes in the environment and vegetation of the Mackenzie valley. By studying how plants colonized the Mackenzie Basin, and how past changes in climate have affected vegetation, we can gain important insights into how the present environments of the Mackenzie valley evolved and how future climate changes, both natural and human, may affect the valley.

The changes in vegetation and climate that have occurred in the Mackenzie valley over the postglacial period (past 10 000 years) have been reconstructed using two lines of evidence. The first line of evidence is the record that comes from fossil pollen grains produced by plants in the past and deposited by wind and water in lake sediments and peat. The second is the analysis of tree-rings from dead and living trees. Both of these approaches are outlined below and the insights they provide on the past vegetation and climate of the Mackenzie valley are presented.

FOSSIL-POLLEN ANALYSIS

Common boreal plants such as white and black spruce (*Picea glauca* and *P. mariana*), jack pine (*Pinus banksiana*), birch (*Betula*), alder (*Alnus*), willow (*Salix*), grass (Poaceae), and sedge (Cyperaceae) produce large quantities of wind-borne pollen. The pollen grains are about the same size as silt (10–150 µm in diameter) and are easily blown away from the parent plant. Much of this pollen never fulfills its reproductive function and is deposited in lakes and bogs. Although the living interior of the grains decays rapidly, the exterior walls of the pollen grains are made up of a polymer called sporopollenin that is resistant to decay in settings where oxygen is limited, such as lake bottoms and damp, fast-growing peatlands. In these settings, pollen can be preserved for millions of years (Fig. 1).

The sediments on the bottom of a lake or in a peat bog collect over time, so that the oldest sediments are at the base and the youngest are at the surface. By studying the distribution of fossil pollen grains stratigraphically from the base of the sediments to the top, it is possible to reconstruct vegetation changes from the initial formation of the lake or bog until the present day. To undertake this research, cores are taken from the lake sediments and peat. In the laboratory, subsamples of a few millilitres are taken from the cores and chemical treatments are used to separate the pollen grains from the other organic and inorganic components of the sediments and peats. As many as 50 000 pollen grains can be recovered from 1 mL of sediment from small lakes in the Mackenzie valley.

The pollen grains are identified using a microscope at a magnification of about 400 x. Usually some 300 to 1000 grains are identified per sample. For some plants it is possible to identify the precise species that produced the pollen. For example, one can often tell apart the pollen of white spruce from black spruce. In the case of other plants, such as willows, one can only tell the genus. For some plants, such as grasses, it is generally only possible to identify the pollen as coming from the grass family (Poaceae). There are many plants, such as tamarack (*Larix laricina*), that do not produce large quantities of pollen. Other plants such as aspen and balsam

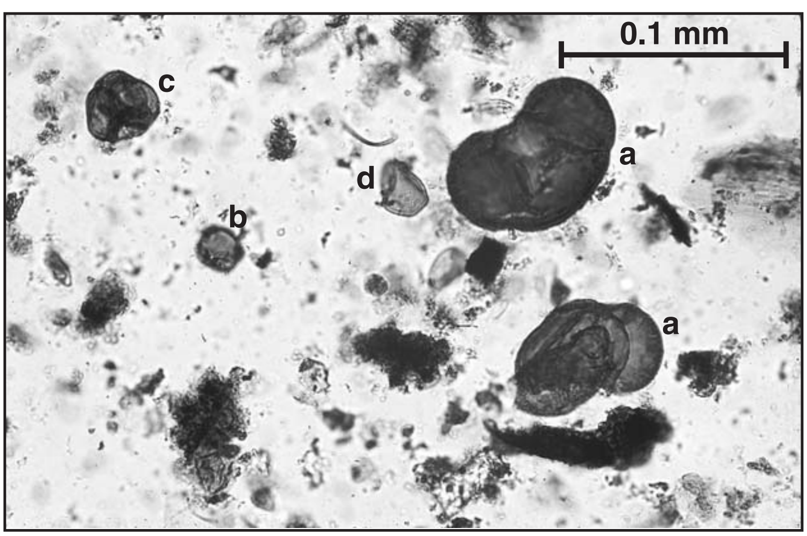

Figure 1.

A photomicrograph of fossil pollen grains from a lake near Yellowknife. The pollen includes: a, spruce, b, birch, c, heath, d, sedge. The pollen dates from about 5000 BP. Photograph by G.M. MacDonald. GSC 2000-027 (100 µm = 0.1 mm) poplar (*Populus tremuloides* and *P. balsamifera*) produce pollen that is not well preserved in sediments. It is harder to reconstruct the history of these species on the basis of pollen. Despite this, the general character of the vegetation surrounding a lake or bog can usually be reconstructed from the pollen deposited in the sediments. Small lakes (1 ha or less) tend to receive pollen locally. Larger lakes receive a greater proportion of pollen from outside the immediate area and present a more regional picture of vegetation change.

Scientists studying fossil pollen pay careful attention to the modern relationship between vegetation and climate. If aspects of modern vegetation can be related to climate, such as the northern treeline, which appears to correspond with the northernmost locations where the mean July temperature is about 10°C, then past climate conditions can be inferred from the past distribution of vegetation. A past northward shift in treeline might be apparent in fossil-pollen records and interpreted as a period when July temperatures were warmer than at present. As many lakes in the Mackenzie valley formed at the end of the last ice age, it is possible to reconstruct the entire postglacial history of vegetation and climate in the valley.

Ages are assigned to the stratigraphic records of vegetation change by radiocarbon dating organic material in the sediments. Radioactive carbon (^{14}C) is formed naturally in the stratosphere and makes up a very small proportion of the total carbon found in the atmosphere and on Earth. Plants incorporate ^{14}C during photosynthesis and it is then passed up the food chain through animals that feed on plants. All living organisms contain a small amount of radioactive carbon. When a plant or animal dies, the ^{14}C undergoes radioactive decay and becomes nitrogen (^{14}N). Half of the radioactive carbon decays to nitrogen every 5730 years. By measuring the amount of radioactive carbon left in the remains of a plant or animal, scientists can determine how long ago the organism died. By convention, such ages are reported as radiocarbon years Before Present (BP) with 'Present' taken to be AD 1950. Radiocarbon years do not exactly equal calendar years. For the last few thousand years the difference between radiocarbon and calendar years is small. However, this difference increases back in time and a radiocarbon date of 10 000 BP represents a calendar date closer to 12 000 years ago.

A latitudinal transect of fossil-pollen records from lakes and bogs is available from Mackenzie valley (Fig. 2). This transect can be used to reconstruct vegetation change since the end of the last glacial period. In the extreme north, a fossil-pollen record from Sleet Lake in the vicinity of Tuktoyaktuk presents a record of 13 000 years of vegetation change (Spear, 1983, 1993; Fig. 2a). At 13 000 BP the region was newly free of glacial ice and supported tundra vegetation dominated by shrub birch. This is evidenced in the pollen record by the large amounts of birch pollen deposited at that time and very low amounts of tree pollen. Between 10 000 and 9000 BP there was an increase in spruce pollen. The abundance of spruce pollen remained high until about 7000 BP and then declined until 4000 BP. These changes in spruce-pollen abundance suggest that between 9000 and 7000 BP the region near Tuktoyaktuk was likely forested (Spear, 1983, 1993; Ritchie, 1984, 1987). This forest thinned and retreated between 7000 and 4000 BP. The present mean July temperature (an indication of how warm the summer is) on the Tuktoyaktuk Peninsula is approximately 7–9°C. The presence of spruce forest in this region between 9000 and 7000 BP suggests that the mean summer temperatures at that time may have been approximately 1–3°C warmer than those of today. In addition, there is evidence that the treeline was slightly higher in elevation and high-elevation forests were denser in the Mackenzie Mountains during this warm period (MacDonald, 1983; Szeicz et al., 1995) Needles of spruce have been found in peat deposits that lie above the modern treeline near the headwaters of the Natla River in the Mackenzie–Selwyn mountains (Fig. 2). The modern tundra that makes up the vegetation of the Tuktoyaktuk Peninsula today was established after 5000–4000 BP (Fig. 2a). In the pollen record this tundra is evidenced by large amounts of birch (shrub birch), alder, heath, grass, and sedge pollen (Ritchie, 1984, 1987; Anderson et al., 1992). It is known that the orbital geometry of the Earth was different in the past, and summer insolation in the high northern latitudes was about 4% higher than today between 11 000 and 6000 BP. This increased insolation would have caused summers to be warmer and could explain this northward shift of treeline in the Mackenzie valley and the upward shift of the treeline in the mountains (Ritchie, 1987; Spear, 1993; Szeicz et al., 1995).

Fossil pollen from Lac Mélèze and Lac Demain located further south in the valley near Norman Wells (MacDonald, 1987; Fig. 2b) and Fort Simpson (MacDonald, 1987; Fig. 2c) present a different pattern of vegetation history. The southern and central valley was deglaciated later than the Tuktoyaktuk region and the pollen records do not begin until 11 000 to 10 000 BP. Both of these sites were first occupied by a herb- and grass-dominated vegetation. This is recorded in the pollen records by large percentages of sage (*Artemisia*) and grass pollen. There is some pollen from aspen or balsam poplar in the sediments of this age, so it is possible that these trees were present. Aspen/poplar pollen is also present in the sediments deposited in Sleet Lake at this time (MacDonald, 1987; Spear, 1993; Fig. 2a). Perhaps the climate of the central and southern Mackenzie valley was dry rather than extremely cold when this initial vegetation was established. The herb, grass, and poplar vegetation was rapidly replaced by shrub birch. The pollen records show that the birch-dominated vegetation was in turn rapidly replaced by spruce forest between 10 000 and 9000 BP. Dense spruce forest has persisted at these sites from about that time until the present (Ritchie, 1984, 1987; Ritchie and MacDonald, 1986; MacDonald, 1987; MacDonald and McLeod, 1996).

At about 7000–6000 BP there was an increase in alder pollen in all sites along the Mackenzie valley (Fig. 2a–2c). Alder is typically associated with damp acidic soils and is often found growing around lake margins and stream channels. The near-synchronous increase in alder throughout the valley suggests an increase in moisture at 7000–6000 BP. This may have been due to decreasing summer warmth, increasing precipitation, or some combination of these factors (MacDonald, 1987).

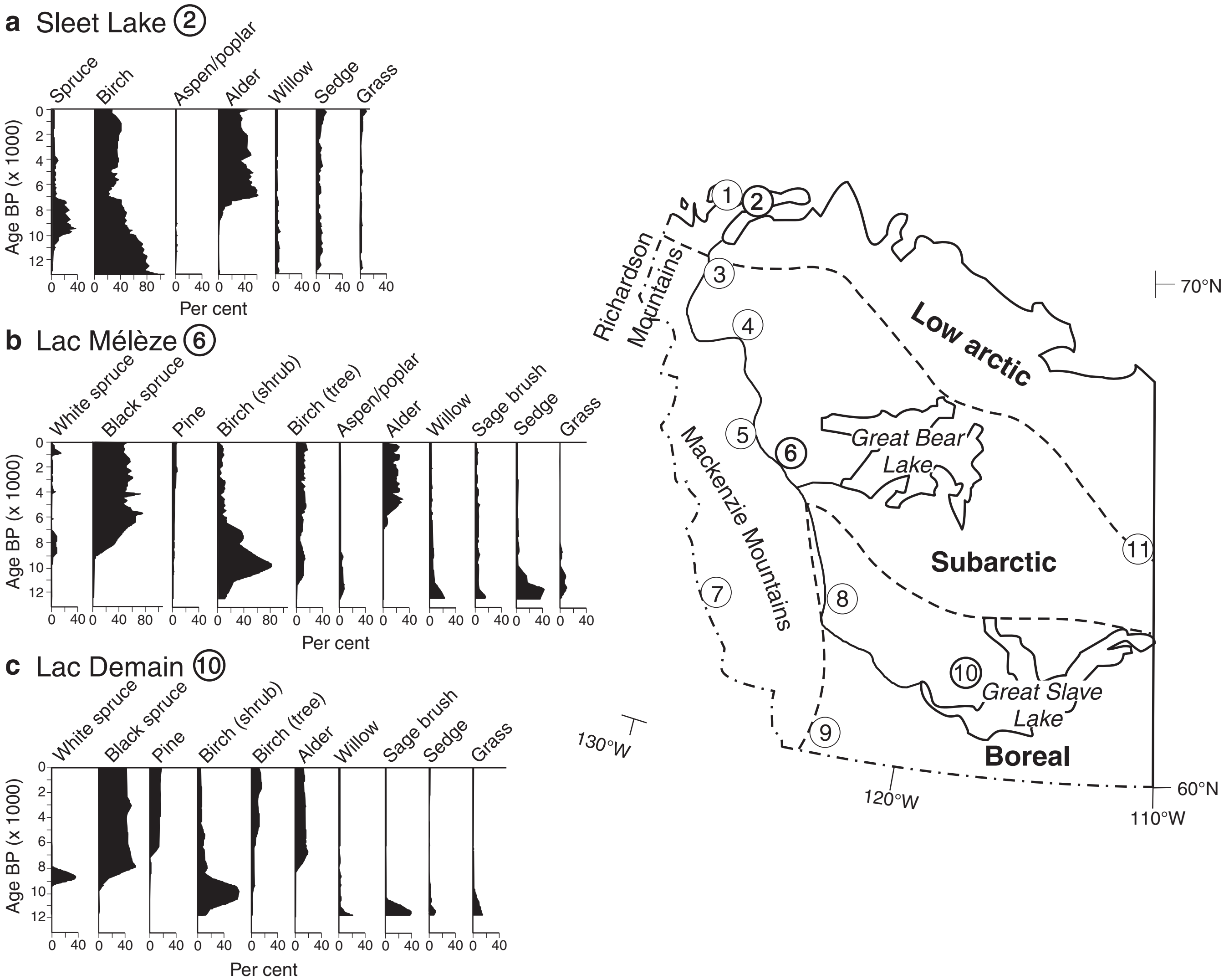

Figure 2. *The location of fossil pollen sites along the Mackenzie valley and fossil pollen diagrams from* ***a****) Sleet Lake (2),* ***b****) Lac Mélèze (6), and* ***c****) Lac Demain (10).*

Peatlands are an important part of the landscape of the Mackenzie Valley today and have been in the past (Zoltai and Tarnocai, 1975). The spores of *Sphagnum* peat moss are deposited and preserved in lakes and bogs along with pollen. The amount of these spores in sediment deposits increased to modern abundances between about 7000 and 4000 BP in the Mackenzie valley (MacDonald, 1987; MacDonald and McLeod, 1996; Fig. 3). This suggests that the extensive peatlands that typify the region today were established at that time. Peatlands are generally associated with cool and moist climates. A decrease in summer warmth and increasing moisture after 6000 BP may have aided the development of extensive peatlands in the Mackenzie valley (MacDonald, 1987).

The last important change was the arrival of jack pine in the southern and central Mackenzie valley at about 4500 BP. Both spruce and jack pine spread to the Mackenzie valley from the south (Ritchie and MacDonald, 1986; MacDonald, 1987; MacDonald and MacLeod, 1996). Spruce spread rapidly up the valley and was present from the Alberta–Northwest Territories border to Tuktoyaktuk by 9000 BP. The rapid spread of spruce may have been due to the northward transport of seed by northwesterly blowing winds coming from the remaining Laurentide Ice Sheet (Ritchie and MacDonald, 1986). In addition, seeds, and even trees and soil dislodged by floods may have been transported northward by the Mackenzie River. The reasons for the slower spread of pine compared to spruce are unclear. It may have been that the glacial refuge of jack pine was located further to the southeast than that of spruce. In addition, jack pine cannot tolerate shade and may have had a hard time establishing once spruce forests were dominant in the region. Finally, a changing climate after 6000 BP, perhaps to moister conditions, may have helped the spread of jack pine.

Taken together, the fossil-pollen evidence from sites along the Mackenzie valley suggest that at 13 000 BP the northern end of the valley supported a shrub-birch-dominated vegetation (Fig. 4a). By 10 000 BP, the southern and central portions of the valley were ice-free

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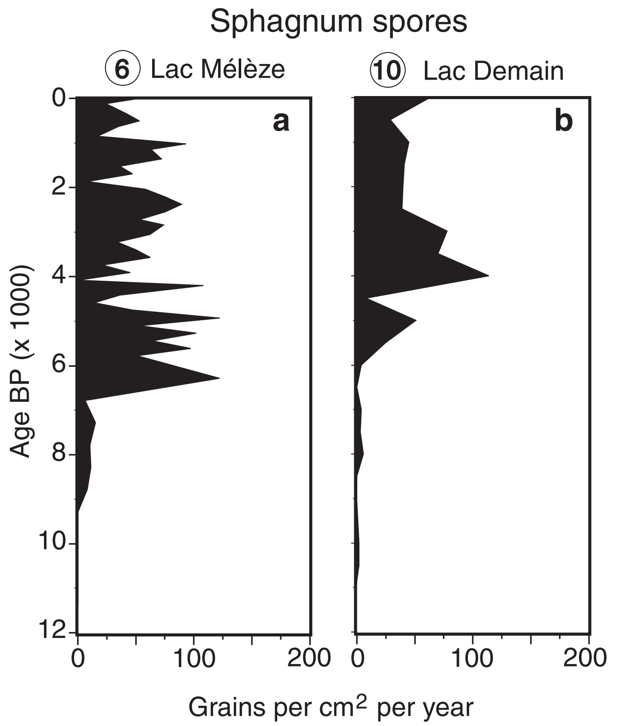

Figure 3. *The accumulation rates of peat-moss spores (Sphagnum) in lake sediments at **a**) Lac Mélèze (6), and **b**) Lac Demain (10). The large increase at around 6000 BP likely represents the establishment of widespread peatlands in the Mackenzie valley.*

and supported a herb-, shrub-, and poplar-dominated vegetation (Fig. 4b). Warm, and relatively dry summers occurred between 10 000 and about 6000 BP. The initial herb, birch, and poplar vegetation was rapidly replaced by spruce forest between 10 000 and 9000 BP. Spruce trees grew well north of their modern range between 10 000 and 5000 BP (Fig. 4c). Between 6000 and 4000 BP, the spruce treeline retreated southward, alder and peatlands became important components of the vegetation, and jack pine appeared in the valley. By about 4000 BP, the modern vegetation of the region was established. Since that time there has been little in the way of major vegetation or climatic changes. However, it is likely that smaller scale changes in vegetation have occurred and some of these will be examined below.

ANCIENT WOOD AND TREE-RING ANALYSIS

The study of ancient wood and tree rings provides another line of evidence for reconstructing past vegetation and climate in the Mackenzie valley. In the cold and relatively dry conditions of the Arctic, wood may be preserved for thousands of years following the death of a tree. Several stumps of white spruce and tamarack have been found on the tundra of the Tuktoyaktuk Peninsula (Spear, 1983; Ritchie, 1984). These stumps date to a time when forests grew on the peninsula. Radiocarbon dates from the stumps range from 7500 to 4900 BP. The presence of these stumps, and the radiocarbon ages obtained for them, confirm the evidence from the fossil-pollen records of forest development on the Tuktoyaktuk Peninsula in the past.

Trees such as spruce produce distinctive concentric growth rings during each summer. The rings are made up of couplets of light 'early wood' that forms at the start of the

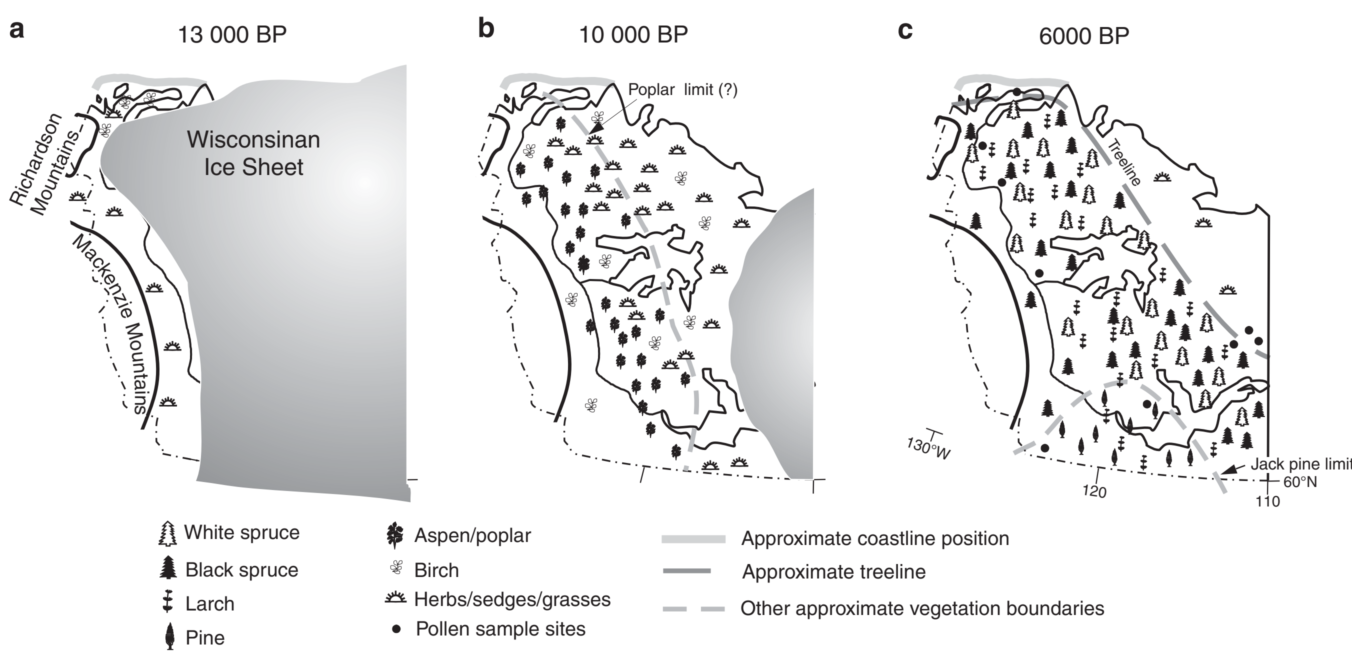

Figure 4. *Summary maps of vegetation in the Mackenzie valley at **a**) 13 000 BP, **b**) 10 000 BP, **c**) 6000 BP.*

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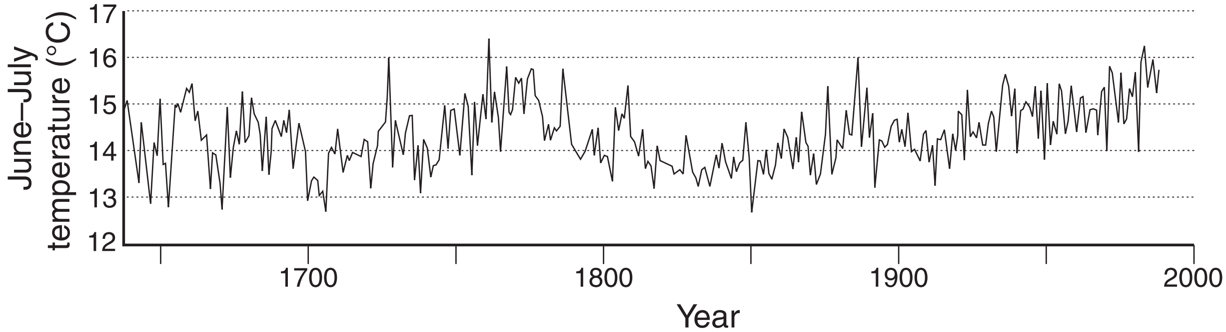

Figure 5. *Reconstruction of June–July temperatures in the Mackenzie valley region for the period AD 1638–1988.*

summer and darker 'late wood' that forms as growth slows down towards the middle of summer. The oldest rings are found at the centre (or pith) of the tree and the most recent near the bark. By counting these rings it is possible to determine the age of the tree. In general, trees growing under good conditions produce wide rings. For spruce trees near treeline in the northern Mackenzie, or in the mountains, the rings produced during years with warm summers are generally wide, while those produced during cool years are narrow (D'Arrigo and Jacoby, 1993; Szeicz and MacDonald, 1995a). Thus, the width of the rings can be used to gauge the general conditions under which the tree grew, and also the specific changes in weather from year to year.

The ring widths and growth rates of the stumps from the Tuktoyaktuk Peninsula have been measured and compared with the growth rates of living trees. The rate of growth of these ancient trees was found to be similar to that of the modern spruce trees growing near Inuvik. This suggests that summer climate near Tuktoyaktuk 6000 years ago was similar to the present climate at Inuvik, which is further south (Ritchie, 1984). This would represent a warming of July temperatures at Tuktoyaktuk of approximately 3°C. This estimate of summer warming corresponds well with estimates based on the fossil-pollen record as outlined above.

Tree-ring data collected from living white spruce trees, and spruce trees that have died recently, have been used to produce detailed records of summer temperature variations over the past 300 to 500 years in northwestern Canada and Alaska (D'Arrigo and Jacoby, 1993). In these studies, tree-ring records from a number of trees at several sites are mathematically combined to produce estimates of summer and annual temperatures. These reconstructions indicate that temperatures were low during the last century and have been generally increasing since that time. Detailed analysis of trees growing in the Mackenzie and Franklin mountains have recently been used to reconstruct July mean temperatures (Szeicz and MacDonald, 1995a). This reconstruction provides similar evidence of cold conditions during the period AD 1800 to 1880 with a general warming since that time (Fig. 5). During the very coldest period, at around AD 1850, temperatures in the Mackenzie valley were likely 2–3°C cooler than at present.

The cold conditions of the last century have been recognized at many sites in northern North America and Europe. The time period from AD 1600 to 1880 was generally cool and is often referred to as the Little Ice Age. Some have attributed the warming trend following AD 1880 with the greenhouse effect, caused by increasing level of anthropogenic carbon dioxide in the atmosphere that started with the Industrial Revolution. The warming since 1880 may, however, be a natural variation in climate (*see* Dyke and Brooks, 2000).

The impact on vegetation of these cold temperatures of the last century and the more recent warming has been studied at the treeline in the Mackenzie and Franklin mountains (Szeicz and MacDonald, 1995b). It has been found that many trees at the treeline died during the middle of the last century. Many of the trees which grow at these high-elevation sites today did not begin to grow until the start of warming in the 1880s. Thus, the treeline vegetation that we see today is actually a product of changing climate over the past 200 years.

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