Climate change and impacts in southern British Columbia

- a palaeoenvironmental perspective

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Abstract

A comprehensive review of palaeoenvironmental data has been prepared, providing the basis for evaluating natural variability in climate and ecosystem dynamics in southern British Columbia. Key indicators of past changes include palaeolimnological, palaeobotanical, glaciological, and dendroclimatological evidence. Long records of change pre-dating the last glacial maximum in British Columbia are scarce; thus, data derived from adjacent unglaciated parts of Washington state provide essential background for understanding British Columbia's glacial history.

Abundant evidence preserved in the sediments of British Columbia lakes records the rapid transition from a glacial to an interglacial climate over the interval 12,500 to 9000 ^{14}C yr BP. Peak summer temperatures (about 3°C warmer than present) and minimum precipitation were recorded for southern British Columbia ca. 9000 to 7000 ^{14}C yr BP, but were likely accompanied by winter temperatures colder than today's. A strong Pacific high prevailed during the summer months.

Summer temperatures gradually declined over the interval 7000 to 3000 ^{14}C yr BP, as wetter conditions and a stronger Aleutian low developed. Many glacial advances have been recorded in the past 3500 years, with most glacial maxima dating to the mid 19th century. A general retreat of these glaciers has accompanied the recent warming trend in British Columbia.

Palaeoenvironmental records, such as these, are critical for the development and testing of climate models, and for an understanding of the long-term dynamics of ecosystems. The perspective offered by these long-term changes in ecosystems is needed for the appropriate management of forest resources and development of protected area strategies.

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Introduction

Many palaeoecological and palaeoclimatological studies have been conducted in southern British Columbia. These studies provide essential background information on the natural variability of climate and ecosystem dynamics, prior to the period of rapid industrialisation of North America and Europe. Although climate had been relatively stable in southern British Columbia in recent millennia (last 3000 years), rapid climatic changes are occurring now, and have also occurred in the past.

Palaeoecological studies reveal how natural systems have responded to these climatic changes, providing a unique perspective on how today's ecosystems are likely to respond to future changes. In this paper we focus on the palaeoenvironmental record 1) to provide background information on the magnitude and nature of past climatic changes in the region and their impact on natural systems, and 2) to draw on existing palaeoecological data to project how natural systems are likely to respond to future changes in the climate system.

History of Quaternary Palaeoecological and Palaeoclimatological Work in the Region

Pollen & Plant Macrofossil Analyses

Quaternary palaeoecological research in southern British Columbia was pioneered by Hansen (Hansen, 1940, 1947a, b, 1950a, b; 1955, 1967), who conducted pollen analytical studies at many sites distributed throughout northwestern North America, including southern coastal British Columbia, and the dry, central and southern interior of the province. His extensive sampling provides a basic framework for later palaeobotanical studies, but without the benefit of radiocarbon dating of his sediments, his chronologies, and hence interpretation of his results, are equivocal. Subsequent studies by Heusser and Terasmae provide additional data, principally from maritime coastal regions, including sites in the adjacent states of Washington and Alaska, as well as British Columbia (Heusser, 1952, 1953, 1955, 1960, 1964, 1973, 1974, 1977, 1983a, 1985; Heusser & Heusser, 1981; Heusser *et al.*, 1980, 1985, 1999; Terasmae & Fyles, 1959).

In recent years, detailed studies of past vegetation dynamics in British Columbia have been led by Richard Hebda and Rolf Mathewes, two former students of Glen Rouse at University of British Columbia. They later secured positions at the Royal British Columbia Museum and Simon Fraser University, respectively. Hebda, Mathewes, and their students are responsible for the great bulk of palaeobotanical data now available for the province (e.g., Allen *et al.*, 1999; Brown & Hebda, 1998; Gottesfeld *et al.*, 1991; Hebda, 1983a, 1997b, 1998; Hebda & Allen, 1993; Hebda & Brown, 1999; Hebda & Whitlock, 1997; Heinrichs *et al.*, 1999a; Mathewes, 1980, 1989, 1991a, b; Mathewes & Clague, 1982; Mathewes & Jarzen, 1987; Mathewes & Rouse, 1975; Pellatt & Mathewes, 1994, 1997; Pellatt *et al.*, 1997, 1998; Quickfall, 1987; Wainman & Mathewes, 1990; Warner *et al.*, 1984; White & Mathewes, 1982, 1986; White *et al.*, 1979, 1985; Williams & Hebda, 1991). Research in Richard Hebda's laboratory has focused on past vegetation dynamics in two regions, Vancouver Island and the dry southern interior of British Columbia, whereas Rolf Mathewes' interests have encompassed principally forest development in coastal mainland British Columbia and Haida Gwaii (the Queen Charlotte Islands).

For the dry interior, important additional contributions have been provided by Anderson (1973), Alley (1976), Cawker (1983) and Evans (1997). Farther east, in the Kootenay region of British Columbia, and the Rocky Mountains of adjacent southern Alberta, results of palaeobotanical investigations have been reported by many different authors (e.g., Beaudoin, 1986; Beaudoin & Reasoner, 1992; Cwynar & MacDonald, 1987; Fahnestock & Agee, 1983; Fulton *et al.*, 1989; Hazell, 1979; Johnson & Larsen, 1991; Kearney & Luckman, 1983b; Luckman, 1986; MacDonald, 1982, 1987; MacDonald & Cwynar, 1985; Mott & Jackson, 1982; Reasoner, 1988; Reasoner & Huber, 1999; Vance *et al.*, 1995).

Palaeoecological studies from sites in the adjacent states of Washington, Idaho and Montana are also important to our understanding of the climatic history and vegetation dynamics of southern British Columbia (e.g., Barnosky, 1981, 1984, 1985a, b, 1989; Cwynar, 1987; Davis, 1973; Dunwiddie, 1986, 1987; Gavin & Brubaker, 1999; Hansen & Easterbrook, 1974; Leopold *et al.*, 1982; Mack *et al.*, 1976, 1978a, b, c, d, 1979; Mehringer, 1985, 1996; Mehringer *et al.*, 1977; Tsukada *et al.*, 1981; Whitlock, 1992; Whitlock & Bartlein, 1997)

Palaeolimnology

Until recently, very little palaeolimnological work had been conducted in British Columbia. The earliest studies, conducted by Mathewes & D'Auria (1982) and Stockner (e.g., Ennis *et al.*, 1983; Stockner & Costella, 1980; Stockner & Northcote, 1974), focused entirely on recent lake sediments, and especially human impacts on aquatic communities. Longer lake records were later used by Mathewes & King (1989) to reconstruct water level fluctuations and their climatic causes in the Lillooet region, based principally on mollusc fossils and sediment lithology, in addition to palaeobotanical evidence.

In 1983, Ian Walker, as a student of Rolf Mathewes, began studying midge fossils preserved in the sediments of British Columbia lakes, and concluded that they too provided good evidence of past climatic conditions (Walker, 1988; Walker & Mathewes, 1987a, b, 1988, 1989a, b, c). At this time, the idea of using palaeolimnological data to infer past climatic changes was considered controversial (Hann & Warner, 1987; Hann *et al.*, 1992; Warner, 1984; Warner & Hann, 1987; Warwick, 1989). In more recent years, palaeolimnological indicators, and midge remains in particular, have gained general acceptance as one of many possible tools useful in the reconstruction of past climatic changes (Battarbee, 2000; Smol, 1990, 1992; Smol *et al.*, 1991; Walker, 1993, 1995, in press). Students have since expanded upon Walker's initial research (Heinrichs, 1995; Heinrichs *et al.*, 1997, 1999, in press; Palmer *et al.*, 1998, submitted; Rück *et al.*, 1998; Smith *et al.*, 1998), and the development of quantitative midge-temperature inference models (Palmer, 1998), provides independent means of assessing earlier, palaeobotanical inferences with respect to climatic change.

In recent years the potential of saline lake sediments as archives for palaeoclimatic data has also attracted considerable interest, not only within British Columbia, but western Canada generally (Bennett, 1996; Bennett *et al.*, 2001; Chaudhari, 1998; Cumming & Smol, 1993; Cumming *et al.*, 1993; de Grace, 1999; Elder, 1997; Heinrichs, 1995; Heinrichs *et al.*, 1997, 1999b, in press; Laird *et al.*, 1996; Lowe *et al.*, 1997; Salomon, 1996; von Westarp, 1997; Wilson *et al.*, 1994, 1996; Zeeb & Smol, 1995). In British

Columbia, saline lakes are abundant in the dry, southern interior, and many of these lakes were sampled as a part of the PISCES ("Palaeolimnological Investigations of Salinity, Climate and Environmental Shifts") project. Quantitative salinity inference models were key deliverables promised in the project proposal; thus, quantitative means are now available for inferring past lakewater salinity from diatom (Cumming & Smol, 1993; Wilson *et al.*, 1994, 1996), chrysophyte (Cumming *et al.*, 1993; Zeeb & Smol, 1995), crustacean (Bos *et al.*, 1999), sponge (Cumming *et al.*, 1993) and midge (Heinrichs *et al.*, in press; Walker *et al.*, 1995) remains. Application of these models has recently begun. The resulting palaeosalinity records are likely to provide key evidence of past precipitation in much of southern interior British Columbia (Bennett, 1996; Bennett *et al.*, 2001; Chaudhari, 1998; de Grace, 1999; Elder, 1997; Heinrichs, 1995; Heinrichs *et al.*, 1997, 1999b, in press; Salomon, 1996; von Westarp, 1997).

Dendroclimatology

Most dendroclimatological research in British Columbia and the adjacent Rocky Mountains has been conducted in Brian Luckman's laboratory at the University of Western Ontario (e.g., Colenutt & Luckman, 1991; Luckman, 1993, 1997; Luckman & Kavanagh, 1998; Markgraf *et al.*, 2000; McCarthy & Luckman, 1993; McCarthy *et al.*, 1991), or, more recently, at the University of Victoria Tree-Ring Laboratory (e.g., Lewis & Smith, 1999; Smith & Laroque, 1998a). Their research has culmiinated in several long proxy records of past temperature and precipitation (e.g., Watson, 1998) in the region.

Glacial History

The glacial history of British Columbia is largely revealed in the published works of a small number of Quaternary geologists, including such prominent scientists as Bill Mathews (e.g., Mathews, 1991), Robert Fulton (Fulton, 1965, 1969, 1984, 1991; Fulton & Warner, 1990; Fulton *et al.*, 1986), and June Ryder (Ryder, 1987; Ryder & Thomson, 1986; Ryder *et al.*, 1991). John Clague, now occupying a distinguished research chair in the Department of Earth Sciences at Simon Fraser University, but employed previously as a Quaternary scientist with the Geological Survey of Canada, has emerged in recent years, as a leader in this field, particularly for the coastal belt of the province (e.g., Clague, 1981; Clague & Luternauer, 1982; Clague *et al.*, 1982a, b, 1983, 1987, 1989, 1991; Mathewes & Clague, 1994; Warner *et al.*, 1982).

The initial surveys of Heusser (1956), and recent, more detailed investigations especially of Luckman and Osborn (e.g., Luckman, 1994, 1995, 1998a; Luckman *et al.*, 1978, 1986, 1993, 1999; Luckman *et al*; Luckman *et al*; Osborn, 1986; Osborn & Luckman, 1988) provide parallel data for glacier fluctuations in the Canadian Rockies.

Other Indicators

Many palaeoenvironmental indicators have been little exploited for the development of climate proxy records in British Columbia. Several papers pertaining to palaeoceanography and stable isotope analyses have been published (Clague *et al.*, 1992; de Vernal & Pedersen, 1997; Edwards & Luckman, 1996; Marret *et al.*, 2001; Sabin & Pisias, 1996), and borehole temperature data and reconstructions are available for 12 sites in southern British Columbia (Huang & Pollack, 1998; Huang *et al.*, 1999). However, we could find no locally relevant compilations pertaining to fossil beetles, ice cores, speleothems or pack-rat middens. These are obvious areas for future research.

Interpretation

In this paper, we place considerable emphasis on the palaeolimnological record, and its implications for our understanding of past climatic changes in British Columbia. There are two key arguments for this approach:

1) little palaeolimnological data was available when earlier reviews (Hebda, 1983b, 1995; Mathewes, 1985) were being written; thus these data offer a new approach, and are especially likely to yield new insights.

2) since aquatic organisms have much shorter lives than trees (the source of most terrestrial pollen and plant macrofossils), many palaeoecologists have argued that aquatic species likely responded much more rapidly to past climatic changes (e.g., Smol *et al.*, 1991; Walker *et al.*, 1991b)

In recent years, aquatic midges (particularly the Chironomidae) have emerged as a leading new indicator of past summer temperatures. Battarbee (2000: 107), for example, has concluded that chironomid analysis is the "most promising biological method for reconstructing past temperature". Thus, we will begin our discussion with an interpretation of their record.

Chironomids & Summer Palaeotemperature

The "non-biting midges" or Chironomidae comprise a family of true flies (Class Insecta, Order Diptera) that are aquatic in their larval and pupal stages, but the adults are equipped with a pair of wings, allowing them to readily disperse to new aquatic habitats. During their larval stages these insects live principally in bottom habitats of lakes and streams. As they grow, the larvae shed four head capsules. These head capsules preserve in lake sediments, are readily identifiable, and changes in the species composition of their communities are frequently used to infer past changes in temperature, water salinity, and nutrient loading in lakes (Walker, 1987, 1995, in press).

Walker & Mathewes pioneered the use of chironomids as palaeotemperature indicators (Walker, 1987, 1988, 1991; Walker & Mathewes, 1987a, b, 1988, 1989a, b, c). Their papers provide low resolution, qualitative records of past temperature changes in coastal British Columbia – clearly documenting the rapidly increasing temperatures following deglaciation, beginning ca. 12,500 ^{14}C yr BP, and continuing into the early Holocene. Evidence is also presented for decreasing temperatures in the late-Holocene.

In recent years, chironomid analyses in British Columbia have become increasingly refined. Recent studies have advanced significantly, by greatly increasing the temporal resolution of the analyses, by focusing on sensitive ecotonal sites near present tree-line, and via the development of quantitative midge-palaeotemperature inference models (Palmer, 1998; Pellatt *et al.*, 2000; Smith *et al.*, 1998). The resulting high resolution, quantitative temperature reconstructions offer perhaps our best indication of how British Columbia summer temperatures have changed over the postglacial.

At this time, quantitative summer palaeotemperature records are available for four sites: Cabin Lake, 3M Pond, Crater Lake and Lake of the Woods. These sites are all located close to treeline in the Engelmann Spruce Subalpine Fir zone of south-central British Columbia. Averaging Palmer's (1998) reconstructions for all four sites, yields a "consensus" record (Fig. 1) of summer temperature spanning the last 12,000 years. As would be expected, cold temperatures (up to 4°C cooler than present) prevailed at these sites during the late-glacial.

Summer temperature rose rapidly ca. 10,000 ^{14}C yr BP at the beginning of the Holocene epoch. From 9000 to ca. 5000 ^{14}C yr BP, reconstructed summer temperature averaged 2 to 4°C warmer than current normals, confirming Mathewes' (1985) and Hebda's (1995) concept of an early Holocene interval of maximum warmth, the so-called "xerothermic". The summer temperature record for the interval 7700 to 4400 ^{14}C yr BP varies markedly among the four cores, with the Cabin Lake record suggesting a rapid cooling at about 7500 ^{14}C yr BP., whereas the 3M Pond record indicates that summer temperature remained 3 to 6 °C higher than current normals, until about 4000 ^{14}C yr BP.

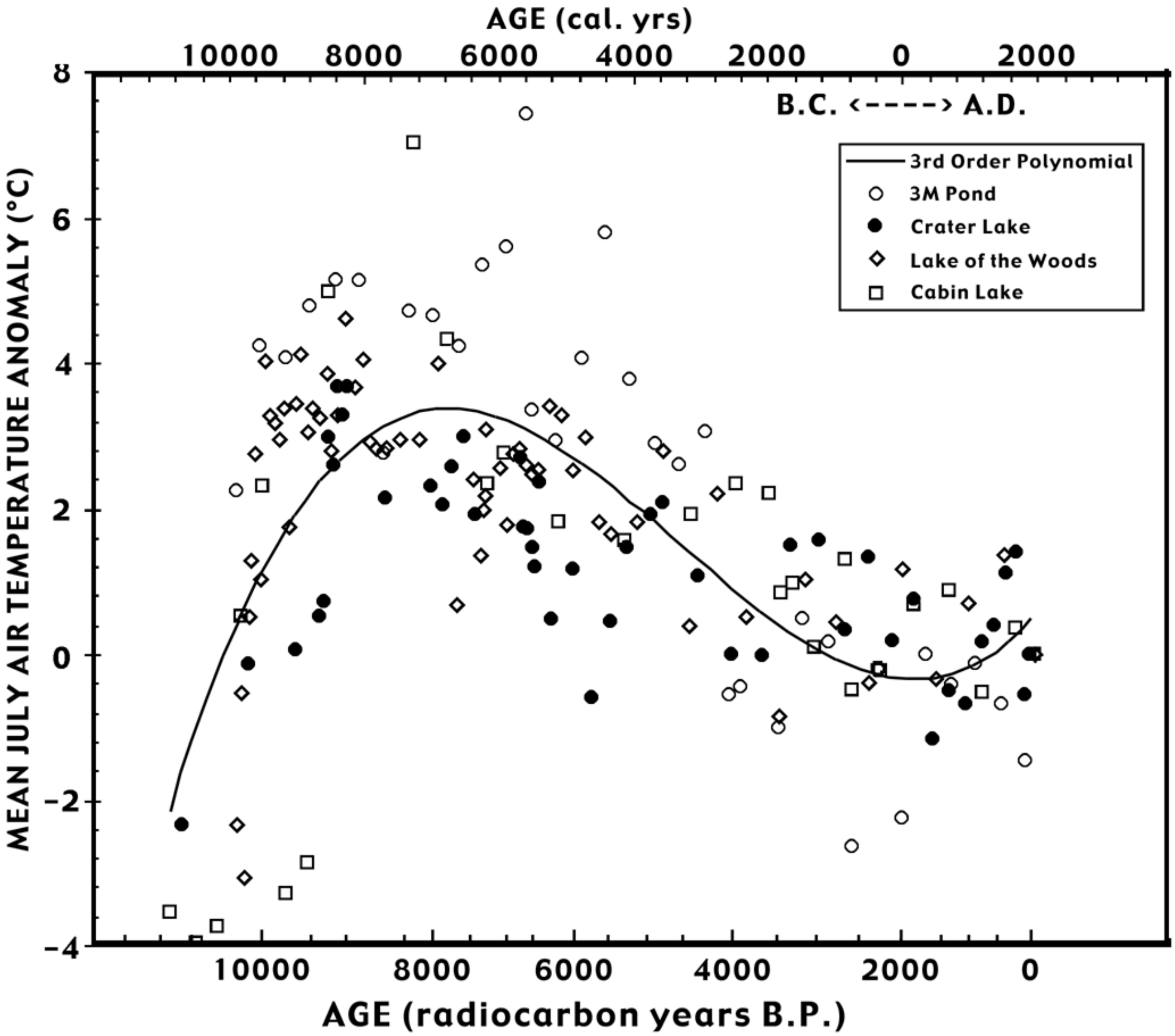


Figure 1. Chironomid-inferred palaeotemperature record for southern British Columbia (based on Palmer, 1998)

It is unclear why the mid-Holocene cooling appears time-transgressive among sites. It is perhaps best to adopt the average or “consensus” reconstruction (Fig, 1) as most probable. This indicates a gradual mid-Holocene cooling. It portrays a compromise between Mathewes and Heusser’s (1981) reconstruction indicating an early abrupt end to xerothermic warmth, and Hebda’s (1995) suggestion of a mid-Holocene “mesothermic” interval that was just as warm as (but wetter than) the xerothermic.

The reconstructions indicate that the coolest Holocene summer temperatures occurred since 3500 ^{14}C yr BP, in an interval commonly referred to in British Columbia as the “neoglacial”.

Although the midges are now generally accepted to be good indicators of summer temperatures (e.g., Walker, in press), they likely offer little or no information pertinent to winter conditions. These aquatic insects are protected from severe winter conditions, finding refuge in their larval stages beneath winter ice cover in lake and stream habitats.

Pollen and Plant Macrofossils

Many palaeobotanical studies have been conducted in British Columbia; thus, they have long provided the principal evidence used to reconstruct past climatic change in the region (see for example reviews by Mathewes (1985) and Hebda (1995)). The distributions of plant species are clearly climatically regulated, with temperature (summer and winter) and precipitation both being important environmental controls. A significant challenge in interpreting this record is separating the independent seasonal effects of temperature and precipitation, and also the influence of plant migrations into new areas.

The northern limits of plant species distributions are often related to low summer temperatures (Barnosky *et al.*, 1987). At high elevations in the Coast Mountains, the considerable snowpack also persists long into summer and greatly restricts the growing season (Brooke *et al.*, 1970). The southern and lower limits of plant species, especially in dry interior valleys, are often related to available summer soil moisture. Although soil moisture is in part controlled by precipitation, summer temperature and diverse edaphic (site-specific) features (e.g., slope, aspect, texture and other soil properties) are perhaps equally important.

Tree seedlings are especially sensitive to climatic conditions as compared to established trees. Since soils develop slowly, and since tree species in British Columbia are especially long-lived, long lags might exist between climatic changes and vegetation response. With these considerations in mind, we now consider the plant fossil record.

The number of palaeobotanical records from British Columbia greatly exceeds the number of midge records. British Columbia was almost completely glaciated during the most recent, “Fraser” glaciation, but a few scattered interglacial deposits provide a glimpse of earlier vegetation (Alley, 1979; Alley *et al.*, 1986; Clague *et al.*, 1990; Fulton & Warner, 1990; Mathewes, 1979). In addition, several long records from adjacent Washington state provide good palaeobotanical evidence of past changes in vegetation and climate.

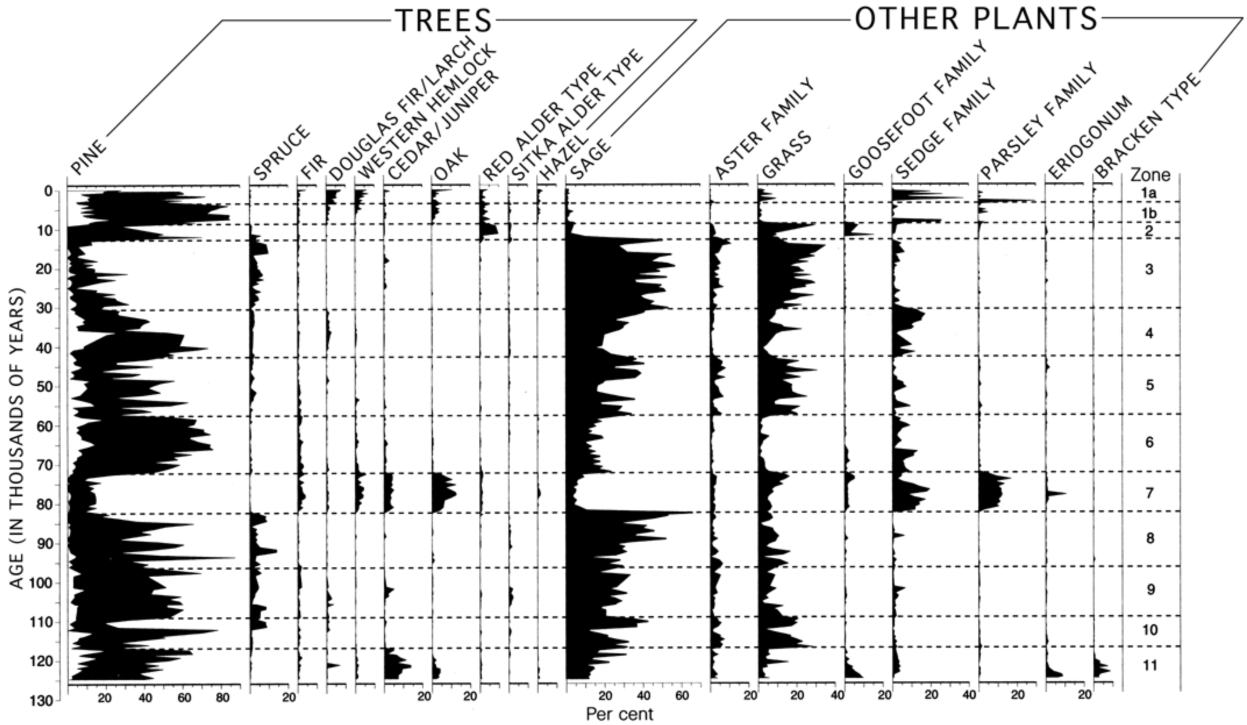


Figure 2. A 125,000 year pollen record from the forest/steppe border at Carp Lake in the eastern Cascade Mountains, Washington (adapted from Whitlock & Bartlein, 1997).

Whitlock and Bartlein (1997) for example provide a 125,000 year record from the forest/steppe border at Carp Lake in the eastern Cascade Mountains of southernmost Washington (Fig. 2). Thermophilous pollen (e.g., oak (*Quercus*), western hemlock (*Tsuga heterophylla*), cedar (Cupressaceae)) are especially well represented in Carp Lake sediments dating 133,000 to 116,700 ^{14}C yr BP, and 85,100 to 73,900 ^{14}C yr BP, indicating temperatures warmer than those recorded for the past 13,000 years (the present interglacial). Pollen from sage (*Artemisia*), grasses (Poaceae), and spruce (*Picea*) prevail from 116,700 to 85,100, and from 73,900 to 13,200 ^{14}C yr BP, indicating that glacial climates were not only colder, but also generally drier than present. The climatic inferences closely match the marine oxygen isotope record (a proxy record indicating glacial ice sheet volumes). Each warm interval was preceded by a maximum in July solar insolation.

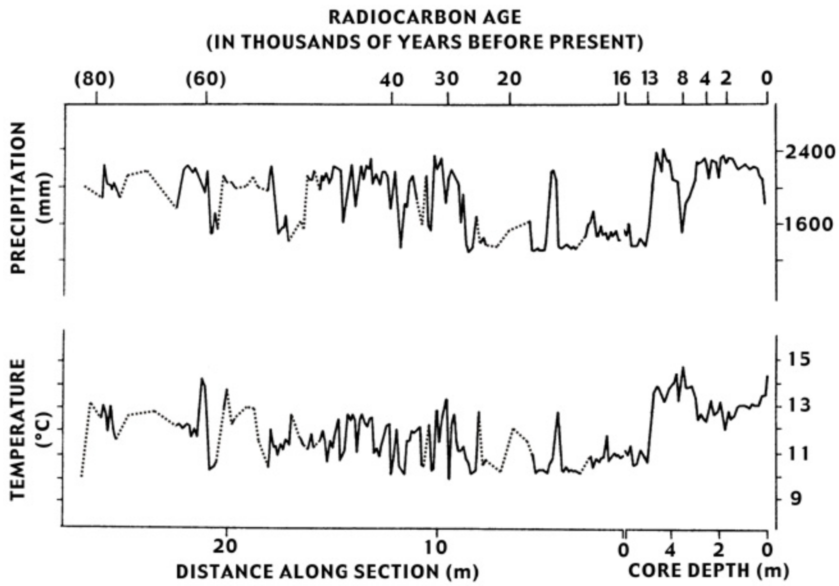


Figure 3. An 80,000 year record of temperature and precipitation from the western margin of Washington's Olympic Peninsula (adapted from Heusser *et al.*, 1980).

Heusser et al. (1980) have used transfer functions to reconstruct an 80,000 year record of temperature and precipitation from the western margin of Washington's Olympic Peninsula (Fig. 3). The record indicates that the past 13,000 ^{14}C years have been warmer than any other time in the last 60,000 years. A prolonged dry phase is evident from about 25,000 to 13,000 ^{14}C yr BP, extending through the last (Fraser) glaciation.

Although the climatic trends in British Columbia may not have always paralleled those in Washington, almost all of British Columbia was glaciated, and thus no comparably long records are available in the province. The longest continuous cores from southern British Columbia, span the last 13,000 ^{14}C years. Although these cores do not record climatic conditions spanning the last glaciation, or earlier glacial/interglacial cycles, they do provide a long late-glacial record, and a complete Holocene sequence.

Abrupt oscillations between glacial and inter-glacial conditions are a dominant feature of the late-glacial (13,000 to 10,000 ^{14}C yr BP) as recorded in eastern North America and Europe (Berger, 1990; Broecker, 1987; Broecker *et al.*, 1985; Dansgaard *et al.*, 1989; Levesque *et al.*, 1993a, b; Lowe *et al.*, 1994; Mayle & Cwynar, 1995; Mott *et al.*, 1986; Peteet *et al.*, 1990; Walker *et al.*, 1991a; Wright, 1989). These oscillations may have been caused by catastrophic episodes of freshwater meltwater influx. The meltwater may have interrupted North Atlantic Deep Water formation and thus, the "conveyor" drawing warm water northward to polar latitudes from the tropical Atlantic ocean (Broecker, 1987; Broecker & Denton, 1990a, b; Broecker *et al.*, 1985; Rooth, 1990). Although the Allerød, Amphi-Atlantic (Killarney and Gerzensee), and Younger Dryas oscillations are clearly evident at sites bordering the North Atlantic (Europe, Greenland and Atlantic Canada), the geographic extent and climatic expression of these oscillations is uncertain elsewhere (Kudrass *et al.*, 1991; Markgraf, 1991; Peteet, 1995, 1987; Peteet *et al.*, 1990). Some evidence, including dated moraines in the Rocky Mountains (Osborn *et al.*, 1995; Reasoner & Jodry, 2000; Reasoner *et al.*, 1994), and pollen and benthic foraminiferal records from coastal British Columbia and Alaska (Engstrom *et al.*, 1990; Mathewes, 1993; Mathewes *et al.*, 1993), suggests that these oscillations may have been expressed in southern British Columbia. As of yet, the evidence is derived from widely scattered sites, and at low temporal resolution. These records are not as convincing as the North Atlantic evidence, but a more detailed assessment is clearly needed.

Mathewes' (1973) pollen record for Marion Lake is typical for many sites in southwestern British Columbia. Tundra or tundra-like successional vegetation existed briefly on the newly deglaciated landscape, and was quickly replaced by coniferous forests. Pine pollen (most likely derived from lodgepole pine - *Pinus contorta*) dominates the late-glacial record (12,500 to 10,000 ^{14}C yr BP). However, since pine produces much more pollen than other species, pine may not have been the dominant tree in surrounding forests. These initial forests were likely composed of a mix of lodgepole pine, true fir (*Abies*), and spruce (*Picea*). Lodgepole pine needles and cones are common in late-glacial macrofossil collections (e.g., Clague *et al.*, 1997; Wainman & Mathewes, 1987), confirming that pine was at least a common tree locally; thus, the abundant pine pollen is not an artifact of long-distance transport, and "over-production" of pine pollen.

The boreal character of this forest clearly indicates a cool or cold late-glacial climate. It is more difficult to infer precipitation trends. Although one variety of lodgepole pine (*Pinus contorta* var. *latifolia*) is fire-dependent and typically associated with dry, well-drained soils, another variety (*Pinus contorta* var. *contorta*) prevails in coastal British Columbia today, and is commonly associated with coastal bog vegetation. Barnosky et al. (1987) note the earlier occurrence of xerophytic subalpine taxa (e.g., Engelmann spruce) in the Puget Trough as evidence for drier conditions close to the glacial maximum. The increasing abundance of mountain hemlock (*Tsuga mertensiana*) pollen suggests a gradual shift from cool dry to cool wet conditions through the late-glacial.

Douglas fir (*Pseudotsuga menziesii*) pollen, alder (*Alnus*) pollen, and bracken fern (*Pteridium aquilinum*) spores are abundant in early Holocene sediments, especially during the interval 10,000 to 6700 ^{14}C yr BP. Also noteworthy is the very low abundance, or absence, of western hemlock (*Tsuga heterophylla*), western red cedar (*Thuja plicata*), true fir (*Abies*) and spruce (*Picea*) in the earliest Holocene forests. This pattern has been widely interpreted as indicating a warm, dry period with frequent fires, generally referred to as the xerothermic in British Columbia (Mathewes, 1985).

The progressive increase in western hemlock, beginning by about 8000 ^{14}C yr BP, followed by true fir (*Abies*) and birch (*Betula*) ca. 7200 ^{14}C yr BP, western red cedar (*Thuja plicata*) 6700 ^{14}C yr BP, and spruce (*Picea*) ca. 6000 ^{14}C yr BP, suggests a gradual shift from a very warm dry state in the earliest Holocene to a cooler, more moist state in the mid to late Holocene. The late arrival of western red cedar at the site has been linked to the slow, northward migration of the species (Hebda & Mathewes, 1984); thus, palaeoecologists must interpret its mid-Holocene increase very cautiously in climatic terms.

The abrupt increases in alder, and corresponding decreases in western hemlock, douglas fir and cedar in the uppermost sediments clearly result from recent European settlement and logging activity, not climatic change.

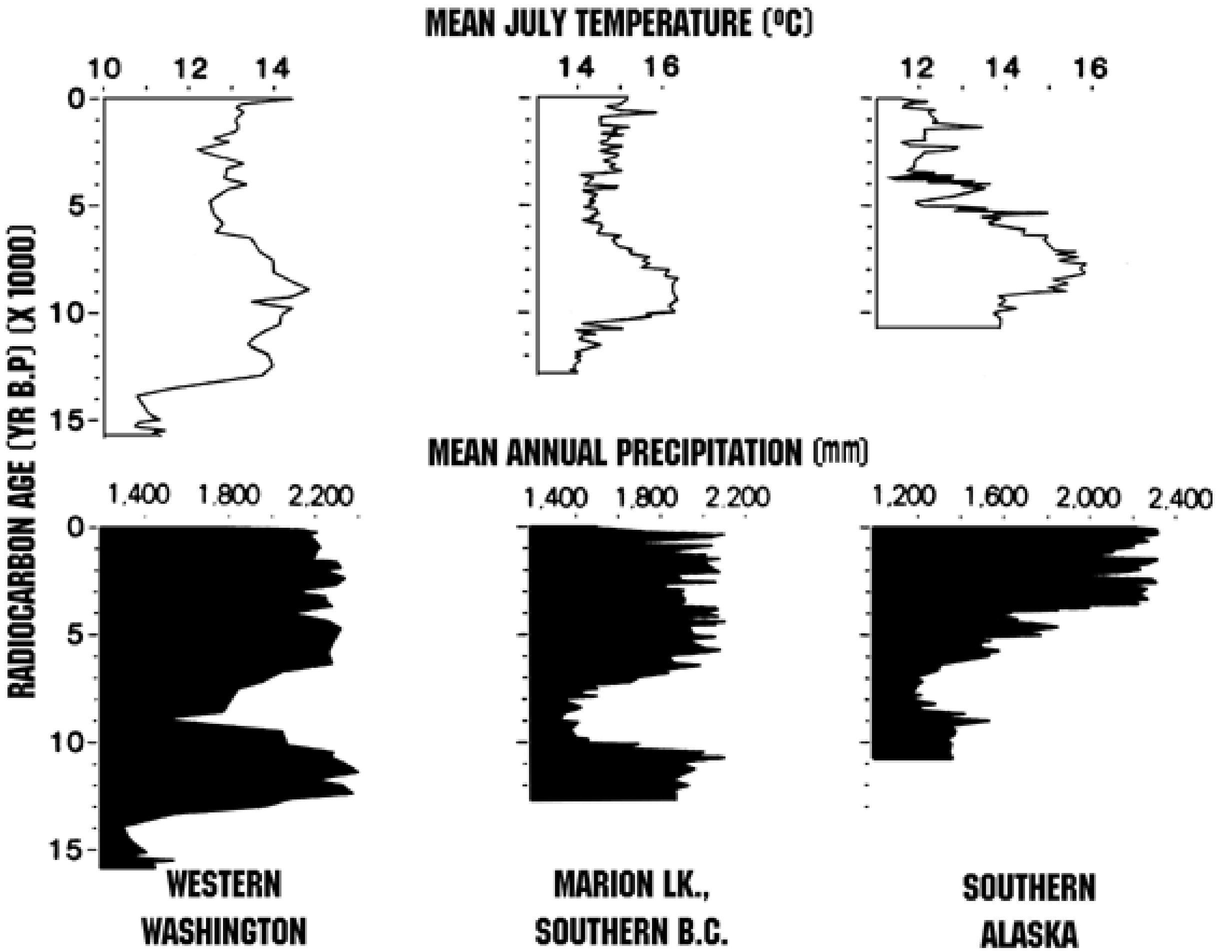


Figure 4. Comparison of pollen-based reconstructions of temperature and precipitation at three coastal sites: a) Olympic Peninsula, Washington, b) Marion Lake, near Maple Ridge, British Columbia, and c) near Icy Cape, Alaska (adapted from Heusser *et al.*, 1985).

Mathewes and Heusser (1981) based their postglacial reconstructions of mean July temperature and mean annual precipitation on the Marion Lake record. The reconstruction (Fig. 4) indicates a rapid transition from maximum warmth ca. 8000 ^{14}C yr BP to a July temperature minimum by 6000 ^{14}C yr BP. Cautious inspection of the pollen diagram suggests that the transition was more gradual, beginning by ca. 8000 ^{14}C yr BP, but much prolonged, perhaps extending to the present-day.

Heusser et al. (1985) compare temperature and precipitation trends reconstructed from three coastal sites (Fig. 4), distributed from northernmost Washington (the Olympic Peninsula) to southern Alaska (near Icy Cape in the Malaspina Glacier district). Maximum postglacial inferred temperatures span the 12,500 to 7500 ^{14}C year interval on the Olympic Peninsula, but persist more briefly, from 9000 to 6500 ^{14}C year BP, in southern Alaska. Minima in precipitation are inferred from 9500 to 7000 ^{14}C yrs BP on the Olympic Peninsula, and from 8500 to 6000 ^{14}C yr BP near Icy Cape. Comparing these three sites, the time-transgressive nature of the changes in temperature and precipitation is clearly evident, propagating northward along the coast. Heusser et al. (1985) inferred that the monsoon-type circulation was established with deglaciation, and that the intensity of the Aleutian low pressure centre gradually intensified later in the Holocene (see also Heusser *et al.*, 1999). Clague & Mathewes' (1989) discovery of abundant fossil wood dating from 9100 to 8100 ^{14}C yr BP, at a site 60 m above present timberline, provides additional evidence for early Holocene warmth. They infer early Holocene temperatures in the southern Coast Mountains, at least 0.4 °C warmer than current normals.

One curious feature of the coastal pollen record is the rather late arrival of Garry Oak (*Quercus garryana*) in southern British Columbia (Allen, 1995; Heusser, 1983b; Pellatt *et al.*, 2001). Although the maximum summer temperature has been inferred for the interval 9000 to ca. 5000 ^{14}C yr BP on the basis of midge fossils (Palmer, 1998), and from 10,000 to 8000 ^{14}C yr BP based on pollen records (Mathewes & Heusser, 1981), *Quercus* pollen was rare prior to 7500 ^{14}C yr BP and peaked at 7000 ^{14}C yr BP or later on southern Vancouver Island (Allen, 1995; Heusser, 1983b; Pellatt *et al.*, 2001).

A slow northward migration across the southern Gulf Islands to Vancouver Island, and thus, a long time lag following climatic change, offers a possible explanation for the tree's late arrival. However, it should also be noted that Garry oak is the only broad-leaved evergreen tree well represented in the regional pollen rain. Unlike the other pollen indicators, this species is likely very sensitive to winter temperature minima. Its absence may, therefore, indicate that Garry oak could not grow on the island during the early Holocene, despite higher summer temperatures, because winter temperatures were too cold. Greater seasonality may have been an important feature of early Holocene climate.

Pellatt et al. (2001) also note that Garry oak persists into the late Holocene, when summer temperatures are thought to have cooled significantly from their early Holocene maximum. They speculate that aboriginal burning practices may have played an important role in maintaining the oak savannah on southernmost Vancouver Island, despite less favourable climatic conditions.

Farther east, in the dry interior of British Columbia, pine pollen dominates the entire postglacial pollen record. The most noteworthy feature is a marked increase in grass and sage pollen during the early Holocene. This is especially evident at sites lying close to the present forest/grassland ecotone (e.g., Pemberton Hill Lake - Hebda, 1995; Crater Lake - Heinrichs, 1999; Fishblue and Chilhil Lakes - Mathewes & King, 1989), and indicates an interval of greater drought. The pollen record does not provide good means for resolving the precise reason for this drought. As indicated by the midge record, summer temperatures during the early Holocene were clearly 2 to 3 °C higher than today, and this would have contributed to drought conditions. However, it is difficult to resolve whether actual precipitation was lower than present, the same as present, or perhaps even slightly greater than that measured today. Furthermore, frequent fires likely played an important role in maintaining the more extensive early Holocene grasslands.

Low elevation pollen records from the Kootenay region of British Columbia are more difficult to interpret. Most records are dominated by pine, indicators of more extensive grasslands are less evident, and indications of early Holocene warmth or drought are more equivocal (see the pollen diagrams, for example, in Hills *et al.*, 1985). Nevertheless, Hebda's (1995) diagram for Bluebird Lake in the southern Rocky Mountain Trench indicates high sage (*Artemisia*) and grass (Poaceae) values in the early Holocene. It is therefore likely that grasslands were more extensive at this time.

Tree-line pollen records from the Rocky Mountains are more informative. Although the Wilcox Pass chronology (Beaudoin & King, 1990) from Jasper National Park is only based on two radiocarbon dates, the pollen record provides good evidence that local subalpine forests were already established in the early Holocene. A gradual decline in tree pollen (Pine, spruce, and fir) and increases in tundra vegetation components (e.g., willows, sedges, juniper, and spikemoss) indicate a late-Holocene retreat of timberline. Kearney and Luckman (1983a) inferred that timberline fluctuated, but was generally much higher 8700 to 5200 ^{14}C yr BP. Timberline retreated to lower elevations after 5200 ^{14}C yr BP, reaching the lowest recorded position within the last 500 years. Luckman (1988a) reports additional evidence for higher tree-line, in the form of 8000 year old wood from the snout of the Athabasca glacier.

Near treeline in Yoho National Park (Reasoner & Hickman, 1989), the pollen record is equivocal, but plant macrofossils and sediment lithology provide key evidence for more extensive forests in the early to mid-Holocene, and a down-valley retreat of timberline beginning ca. 5000 ^{14}C yr BP.

Salinity: Diatom and Midge Records

Like other aquatic organisms, diatom and midge species are strongly influenced by lake water salinity via the osmotic stress it imposes. Thus some species are restricted to freshwater habitats, whereas others may prefer moderately or highly saline environments. The factors controlling salinity can be complex, but salinity is generally expected to be high, and lake levels low, during episodes of drought, and/or when higher than normal temperatures prevail. At these times evaporative concentration of dissolved salts can increase the salinity of closed basin lakes.

In humid regions, such as coastal British Columbia, heavy precipitation regularly flushes lake basins with fresh water. However closed basin saline lakes are abundant throughout much of the southern and central interior of the province, and are potentially sensitive sites for monitoring past changes in evaporation/precipitation balance.

Although models have been developed to infer salinity changes from a variety of organismal remains, the reconstructions currently available are all based on diatoms and midges. We have already provided a brief synopsis of chironomid biology, but we have said nothing of diatoms.

Diatoms are unicellular algae and occur in essentially all aquatic habitats. They are especially sensitive to water pH and salinity, but are also widely used as palaeoindicators of phosphorous loading to lakes, and sometimes as indicators of water depth. Unlike other algae, the diatom cell wall is composed of silica. These ornate, highly sculptured cell walls (= frustules) preserve in lake sediments, and like midge remains, are readily identifiable.

Heinrichs et al. (1997) used both diatoms and midges to reconstruct the salinity history for Mahoney Lake, in the southern Okanagan Valley (Figure 5). The salinity reconstruction indicates an initial low salinity stage prior to 10,400 ^{14}C yr BP, perhaps a reflection of the prevailing, cool, late-glacial climate (as independently recorded by midge and pollen analyses previously outlined in this paper); however, the low salinity might just be attributable to an initial flush of glacial meltwater.

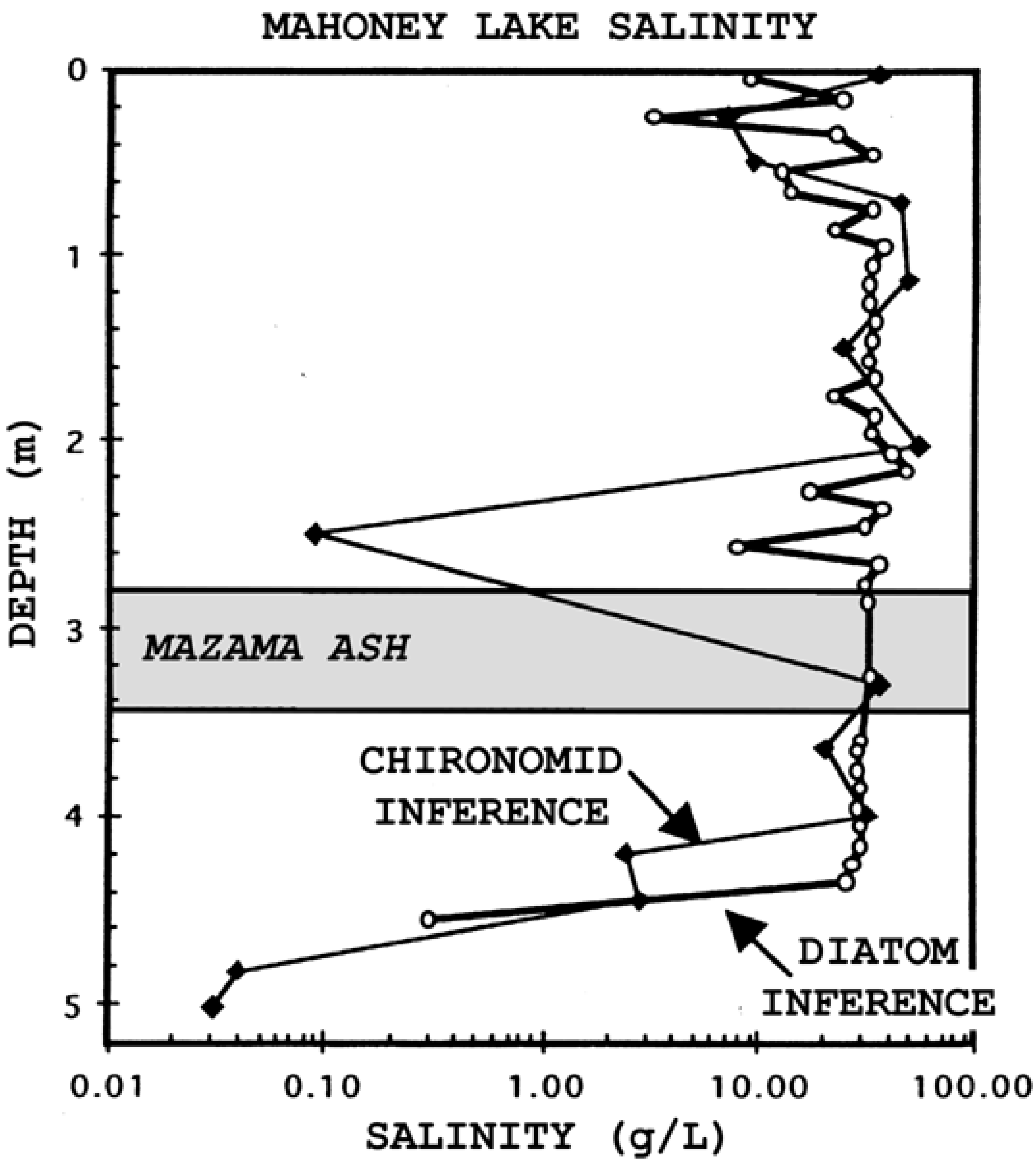


Figure 5. Salinity history for Mahoney Lake, southern Okanagan Valley based on diatoms and midges (adapted from Heinrichs *et al.*, 1997).

Salinity increased rapidly at Mahoney Lake during the transition from the late-glacial to the early Holocene (ca. 10,300 to 9000 ^{14}C yr BP), as would be expected with rapid postglacial warming. Salinity remained very high throughout the Holocene, except for a brief decrease in the mid-Holocene (ca. 6000 ^{14}C yr BP), and evidence for less saline waters in the late Holocene (after ca. 1000 ^{14}C yr BP).

Diatom and midge analyses are also complete for Big Lake in the Cariboo region, and Kilpoola Lake near Osoyoos (Bennett *et al.*, 2001; Heinrichs *et al.*, in press; Salomon, 1996). At Big Lake the diatom record indicates high salinities from 7800 ^{14}C yr BP to 5800 ^{14}C yr BP. The midge and pollen temperature reconstructions for other sites described earlier in this paper indicate higher summer temperatures throughout much of this interval. Unfortunately midge remains were scarce in Big Lake sediments dating to this time and the midge salinity reconstruction does not provide a good indication of postglacial salinity trends at Big Lake (Heinrichs *et al.*, in press). At Kilpoola Lake a mid-Holocene (ca. 6700 to 4500 ^{14}C yr BP) salinity peak is indicated in both the diatom and midge reconstructions (Heinrichs *et al.*, 1999b, in press; von Westarp, 1997).

Chaudhari (1998) used diatoms to infer postglacial changes in water depth at Burnell Lake, near Oliver in the south Okanagan. The diatoms indicate that water levels were generally low from ca. 8000 to 5200 ^{14}C yr BP, with water depth increasing in the late Holocene. Water depth, as inferred by the diatoms, has decreased gradually over the last 2000 years.

Elder (1997) and de Grace (1999) provide shorter (late-Holocene) diatom records for Fir Lake and Valentine Lake, respectively, in the Cariboo region. Valentine Lake became progressively fresher and deeper from 4200 to about 975 ^{14}C yr BP, when diatoms characteristic of shallower, more saline waters returned. Lower salinities have mainly prevailed since ca. 675 ^{14}C yr BP (de Grace, 1999).

At Fir Lake, salinity was high from ca. 2300 to ca. 1500 ^{14}C yr BP. From 1500 ^{14}C yr BP to the present-day, low salinities have prevailed, except for a brief, high salinity interval dating to about 1100 ^{14}C yr BP (Elder, 1997).

It would appear that the saline lake records vary greatly among basins. Changes in precipitation patterns may have been much more complex than originally anticipated. Shifting wind directions could produce spatially complex changes in the distributions of orographic precipitation and rain shadow effects. Alternatively, the complexity of these basins' hydrology may account for the differences evident among lakes.

Glacial History

The extent of glaciers and large ice sheets provides another index of past climatic changes. Growth of the ice sheets is obviously favoured in cold, humid climates, and not in warm or arid settings. Small glaciers and ice caps can respond rapidly to climatic conditions, but larger ice sheets will display a more inertial response (Booth, 1987; Wright, 1984).

About 20,000 ^{14}C yr BP, the Laurentide Ice Sheet reached its maximum late Wisconsinan extent (Booth, 1987; Fulton, 1989; Fulton *et al.*, 1986). The Cordilleran Ice Sheet attained its maximum somewhat later, with retreat in Hecate Strait evident after 15,600 ^{14}C yr BP (Blaise *et al.*, 1990). Farther south, in Washington state, the Puget Lobe of the Cordilleran ice sheet reached its maximum position ca. 14,000 to 15,000 ^{14}C yr BP (Booth, 1987; Ryder *et al.*, 1991), and then began to retreat northward into British Columbia. By ca. 11,000 ^{14}C yr BP, the Fraser Lowland was completely ice free and the Cordilleran ice sheet had collapsed (Ryder *et al.*, 1991; Souch, 1989).

The growth and retreat of the Cordilleran ice sheet seems to have lagged somewhat behind climatic changes in the region (Booth, 1987). Barnosky et al. (1987) note that local alpine glaciers had attained their maximum extent in western Washington between 22,000 and 19,000 ^{14}C yr BP, when the Puget Lobe was still advancing. On the Olympic Peninsula, tundra vegetation was displaced by trees 16,800 ^{14}C yr BP (Barnosky *et al.*, 1987).

At least one major readvance, the Sumas event, has been recognised in the Fraser lowland. Researchers have long speculated on the climatic significance of the readvance, with several scientists (e.g., Heusser, 1977) suggesting a correlation with the Younger Dryas (a cold interval very prominent in the North Atlantic region).

Recently, careful study and dating of deposits in the Fraser Lowland, has revealed that two advances actually occurred (Clague *et al.*, 1997). The first advance occurred before ca. 11,900 ^{14}C yr BP, and may correlate with either the Oldest Dryas cold period, or the Older Dryas cold period, in Europe. The second advance began ca. 11,300 ^{14}C yr BP, and may correlate with the Amphi-Atlantic (Killarney + Gerzensee) Oscillation (Clague *et al.*, 1997). Although the readvances may be climatically significant, sufficient evidence is so far lacking. The cooling was certainly minor relative to that evident in regions bordering the North Atlantic, and probably ended before the Younger Dryas began (Clague *et al.*, 1997). Clague et al. (1997) suggest that the readvances may instead have resulted from dynamic interactions among glacial retreat, isostatic rebound, sea level rise, and glacial surges. By about 9,500 ^{14}C yr BP, the glaciers in the Cordilleran region were no more extensive than they are today (Fulton, 1989).

Porter and Denton (1967) summarised much of the early evidence for subsequent Holocene glacier fluctuations in the North American Cordillera. They inferred that major glacial advances clustered around three time periods, ca. 4600 ^{14}C yr BP, ca. 2600-2800 ^{14}C yr BP, and ca. 200 to 600 ^{14}C yr BP.

Ryder (1989) provides a more recent compilation specific to the Canadian Cordillera. She notes that several very late Pleistocene or early Holocene moraines exist in the Shuswap Highlands and Rocky Mountains. The age of the Shuswap Highland advances is not well constrained by dates. Reasoner (1994) has established that the Crowfoot Advance in the Rockies is late-Pleistocene in age, correlating it with the Younger Dryas cold event. In the northern Cascade Mountains of Washington State, Heine (1998) and Thomas et al. (2000) provide evidence of glacial advances dating between 9000 and 9800 ^{14}C yr BP and 7700 and 8400 ^{14}C yr BP, respectively. Assuming a precipitation regime like that now extant, these advances would indicate temperatures averaging ≤2°C cooler than today's.

Ryder (1989) notes several Neoglacial advances, beginning as early as 6000 ^{14}C yr BP. The early Neoglacial, Garibaldi phase advances of Ryder & Thomson (1986) are recorded in the southern Coast Mountains, and probably date between 5000 and 6000 ^{14}C yr BP.

In the Coast Mountains several neoglacial advances date between about 1900 and 3300 ^{14}C yr BP, and are collectively referred to as the Tiedemann advance (Ryder & Thomson, 1986). At this time, advances of the Tiedemann (maximum 2300 ^{14}C yr BP), Gilbert (maximum 1900 ^{14}C yr BP), Frank Mackie (maximum 2700 ^{14}C yr BP) and Berendon (maximum between 2200 and 2800 ^{14}C yr BP) glaciers are noted (Clague & Mathewes, 1996; Ryder & Thomson, 1986). To the east, the Battle Mountain advance is recorded in the Shuswap Highlands (ca. 2400 to 3400 ^{14}C yr BP) and, in the Columbia Mountains, the Bugaboo Glacier was close to its maximum Neoglacial limit about 2500 ^{14}C yr BP (Osborn, 1986; Ryder, 1989).

Ryder (1987, 1989) notes abundant evidence of Holocene glacial maxima in the last millennium, dating between 100 and 900 ^{14}C yr BP. This interval is commonly referred to as the “Little Ice Age”. Retreat for some of these glaciers began in the 18th century, whereas the response of others was delayed, beginning only in the 19th or early 20th centuries (Ryder, 1987, 1989). At Tzeetsaytul Glacier in Tweedsmuir Provincial Park, Smith (2000b) notes evidence for a 17th century advance. A second advance was evident by 1815, culminating in terminal moraine deposition by 1853 (Smith, 2000b).

Farther north, Wiles et al. (1999) have used tree-ring dating to determine the age of Little Ice Age events. They note three intervals of strongly synchronous glacier movement: 1) a late 12th to 13th century advance, 2) a 17th to early 18th century advance, and 3) a late 19th century advance (Wiles *et al.*, 1999). Using dendrochronology Heikkinen (1984b) dated moraines on Mount Baker, in northwestern Washington, to the early 16th century, ca. 1740, ca. 1823, ca. 1855-56, ca. 1886-87, ca. 1908-12, ca. 1922, and 1978-79. Heikkinen (1984a; 1984c) also notes abundant evidence for forest expansion on Mount Baker over the last 100 years.

In the Rockies, Luckman (1977), used lichen growth curves to estimate the age of a “pre-Little Ice Age” moraine on Mount Edith Cavell. The moraine was dated prior to about 1800 years ago, whereas most other moraines on the same mountain dated to the 18th and 19th centuries. An advance of the Saskatchewan glacier killed trees in its path ca. 2850 ^{14}C yr BP (Smith, 2000a). Smith (2000a) considers this event to be contemporaneous with similar episodes at the Peyto, Yoho and Robson glaciers (Luckman, 2000), and as further confirmation that the Peyto Advance had regional significance.

Luckman (2000), and Luckman and Osborn (1979) report little other evidence of early- or mid-Neoglacial advances. In contrast, Little Ice Age events have been studied in great detail. Luckman (1986) infers, from ^{14}C dated snags, that warm conditions prevailed in the Rockies ca. 700 to 1100 AD. Luckman also notes evidence for a regional glacial advance between 1200 and 1370 AD, followed by episodes of moraine formation in the early 18th and mid 19th centuries (Luckman, 1986, 2000). Smith et al. (1995) have recorded glacial advances prior to the 16th century, in the early 17th and 18th centuries, and the mid 19th century in Peter Lougheed and Elk Lakes provincial parks, Alberta. Luckman (2000), however, concludes that although several moraines have been assigned dates prior to 1700 AD, the only convincing evidence for an earlier regional Little Ice Age event is for the 1200 to 1370 AD advance.

Using dendrogeomorphic techniques, the Little Ice Age maxima for several glaciers has been dated between 1830 and 1862 AD (Luckman, 1988b, 2000). Furthermore, Luckman infers from tree-ring evidence that most of the Little Ice Age advances were induced by cool summers, often accompanied with high summer precipitation (Luckman, 2000). However, Luckman also notes that climate varied continuously throughout the Little Ice Age. A long interval of sustained cool summers was only evident in the 19th century (Luckman, 2000).

In the Premier Range of British Columbia, ice retreated rapidly ca. 1930 to ca. 1955 AD (Luckman *et al.*, 1987). This was followed by glacial readvances, induced via cool summers (ca. 1954 to 1968 AD) in combination with greater winter precipitation (ca. 1951 to 1976 AD). The glaciers had receded slightly in the 1980s, as warmer summers returned, and winter precipitation decreased (Luckman *et al.*, 1987).

Dendroclimatology

Analyses of tree ring thickness have long been recognised as potential sources for long climate time series. Given the remarkable longevity of many tree species native to southern British Columbia, there is probably no region in the world better suited for collection of these data.

Growth of trees, especially near the trees' climatic limits, is obviously related to favourable weather conditions. Trees add all of their incremental growth in the spring and summer, but must build the carbohydrate reserves needed for this growth over the preceding year; thus, weather conditions in both the current year and the preceding year are often important determinants of tree-ring width. Growth at arctic and alpine treeline is widely regarded as being temperature limited. Growth at the grassland-forest ecotone is more often limited via summer drought.

Several investigators have examined tree-ring widths in the Pacific coastal region of North America. The growth responses they report are more complex (Heikkinen, 1985; Laroque & Smith, 1999; Peterson & Peterson, 1994; Smith & Laroque, 1998b; Zhang *et al.*, 2000). They indicate, that growth of subalpine conifers in the Pacific Northwest, is positively influenced by warm early summer temperatures. Nevertheless, drought and unusually high late summer temperatures can stress the trees, leading to reduced growth in the following year.

In coastal British Columbia precipitation also influences subalpine tree growth. Spring snowpack thickness is negatively correlated with most trees' growth. Since low temperatures, and high fall and early winter precipitation promote snowpack accumulation, they too impede tree growth. In contrast, Laroque and Smith (1999) note that February rain may hasten snowmelt and/or increase the soil moisture reservoir, promoting radial growth in Yellow Cedar (*Chamaecyparis nootkatensis*).

El Niño winters seem to favour Mountain Hemlock (*Tsuga mertensiana*) growth (Smith & Laroque, 1998b). These winters are characterised as having either 1) cool summers beginning with a thin snowpack, or 2) very warm summers with deep snowpacks in the spring season (Smith & Laroque, 1998b). The tree-rings have been used to develop a proxy record of variability in the mean Pacific Decadal Oscillation Index, spanning the interval 1600 AD to the present-day (Gedalof & Smith, 1999; Smith, 1999). Smith (1999) and Gedalof & Smith (1999) note that interdecadal climate variability was more pronounced prior to 1875.

Wiles et al. (1996, 1998) have used coastal tree-ring time series to reconstruct spring/summer land air temperature variations over the past few centuries. The temperature reconstruction based on tree-ring series from southwestern coastal British Columbia and adjacent Washington reveals a particularly cool interval ca. 1800 to 1820 AD (Fig. 6). They also note that the land air temperatures are strongly positively correlated with North Pacific sea surface temperatures, an index of intensity of the Aleutian Low, and salmon catches in the Gulf of Alaska. Finney et al. (2000) provide evidence that these salmon returns have strongly influenced the nutrient supply, algal species composition, and zooplankton community of coastal Alaskan sockeye lakes.

In 1984, Luckman et al (1984) reported living seven hundred year old spruce (*Picea*) and Pine (*Pinus*) from the Rocky Mountains, and began detailed studies of this potential source for long climate time series. In addition to tree-ring and densitometric measurements, they incorporated isotopic analyses of tree rings into their work (Luckman *et al.*, 1985). Their preliminary research revealed that tree-ring widths were positively correlated with summer temperature. Latewood density was also positively related to summer temperatures, but earlywood density was negatively related (Luckman *et al.*, 1985).

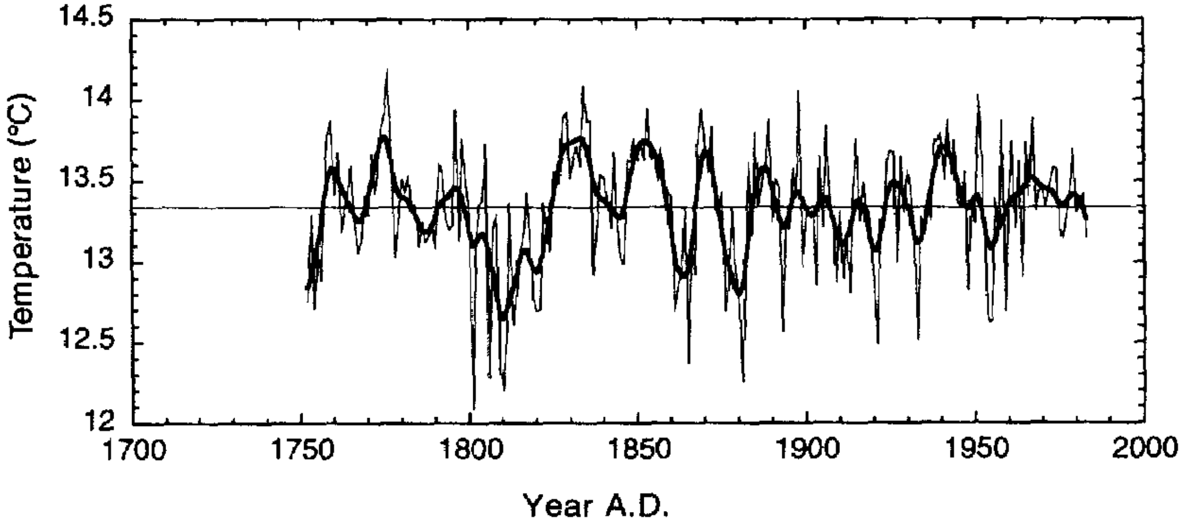


Figure 6. Warm season temperature reconstruction for the Pacific northwest (adapted from Wiles *et al.*, 1996).

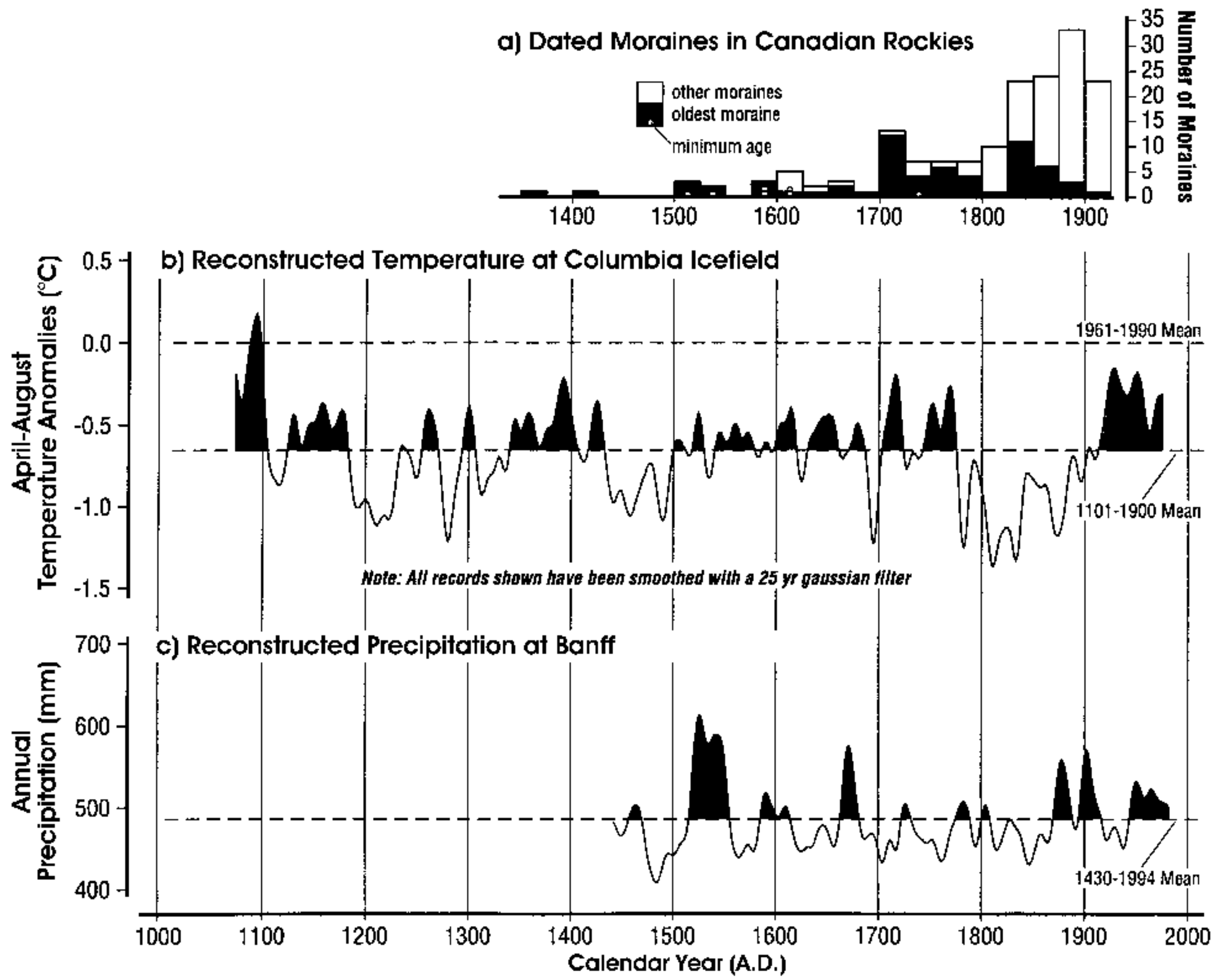


Figure 7. a) Age of moraines, b) Dendroclimatological reconstruction of temperature and c) reconstruction of precipitation trends for the Canadian Rockies (adapted from Luckman, 2000).

Luckman's research has culminated in a 900-year reconstruction (1073 to 1983 AD) of summer temperature anomalies for the Columbia Icefield (Fig. 7a). Prolonged cool intervals are recorded ca. 1190 to 1250, ca. 1440 to 1500, ca. 1690-1700, and ca. 1780 to 1900 AD. The reconstruction clearly reveals that the 19th century was anomalously cold relative to all other intervals in the past millennium. The summer temperature anomaly averaged -1.04°C from 1781 to 1900 (Luckman, 1998b). In contrast, the 20th century was anomalously warm (Luckman, 1998b; Luckman *et al.*, 1997). According to Luckman (1998b, p. 323), "The mean temperature anomaly from 1101 to 1900 was 0.71°C below the 1961-1990 reference period and 0.33°C below the 1891-1990 mean of the instrumental record."

Case & MacDonald (1995) used limber pine (*Pinus flexilis*) to develop precipitation reconstructions for the Alberta foothills. Their 487 year record identifies especially severe droughts in the decade 1790 to 1800, and ca. 1640. Although they chose to reconstruct precipitation from their tree-ring data, the growth of these trees is also adversely impacted by low summer temperature. They do not discuss the difficulty of separating between the temperature and precipitation signals.

Watson and Luckman have also developed long (approx. 550 years) tree-ring records (Fig. 7b) of precipitation from Douglas Fir (*Pseudotsuga menziesii*) in Banff and Jasper National Parks (Luckman, 2000; Watson, 1998; Watson & Luckman, 2001). These records indicate several episodes of anomalously high summer precipitation, dating ca. 1515 to 1550, ca. 1585 to 1610, ca. 1660 to 1680, and ca. 1880 to 1890 AD (Luckman, 2000). Watson (1998) also developed precipitation reconstructions for sites at Kamloops, Westwold, Penticton and Cranbrook in British Columbia, using Douglas fir and Ponderosa pine chronologies, Watson's (1998) data identifies the 1810s, 1870s, 1900s and 1940s as being wet decades, whereas dry decades in southern British Columbia, over the past 200 years, include the 1840s, 1860s and 1920s.

Comparison of the tree-ring reconstructed temperature and precipitation patterns with the glacial record fails to reveal a simple relationship between the two (Luckman, 2000). Luckman (2000) stresses that the Little Ice Age may be characterised as a time of more extensive glacial activity, but it certainly was not a sustained interval of colder or wetter conditions.

Isotopes

Isotopic records have so far been little used for palaeoclimatic research in the Canadian Cordillera. All of the records we consider were collected as a part of dendroclimatological studies of living or subfossil wood (Clague *et al.*, 1992; Edwards & Luckman, 1996; Luckman & Gray, 1990; Luckman & Kearney, 1986; Luckman *et al.*, 1985; Spittlehouse & Livingston, 1999). Spittlehouse and Livingston (1999) indicate that Douglas Fir ring widths on southern Vancouver Island are drought sensitive. While these ring series may potentially be used as palaeo-precipitation indicators, they have also determined that carbon isotopic discrimination in the late-wood will provide a clearer summer rainfall signal (Spittlehouse & Livingston, 1999) .

At Castle Peak, ^{18}O and ^{2}H isotopic analyses on subfossil wood (9100 to 8100 ^{14}C yr BP) indicate that the trees had experienced humidity levels comparable to living trees at modern treeline (60 m below the fossil site). Clague et al. (1992) infer that the present, mean, growing season daytime cloudbase lies at 2285 m, slightly below the cirque rim that encloses the fossil site. In the early Holocene, the cloudbase would have been well above the cirque rim (Clague *et al.*, 1992). This inference suggests that the site would have been drier than today, and is consistent with other evidence for warmer, drier, early Holocene conditions at Castle Peak.

Luckman & Kearney (1986) note strong positive correlations between $\delta^{18}O$ from α-cellulose and mean annual temperature near Jasper. They use this relationship to infer that mean annual temperature was at least 0.5°C warmer ca. 8060 and 8770 ^{14}C yr BP at the Watchtower Basin, and about 1.2 to 1.6°C warmer at Maligne Pass ca. 6000 and 5300 ^{14}C yr BP. Long isotopic time series are being developed for the Canadian Rockies by Edwards & Luckman (1996), but only preliminary results have so far been published.

Ocean Temperature

Few relevant palaeoceanographic studies have been conducted adjacent to the British Columbia coast. Nevertheless, this research has tremendous potential because 1) long records spanning several glacial/interglacial cycles are available, and 2) these records provide essential data revealing the extent of ocean-atmosphere linkages in the northeastern Pacific Ocean. Marret et al. (2001), for example, report a 430,000 year record of sea surface conditions (e.g., February temperature, August temperature, salinity and sea-ice cover) from the Gulf of Alaska. Although the values for reconstructed variables fluctuate widely throughout the record, late Pleistocene conditions (relative to the middle Pleistocene) were generally cooler year-round with lower salinities and more persistent sea- ice cover. The last 10,000 years are exceptional however, recording evidence of the warmest (or nearly the warmest) conditions in the past 430,000 years.

Shorter records are also available. Sabin & Pisias (1996) used radiolarian fossils to infer sea surface temperatures for the past 20,000 years in the northeastern Pacific. At 50°N latitude, postglacial warming culminated in peak sea surface temperatures ca. 12,000 to 8000 ^{14}C yr B.P. Foraminifera at 42°N latitude record dramatic evidence of a cooling event correlative with the Younger Dryas (Mix *et al.*, 1999). This is probably the strongest evidence of a Younger Dryas impact from anywhere along the Pacific Northwest coast.

Solar Insolation, Circulation Models, and Modern Climatic Analogues

A consideration of climate model simulations and past changes in solar insolation provides essential background for interpreting the palaeoclimate record (e.g., Anderson *et al.*, 1988; Barnosky *et al.*, 1987; Berger & Loutre, 1991; Ganopolski *et al.*, 1998; Kutzbach & Guetter, 1986; Kutzbach & Wright, 1985; Rind, 2000; Weaver & Green, 1998). In addition, essential insights can be gleaned from sunspot records, for example, and modern analogues for past climatic conditions (Beer *et al.*, 1996; Mock & Brunelle-Daines, 1999; Vance, 1987).

Over the past two decades, there has been a growing recognition that solar insolation patterns have played a fundamental role in regulating Earth's climate. Cyclic changes in the earth's orbit (tilt (=obliquity), eccentricity and precession), commonly referred to as the Milankovitch cycles, are responsible for long-term changes in solar insolation (e.g., Beer *et al.*, 1996; Berger & Loutre, 1991; Weaver *et al.*, 1998). In addition, short-term changes in insolation are evident due to changes in solar activity (e.g., Beer *et al.*, 2000; Fairbridge & Hillaire-Marcel, 1977; van Geel *et al.*, 1998, 1999).

Insolation curves for mid-to-high latitude northern hemisphere sites (Fig. 8) indicate a summer insolation minimum ca. 20,000 cal. yr BP, a summer insolation maximum ca. 10,000 cal. yr BP, and a progressive decrease in summer insolation over the past 10 millennia (e.g., Barnosky *et al.*, 1987; Kutzbach & Guetter, 1986; Wright, 1984); thus, summer insolation changes generally parallel the long-term climatic trends inferred from British Columbia data. Nevertheless, a lag of 2000 to over 5000 years is apparent in many of the palaeoenvironmental climatic proxy records.

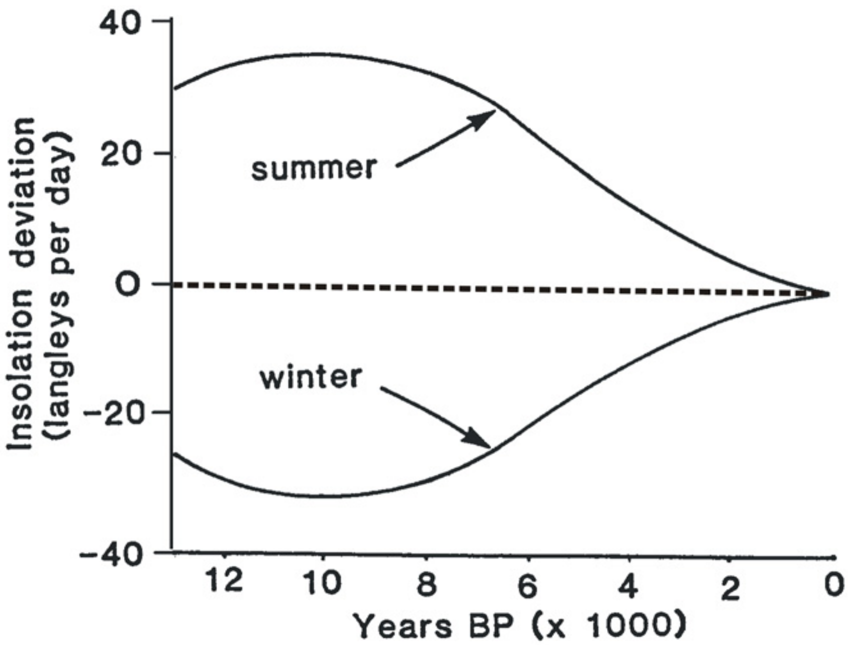


Figure 8. Solar insolation curve for 50°N latitude. Changes in solar insolation are expressed relative to current conditions (adapted from Vance, 1987).

The southern margin of the Cordilleran ice sheet achieved its maximum between 14,000 and 15,000 ^{14}C yr BP, not at the 20,000 cal. yr BP summer insolation minimum. Chironomid-inferred mean July temperatures indicate an early Holocene (ca. 8000 ^{14}C yr BP) maximum, when summer insolation was already declining. In part these discrepancies are an artifact of the real differences between radiocarbon and calendar years (Fig. 9). For example, the systematic error between these dates causes events that occurred 18,000 cal. years ago (16,000 BC) to be dated at 15,000 ^{14}C yr BP. Similarly, events at 15,000 and 12,000 cal. years ago are dated using radiocarbon to 12,500 and 10,200 ^{14}C yr BP, respectively. Thus, the maximum advance of the southern margin of the Cordilleran ice sheet lagged behind the summer insolation minimum by about 2000 years, not 5000 years.

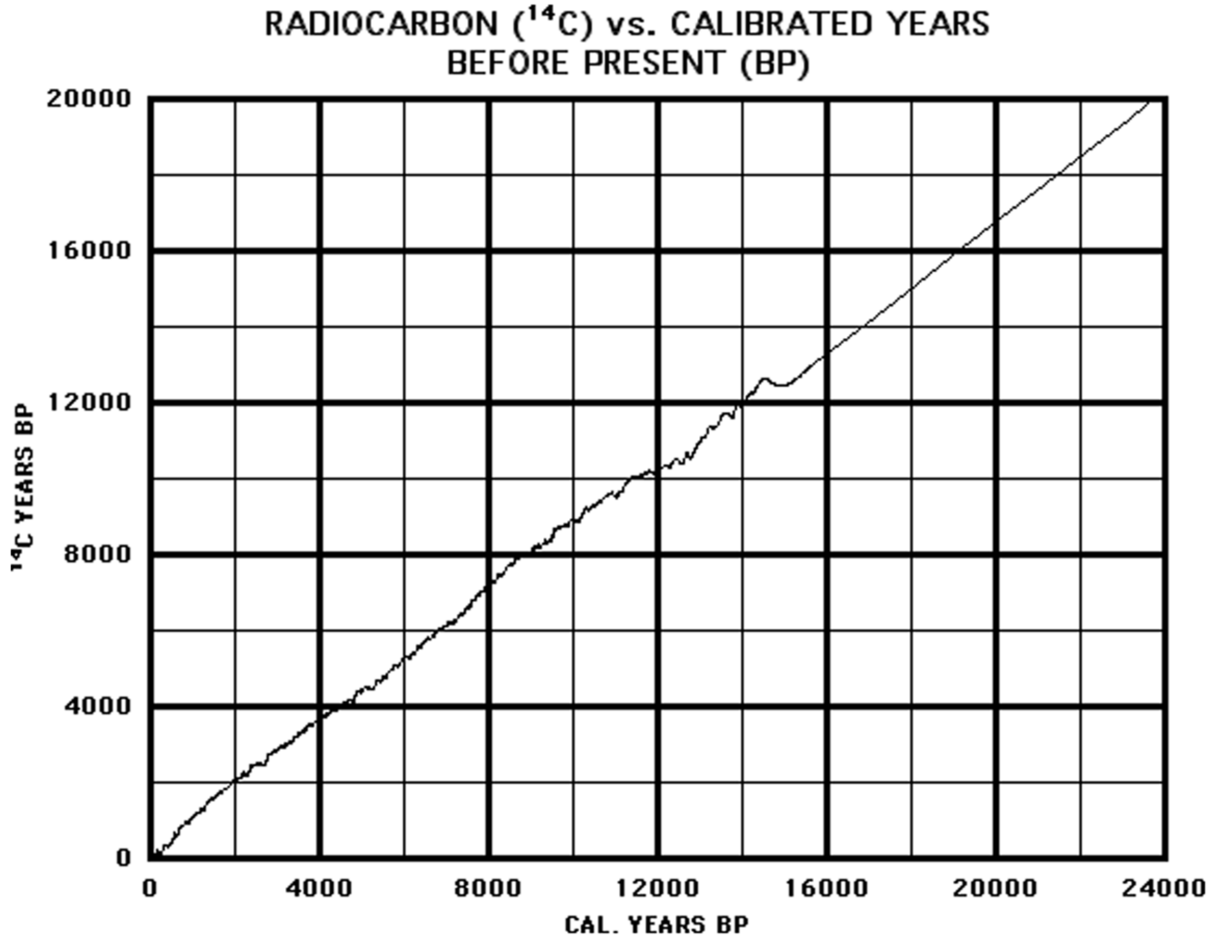


Figure 9. Calibration curve depicting the relationship, and systematic error, between radiocarbon and calendar years, as inferred from tree-ring and varve chronologies (based on data provided with Stuiver *et al.*, 1998).

The remaining (real) lags between solar insolation and the palaeoenvironmental data reflect, not only the response rates typical of ice sheets and organisms to climatic change (Wright, 1984), but also the importance of other factors (e.g., albedo) and feedback mechanisms, in regulating our past and present climate. Having very short life cycles, great reproductive potential, and efficient means for dispersal, insect communities probably responded very rapidly to climatic change; hence their utility in reconstructing past climates. The large ice sheets had a more inertial response, and their contribution to the earth's albedo would have delayed climatic warming in spite of peak summer insolation ca. 10,000 cal. yr BP.

The winter insolation curve (Fig. 8) for this interval is inverse to the summer curve, indicating a maximum prior to 20,000 cal. yr BP, a minimum ca. 10,000 cal. yr BP, and a progressive increase thereafter (e.g., Barnosky *et al.*, 1987; Kutzbach & Guetter, 1986; Wright, 1984). The sharp early Holocene contrast between summer and winter insolation seems to be little apparent in those palaeoenvironmental data we have reviewed. The Garry oak pollen record, however, possibly indicates greater seasonality in the early Holocene with colder winters than those extant today.

The importance of the Milankovitch cycles to long-term (i.e., glacial/interglacial) climatic events has been accepted for almost two decades. The solar contribution to decadal to millennial scale climatic variability has been recognised only very recently (Beer *et al.*, 2000; Crowley, 2000; van Geel *et al.*, 1999). Past solar activity is recorded in historical records of sun spots and aurorae. Cosmogenic isotopes (^{14}C, ^{10}Be, ^{36}Cl) record prehistoric solar activity (Beer *et al.*, 1996).

Beer et al. (2000) note solar cycles of 11, 22, 88 and 208 years duration. Distinct minima in solar activity (Fig. 10) include the Wolf Minimum (1282 to 1342 AD), the Spörer Minimum (1416 to 1534 AD), the Maunder Minimum (1645 to 1715 AD), the Dalton Minimum (1800 to 1820 AD), and the 1900 Minimum (1880 to 1900 AD). According to Beer et al. (2000), these minima seem to coincide with glacier advances, lake level changes, and the sudden deterioration of climate.

The solar minima seem frequently to coincide with cold intervals in Luckman's (2000) records of tree-ring-inferred temperatures and glacial advances in the Canadian Rockies. For example, temperatures were anomalously cold, and glaciers were advancing throughout much of the Wolf Minimum (Luckman, 2000). Many of the moraines in the Canadian Rockies date to the Maunder Minimum (Luckman, 2000). Similarly Wiles et al. (1998) indicate anomalously cold spring temperatures in the Gulf of Alaska region during the Maunder Minimum. The Dalton Minimum coincides with the coldest years in the Pacific Northwest tree-ring temperature reconstruction (Wiles *et al.*, 1996). Over the last 140 years, almost half of the variance in northern hemisphere temperatures can be explained by solar variability (Beer *et al.*, 2000). Similarly, for the northern Hemisphere, Crowley (2000) has determined that the combined influences of solar variability and vulcanism can account for approximately half of the pre-1850 decadal scale temperature variations.

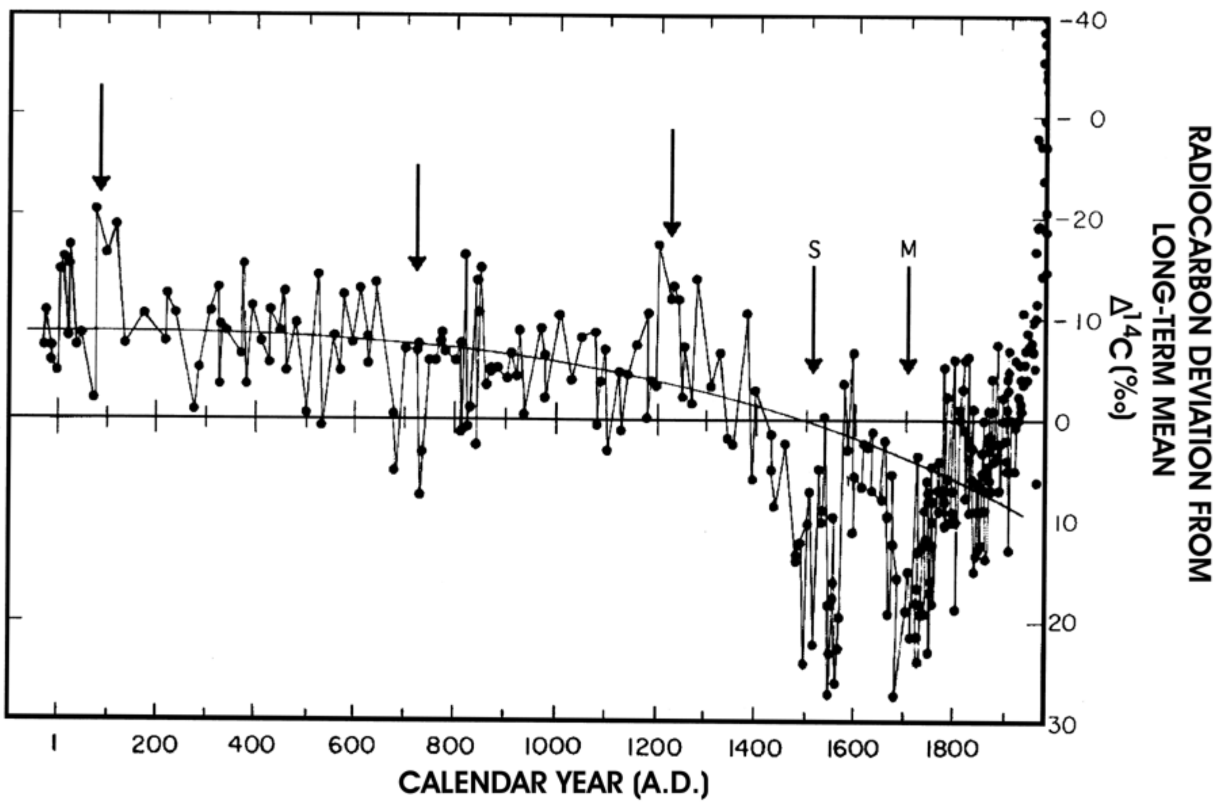


Figure 10. Changes in atmospheric ^{14}C over the past 2000 years. Arrows indicate periods of reduced solar activity, including the Spörer (S) and Maunder (M) minima. Adapted from Bradley (1999).

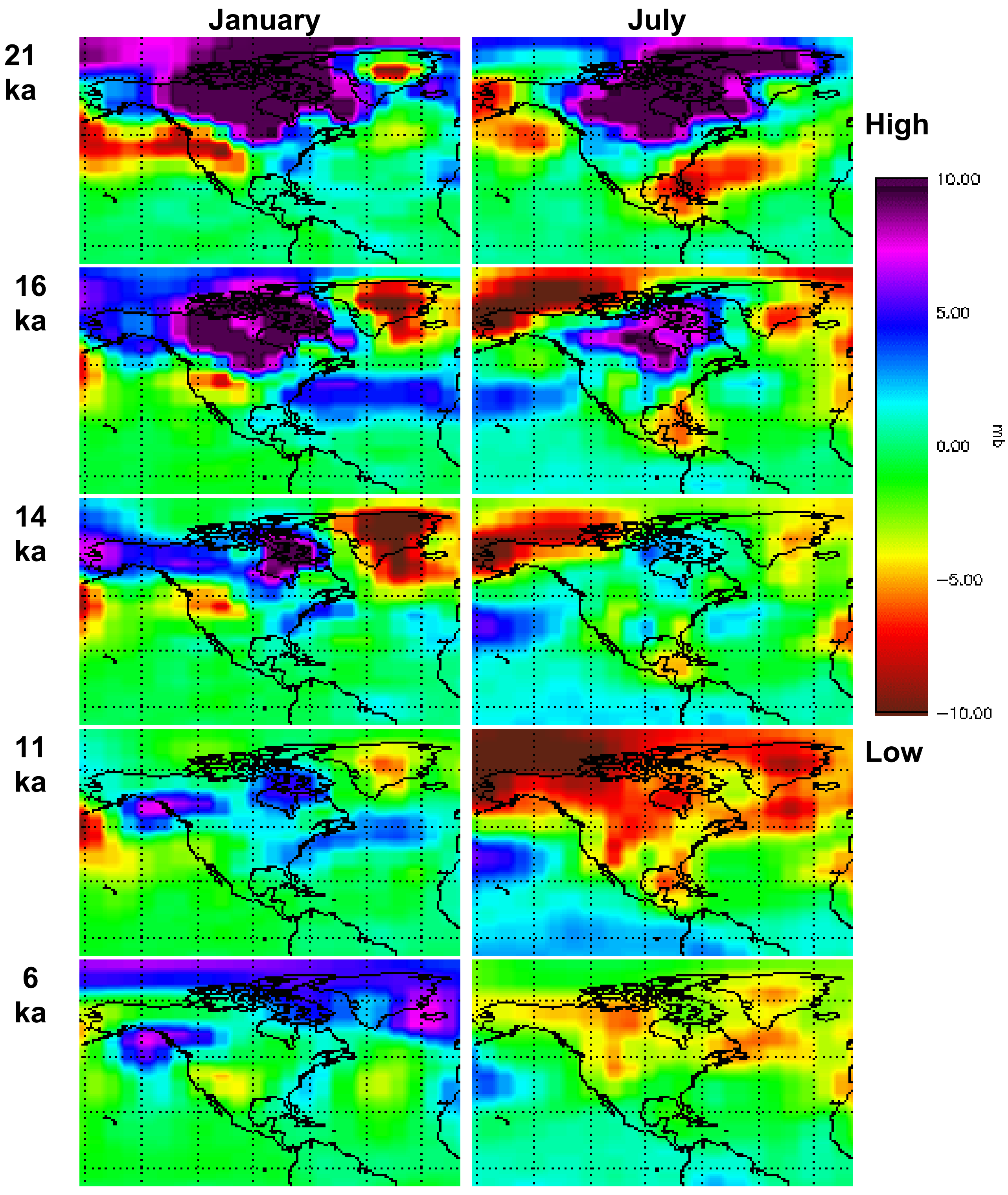


Figure 11. Sea level pressure anomalies (21,000, 16,000, 14,000, 11,000 and 6000 cal yr BP minus modern) for North America, as simulated with the CCM1 general circulation model (Bartlein *et al.*, 1998; Felzer *et al.*, 1998; Kutzbach *et al.*, 1996; Kutzbach *et al.*, 1998). Illustration prepared using NOAA's Paleoclimate Output Visualization Tool (NOAA, 2000).

Palaeoenvironmental reconstructions may be compared against model simulations of past climate 1) to assess the validity of the models' reconstructions, and 2) to gain new insights for interpretation of the palaeoenvironmental data (Anderson *et al.*, 1988; Barnosky *et al.*, 1987; Kutzbach & Guetter, 1986). For western North America (Fig. 11), changes in the winter position of the jetstream, and accompanying changes in the relative strengths of 1) the Aleutian low in winter, 2) the Pacific high during summer, and 3) a persistent anticyclone over the ice sheets, have featured prominently in climate simulations and palaeoenvironmental interpretations (Anderson *et al.*, 1988; Barnosky *et al.*, 1987; Kutzbach, 1994; Kutzbach & Guetter, 1986; Kutzbach *et al.*, 1996; NOAA, 2000; Thompson *et al.*, 1993). Kutzbach & Ruddiman (1993), and Kutzbach et al. (1993) summarise some of the important limitations of these models. Perhaps most importantly, Kutzbach & Ruddiman (1993) recognised that newly emerging evidence of a systematic bias in the radiocarbon ages meant that all of the boundary conditions they had derived from radiocarbon-dated geologic data and had used with the CCM0 model were improperly positioned with respect to the insolation values. This discrepancy created some obvious and unfortunate problems in assessing the model output, and in comparing the output vis-à-vis the palaeoenvironmental record. However, this mismatch between simulations and observations has been corrected in recently published papers (e.g., Bartlein *et al.*, 1998; Felzer *et al.*, 1998; Kutzbach *et al.*, 1998). Fortunately the new CCM1 model output largely concurs with that derived in earlier simulations (Bartlein *et al.*, 1998).

Despite summer and winter insolation values similar to those extant today, the 21,000 cal. yr BP simulation depicts a climate markedly different than today's (Fig. 11); thus, the unusual features are clearly attributable to the extent of the continental ice sheets, sea ice and other boundary conditions inferred from geological evidence (Kutzbach *et al.*, 1993).

During the last glacial maximum the winter jetstream over North was displaced south of its current position (Bartlein *et al.*, 1998). A glacial anticyclone was centred over the Laurentide and Cordilleran ice sheets (Fig. 11), with easterly winds prevailing along the southern ice margin. These easterly winds carried cold, dry, continental air across the Rockies into the Great Basin, and over the Cascade Mountains to coastal regions (Anderson *et al.*, 1988; Barnosky *et al.*, 1987; Kutzbach & Guetter, 1986). This circulation pattern ensured that dry, cold conditions prevailed throughout our study region, during the glacial maximum.

At this time, lower sea surface temperatures also reduced the supply of moisture to the northwest coast, and the glacial anticyclone blocked Pacific air from entering far into the study area. Furthermore, the sharp temperature contrast between land and sea likely ensured that most precipitation was restricted to the westernmost portion of our study area - the exposed continental shelf (Anderson *et al.*, 1988; Barnosky *et al.*, 1987). It is, therefore, not surprising, that palaeobotanical evidence from interior Washington indicates a cold, dry steppe environment. On unglaciated lands west of the Cascade Mountains, tundra or parkland vegetation prevailed at 21,000 cal yr BP (Barnosky *et al.*, 1987). Because the ice sheets were advancing to envelope British Columbia at this time, there is little palaeoecological data available from the province.

The 14,000 cal. yr BP climate simulation suggests, that as the Cordilleran Ice retreated and eventually collapsed, the winter jetstream shifted northward. The glacial anticyclone weakened (Fig. 11), allowing moist Pacific air to penetrate onshore (Anderson *et al.*, 1988; Barnosky *et al.*, 1987). By 14,000 cal. yr BP, parts of coastal British Columbia were ice free, and by 12,000 cal. yr BP (10,000 ^{14}C yr BP) the Cordilleran ice sheet had largely collapsed. The pollen record depicts a progressively warming climate.

In the pollen record a gradual transition from dry to wet conditions is also apparent. Since this dry to wet transition dates to 13,000 cal. yr B.P. (about 11,000 ^{14}C yr BP), researchers have been tempted to draw a correlation with the onset of the Younger Dryas (e.g., Mathewes, 1993; Mathewes *et al.*, 1993).

Further retreat of the Laurentide Ice gradually weakened the glacial anticyclone (Fig. 11), and by 10,000 cal. yr BP), summer insolation had peaked. A strong subtropical high allowed for warmer and drier summers than today's (Anderson *et al.*, 1988; Barnosky *et al.*, 1987; Kutzbach, 1994; Kutzbach & Guetter, 1986; Kutzbach *et al.*, 1996; NOAA, 2000; Thompson *et al.*, 1993). Pollen and chironomid temperature inferences both suggest that July temperatures were warming rapidly ca. 12,000 cal. yr BP in southern British Columbia (Fig. 1, 4). Although July insolation peaked 10,000 cal. yr BP, winter insolation had achieved a minimum 10,000 cal. yr BP, thus greater seasonality, and colder winters may have been typical of the early Holocene in British Columbia (e.g., Vance, 1987). The CCM1 climate model indicates a January high pressure anomaly (Fig. 11) over northern British Columbia (Kutzbach & Guetter, 1986; Kutzbach *et al.*, 1996; NOAA, 2000). Arctic air, and accompanying coastal outflow winds, may have been more prevalent in British Columbia during winter.

The atmospheric circulation pattern was, therefore, changing very rapidly through the late-glacial. The shifts from a cold, dry late-glacial summer climate, to first a cool, wet state, and later, a warm, dry early Holocene summer climate occurred rapidly. In southern British Columbia, the approximate coincidence of these two transitions with the beginning and end of the Younger Dryas suggests a link with the Younger Dryas event. An alternate explanation, however, exists.

The climate simulation indicates that the changes may be explained by a progressive weakening of the glacial anticyclone and winter Aleutian low, and a gradual strengthening of the summer Pacific high (Anderson *et al.*, 1988; Barnosky *et al.*, 1987; Kutzbach, 1994; Kutzbach & Guetter, 1986; Kutzbach *et al.*, 1996; NOAA, 2000; Thompson *et al.*, 1993). This suggests an important test. If these late-glacial climatic changes are linked to the Younger Dryas, they would be sudden, and evident at all Pacific Northwest sites, at precisely the same time. If the Younger Dryas was not involved, and the changes were instead linked to more local and gradual changes in pressure and circulation patterns, the changes would be time-transgressive, with the initial shift to wetter conditions, and the subsequent shift to warmer, drier conditions being at first evident in the south, and later at the more northerly sites. Barnosky et al. (1987), for example, indicate an earlier shift from cold and dry, to cool and humid conditions ca. 17,000 cal. yr BP (15,000 ^{14}C yr BP) in the southern Puget Trough.

Since the early Holocene the Laurentide ice sheet and accompanying anticyclone have disappeared (Fig. 11). The climate simulations indicate that the summer Pacific high has weakened while the Aleutian low has grown stronger (Anderson *et al.*, 1988; Barnosky *et al.*, 1987; Kutzbach, 1994; Kutzbach & Guetter, 1986; Kutzbach *et al.*, 1996; NOAA, 2000; Thompson *et al.*, 1993). In the early Holocene the strong high delivered dry northwesterly air to British Columbia in summer. This feature would result in clearer skies, and when combined with higher summer insolation, readily accounts for the prevailing warm, dry early Holocene summers evident from the fossil evidence. The weaker Aleutian low would have reduced the winter supply of warm, moist air from the south; thus, winters may have been colder but drier. As the last remnants of the Laurentide Ice disappeared from Québec/Labrador in the mid-Holocene, the glacial anticyclone collapsed, and a stronger flow of Pacific air penetrated onto the Prairies in summer, maintaining dry conditions in the centre of the continent (Anderson *et al.*, 1988; Vance, 1987). Since the ice sheets were much smaller in the early Holocene, and had all but disappeared by the mid-Holocene, their presence had little impact on Holocene climate simulations. Consequently Holocene changes in circulation were mostly driven by changes in solar insolation (Kutzbach *et al.*, 1993).

Implications - Applications of Palaeoenvironmental Knowledge

The preceding review reveals that much is known about past climatic trends and variability in southern British Columbia, adjacent regions elsewhere in British Columbia, and inside Alberta, Washington and Idaho. Given that this information exists, we may rightly question, “How can it be applied?”, “How can we harness these tools to better understand future climate?”, and “How might it help us to adapt to the changes, which now seem inevitable?” In this section we will attempt to demonstrate how such data may be applied, and to suggest useful future directions for this research.

Climate Modelling

To date, the most important applications of the palaeoenvironmental data, have related to climate modelling. There has, for example, been considerable debate concerning 1) the reliability of the projected climate trends over the next century, and 2) the extent to which recent climate trends reflect natural climate variability versus anthropogenic greenhouse effects.

The climate models are obviously very complex, but, at the same time, are rather simplistic representations of an even more complex real world. Have the teams of modellers and other atmospheric scientists managed to accurately capture and depict all of the important climate-related processes in their models? If a model seems capable of representing the Earth’s climate, how can it be rigorously tested? We have only one planet Earth. When we develop a model to simulate our planet, we are left with no replicate systems where it may be tested.

In developing the models, palaeoenvironmental data are providing essential information, indicating how sensitive the models need to be to the various natural and un-natural factors forcing current climate. As our earlier presentation of the dendroclimatic evidence indicated, there has been considerable climate variability over the past several centuries. This variability is driven by many processes, including for example variable solar forcing, vulcanism, and greenhouse gas emissions.

Since the various palaeoenvironmental proxy records also comprise isotopic records of past solar variability, historical and ice core evidence of past volcanic eruptions, and direct measurements of atmospheric gases long trapped inside polar ice caps, it has been possible to quantitatively estimate the relative roles of each of these factors in shaping recent climate trends. As previously asserted, over the last 140 years, almost half of the variance in northern hemisphere temperatures can be explained by solar variability (Beer *et al.*, 2000). Similarly, for the northern hemisphere, Crowley (2000) has determined that the combined influences of solar variability and vulcanism can account for approximately half of the pre-1850 decadal scale temperature variations. Nevertheless, it is also these detailed palaeoenvironmental records that allow scientists to confidently conclude, 1) "… that the increase in temperature in the 20th century is likely to have been the largest of any century during the past 1000 years", 2) "this warming was unusual and is unlikely to be entirely natural in origin", and 3) "the projected rate of warming is much larger than the observed changes during the 20th century and is very likely to be without precedent during at least the last 10,000 years (IPCC, 2001).

The ability of a model to simulate the Earth's present climate provides a poor basis for assessing its abilities to simulate future climates. Yet, in the absence of replicate systems, how can the models be properly tested? For their tests, climate modellers have relied greatly on palaeoenvironmental data – instead of projecting climate into the future, the models may be used to project backwards in time. If the models provide backward projections of climate consistent with the palaeoenvironmental data, we may also have confidence in their forward projections of future climates.

Current models portray ancient climates that are largely consistent with the palaeoenvironmental record (Bartlein *et al.*, 1998), and this adds greatly to our confidence in their projections. Nevertheless, the models are based on very coarse geographic grids, and do not portray regional-scale climate variations. In such a mountainous region as British Columbia, the models are incapable of simulating local variations, for example, the wet belts and rain shadows produced by our mountain systems. Modellers are working to overcome this difficulty by linking the global general circulation models to regional models incorporating, for example, a more realistic topography.

It should be clear from this account that palaeoenvironmental data are of great importance to the climate modelling community. Not surprisingly, in its recommendations for further action, the Intergovernmental Panel on Climate Change (IPCC, 2001) recognised the need to "enhance the development of reconstructions of past climate periods".

Resource and Protected Areas Management

British Columbia is a resource-rich province. Although the service and technological sectors of its economy are growing rapidly, much of the province's wealth is supported by resource-based industries. Furthermore, British Columbians greatly value their natural surroundings, and have worked to establish an elaborate system of parks and other protected areas "for preservation, and the enjoyment of future generations". Resource and protected area managers are, therefore, facing a daunting task. Given that the Intergovernmental Panel on Climate Change (IPCC, 2001) has concluded that Earth's climate is very likely changing at a pace unprecedented in the last 10,000 years, how do we best protect the values of our lands and renewable resources for both ourselves and future generations?. To effectively respond to this challenge, our land and resource stewards critically need:

1) reliable climate projections
2) detailed assessments of how climate change will impact, for example, forests, grasslands, water and fisheries
3) a much better understanding of how the rate of climate change affects a natural system's capacity to adjust to future climate states.

In the preceding section, we have already stressed the role of palaeoecology in the development and testing of climate models. These tests lend confidence to their future projections. The projected climate changes vary among models and with region in the province (Taylor & Taylor, 1997). For southern British Columbia, current models project a summer warming of up to 4°C with a doubling in atmospheric CO_2. Winter temperature increases, according to the CCC GCM II model, may be greater, up to 6 or 7°C in southeastern B.C. Precipitation projections vary markedly among the models; thus, it is uncertain whether annual precipitation may increase or decrease (Taylor & Taylor, 1997). However, the warmer temperatures will likely reduce the winter snowpack and increase evaporation; thus, extending the duration of the summer drought..

Suggested outcomes of these climatic changes include:

1) increased outbreaks of insect pests and pathogenic fungi
2) increased extent and frequency of forest fires
3) an upward and northward expansion of grasslands and forests
4) earlier spring freshets, with lower peak discharges
5) loss of all or most glaciers in southern B.C.
6) lower water levels in lakes and streams, especially accompanying the summer drought.

Some species (e.g., many of the rare and endangered grassland species of the South Okanagan and Similkameen) might benefit from these changes. However, the habitat of alpine tundra species and coldwater fishes would contract. These species would be particularly at risk, because of the natural isolation of individual watersheds and mountain peaks. Given the discontinuous nature of these habitats, obvious corridors for species migration are not apparent, and the extinction of sensitive species seems likely. Ranges of long-lived plant species (e.g., alpine plants and many trees) are unlikely to keep pace with the changes, and are therefore also susceptible to losses. Where possible (e.g., for low elevation terrestrial species) corridors for species migration should be provided. In other cases, where habitats are discontinuous, human intervention might be needed to transplant species, facilitating their northward spread. In the case of commercially valuable species, this spread would be accommodated easily by adapting silvicultural practices. However, to what extent would people, or their governments, be willing to assist other species?

The changing species' distributions in British Columbia are also likely to be more complex than commonly assumed. In British Columbia, the biogeoclimatic zone concept is a tool widely used in forest management, and has been used in Hebda's (1997a) initial assessments of climate change impacts. As Hebda clearly recognised, however, the biogeoclimatic zone system is based on a Clementsian view of plant communities, to which few plant ecologists now adhere. These biogeoclimatic units, including their characteristic assemblages of plant and animal species, may be readily identifiable in the context of our current climate, but as species react to climate change, their individualistic responses may generate new species assemblages (biogeoclimatic units) without present-day analogues. This Gleasonian view of plant assemblages is widely acknowledged in palaeoecology, where late-glacial and early Holocene species assemblages often have no analogues in the modern world.

The dilemma for protected area managers is particularly troublesome. Protected areas (e.g., national parks) are set aside to preserve a range of values, including ecosystem integrity (i.e., the composition and abundance of native species and biological communities, rates of changes and supporting process) (Parks Canada, 2000). Should that integrity be based on the ecological make-up of the area:

1) prior to European settlement?
2) as derived from First Nation land-use patterns?
3) when the protected area was established?
4) given current conditions?
5) with adaptation to future global warming?

Some of the features we wish to preserve (e.g., many geological or landscape features) may be insensitive to climate change, whereas others (especially glaciers and biota) may be greatly threatened. In the face of changing species ranges it might be necessary to adjust protected area boundaries to protect target species. We may, therefore, need to develop a dynamic view of area boundaries. Perhaps a more viable alternative, however, is to consider migration corridors and the projected ranges of plant and animal species when selecting areas for protection. One of the important lessons of palaeoecology is that species ranges and assemblages are dynamic. Although the rate of change may be increasing, communities are naturally dynamic entities. It is unrealistic to expect them to remain as constants through time. As climates evolve, so do our biota and landscapes.

It will be critical in coming years to develop a much more detailed understanding of how climate change affects biota. Some kinds of change are reasonably well understood, and are clearly evident in the fossil record (e.g., the transition from tundra to forest to grassland with a warming and drying climate). Other changes are less well understood, and may only be apparent in observations of long-term correlations with climate events, as revealed in the palaeoenvironmental record. In other cases, ideas developed from short-term observations may be best tested using the much longer records obtained using palaeoenvironmental techniques.

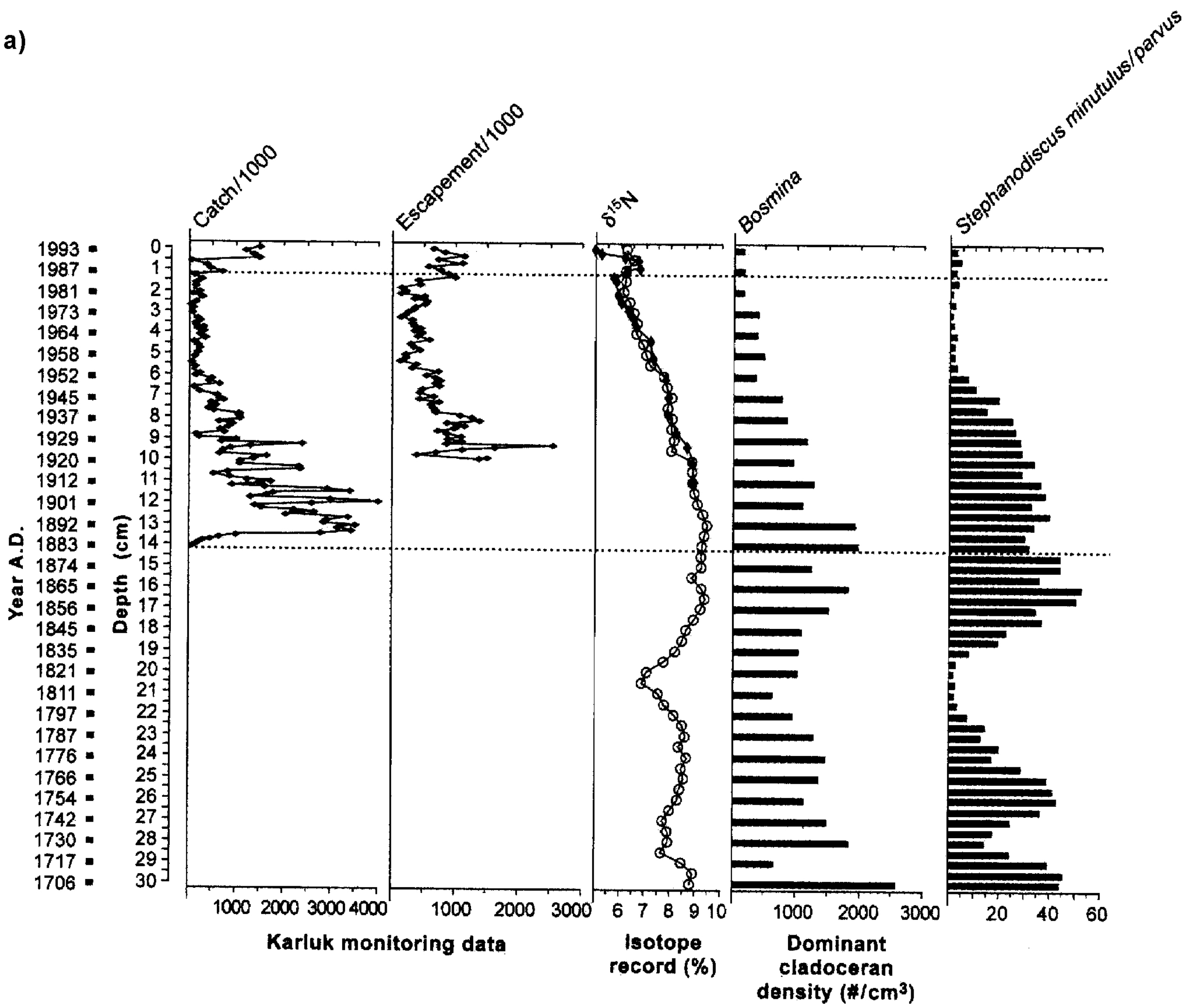
Recent fisheries research provides a dramatic illustration of how palaeolimnological research is proving to be an excellent tool for understanding baseline ecosystems and ecological integrity (e.g., Finney, 1998; Finney *et al.*, 2000). The structure and function of British Columbia's coastal ecosystems is strongly linked to complex land-sea interactions and the absence of frequent disturbance events, such as hurricanes and fires (Franklin *et al.*, 1991). Of particular importance to ecological, societal, and economic systems are the anadromous pacific salmon. Recent fisheries research (Bilby *et al.*, 1996; Cederholm *et al.*, 2000; Kline *et al.*, 1990, 1993; Reimchen, 2000) illustrates the intrinsic linkages in the transport of marine derived nutrients into freshwater and terrestrial ecosystems from salmon carcasses and eggs. At the time we are beginning to better understand the nutrient and food web dynamics of these ecosystems and the role of pacific salmon as keystone species, we are also witnessing a dramatic decrease in the number of salmon that are returning to their parent streams.

Research into the role of marine derived nutrients from salmon has illustrated the utility of using the stable isotope ^{15}N, which exists in higher abundance in marine systems than in terrestrial systems, as a marker for the transport of marine derived nutrients into freshwater and terrestrial systems. ^{15}N is also preserved in lake sediments and has been shown to be a proxy for past sockeye salmon (*Oncorhynchus nerka*) abundance (Finney *et al.*, 2000).

Finney et al. (2000) observed a decline in ^{15}N and marine derived nutrient dependent diatoms corresponding with declines in salmon numbers from ~ 2,000,000 to 500,000 between 1910 and 1970 in Karluk Lake, Alaska due to commercial fishing (Fig 12). The loss of marine derived nutrients resulted in impacts on aquatic communities that are dependent on external nutrient sources in these oligotrophic systems.

Current palaeolimnological research being undertaken by MacIsaac and Finney (2001) reveals similar decreases in ^{15}N, as well as changes in *Bosmina:Daphnia* ratios based on decreased predation from juvenile sockeye salmon in the Shuswap Lake system of the southern interior of British Columbia. This research is revealing short and long-term changes in salmon populations in response to commercial fishing, migration blockages (Hell's Gate slide), and climate change. The research of MacIsaac and Finney (2001) indicates that the carrying capacity of Shuswap Lake for juvenile salmon may be less than half of what it was in the late-1800s. This decrease in carrying capacity may be strongly associated with the reduction in sockeye salmon carcasses as a source of nutrients for the lake ecosystems.

a)


Catch/1000
Escapement/1000
δ15N
Bosmina
Stephanodiscus minutulus/parvus
Year A.D.
Depth (cm)
Karluk monitoring data
Isotope record (%)
Dominant cladoceran density (#/cm3)

b)

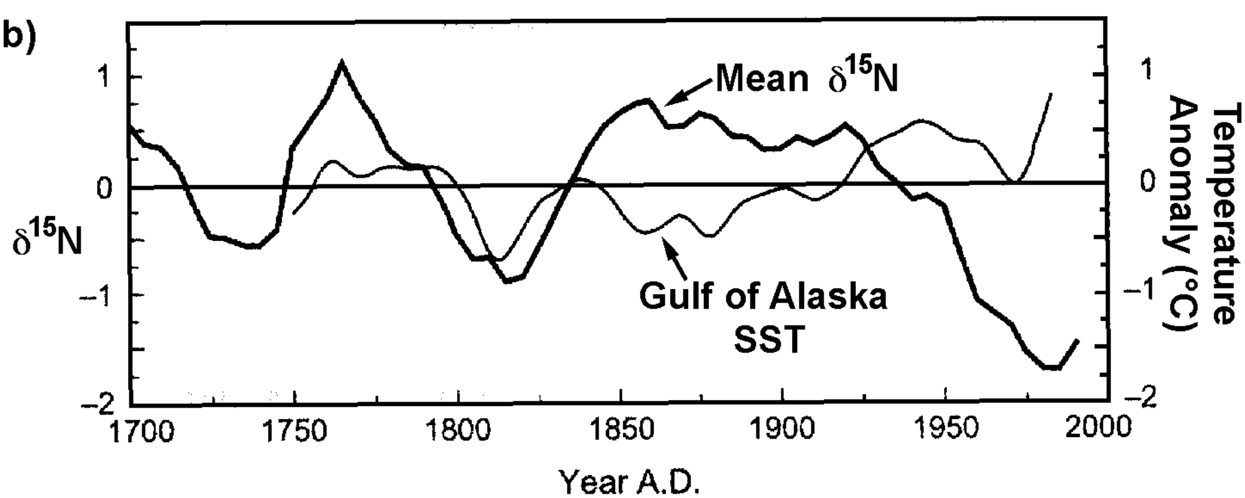

Mean δ15N
Gulf of Alaska SST
δ15N
Temperature Anomaly (°C)
Year A.D.

Figure 12. a) Historical and palaeolimnological records of sockeye runs, nutrient loading and lake biota in Karluk Lake, Alaska; b) Comparison of $\delta^{15}N$ record of marine nutrient loading and tree-ring reconstruction of Gulf of Alaska sea surface temperature (adapted from Finney *et al.*, 2000)

Ongoing field work in the central interior of British Columbia at Quesnel Lake (MacIsaac – Department of Fisheries Ocean) and the west coast of Vancouver Island (Pellatt - Parks Canada) will help fisheries managers and ecologists understand the importance of marine derived nutrients, the commercial salmon fishery, and climate change on aquatic and terrestrial ecosystems of British Columbia.

We expect that as managers attempt to better understand natural drought and forest fire cycles, they will also be drawn to the palaeoecological record. The dendroclimatic reconstructions of Watson (1998), for example, reveal long-term records of drought, a variable of critical importance in agriculture. In the Canadian Rockies, charcoal records in sediment cores are being exploited by Parks Canada as means to understand natural patterns of forest fire frequency (Hallett & Walker, 2000) and insect outbreaks (Rob Walker, pers. communication).

Furthermore, the current rate of climatic changes is unrivalled in the past 10,000 years, but, if we extend this window slightly backward in time, it encompasses the late-glacial/Holocene transition. Ice core studies have revealed that the transition from a glacial to an interglacial climate may have occurred in ≤ 50 years. This time period, therefore, provides an exceptional window for testing the responses of biota to rapid climatic changes. Was this rapid climatic warming accompanied by widespread insect outbreaks? Or fires? How long was the lag time between climatic changes and individual species' responses? The answers to these questions have obvious applicability to understanding how biota will respond to current and future change. By combining isotopic records of climate change with the fossil record of species' responses, we should be able to better understand ecosystem responses to rapid climatic change. Models of species responses that have been developed from modern observations might be best tested against the palaeoecological record, before they are used to simulate ecological responses to future change.

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