Climatic and landscape controls of the boreal forest fire regime: Holocene records from Alaska

JASON A. LYNCH, JEREMY L. HOLLIS and FENG SHENG HU

Department of Plant Biology, University of Illinois, 265 Morrill Hall, 505 South Goodwin Avenue, Urbana, Illinois 61801, USA

Summary

1. The response of ecosystems to past and future climatic change is difficult to understand due to the uncertainties in the direction and magnitude of changes and the relative importance of interactions between climate and local factors. In boreal ecosystems such interactions may dictate the response to climatic change, but the interaction of climate, vegetation composition and the fire regime remains poorly understood.

2. Sediment cores from lakes in south-central Alaska were analysed for lithology, macrofossils, pollen and charcoal to investigate the relationships between moisture availability, species composition and mean fire return intervals (MFI).

3. Macrofossil and lithological evidence suggests that variations in effective moisture occurred over the past 7000 years and that the regional climate has been wetter during the past c. 3800 years than before.

4. Boreal forests existed in the region throughout the past 7000 years. *Picea glauca* and *Picea mariana* were the prevalent forest species around Chokasna Lake, whereas *P. glauca* and hardwood species (e.g. *Betula*) co-dominated the landscape around Moose Lake. *Picea mariana* replaced *P. glauca* as the dominant *Picea* species around Chokasna Lake at c. 2000 BP.

5. MFI was > 500 years before 3800 BP, except from 5800 to 5000 BP at Moose Lake and 5400 to 4550 BP at Chokasna Lake, when values were c. 200 years. MFI decreased to c. 200 years during the late-Holocene at both sites and to c. 150 years after 2000 BP at Chokasna Lake.

6. At both sites, fires occurred more frequently under wetter climatic conditions. Our results therefore support other recent studies demonstrating that warmer/drier climatic conditions do not necessarily induce greater fire importance. A combination of increased ignition by lightning strikes and seasonal-moisture variability probably resulted in more frequent fires under wetter conditions.

Key-words: boreal ecosystem, charcoal analysis, climate change, fire frequency, forest dynamics

Introduction

Predictions of future climatic warming have instigated numerous studies of natural climatic variations and their effects on ecosystems using palaeorecords (e.g. Foley et al. 1994; IPCC 2001). The boreal ecosystems are of particular interest because they are projected to be among the most sensitive to future climatic forcing. Changes in temperature and precipitation are likely to alter disturbance regimes, the most important being fire (Flannigan et al. 2001; Chapin et al. 2003). Boreal forest fires play an important role in determining forest composition, energy fluxes and biogeochemical processes (Shugart et al. 1992). Thus changes in fire frequency induced by climate warming are likely to have major consequences on the dynamics of boreal forest ecosystems.

Fire occurrence depends on weather variations, the long-term trends in climate, the build-up of fine and coarse fuels, the spatial arrangement of fuel, and human activity (Clark 1989a). Weather variations can affect the fire regime on daily to monthly time-scales by changing the
frequency of fire ignition and the amount of moisture in fine fuels (Clark 1989a; Nash & Johnson 1996; Rorig & Ferguson 1999). Humans can alter the fire regime through fire ignition or suppression (Clark & Royall 1995). However, the impact of these variations strongly depends on the amount and arrangement of fuels available for ignition (i.e. fine fuel load) and fire spread (i.e. coarse fuel accumulation, and species composition), as well as other disturbances (e.g. insect outbreaks). These factors vary on decadal to centennial time-scales and on stand to landscape spatial scales (Heinselman 1973; Clark 1989a). The long-term climatic trend ultimately controls the effects of short-term variations in weather (e.g. drought cycles), fuel dynamics through climate-dependent rates of fuel production and decomposition, and local vegetational type (Clark 1990). Therefore, knowledge of the relationships among these factors operating at various temporal and spatial scales is necessary for understanding boreal fire dynamics and predicting future response to climatic change.

The interactions among vegetation, climate and fire are poorly understood for Alaskan boreal ecosystems. Because both fuel dynamics and climate vary at time-scales longer than typical observational studies, the palaeo-ecological approach is particularly useful. Although previous palaeoecological studies in that region provide a wealth of information on vegetational changes, they offer no quantitative estimates of fire frequency (Hu et al. 1993, 1996; Earle et al. 1996) or lack independent reconstructions of past climate (Hu et al. 1993, 1996; Earle et al. 1996; Lynch et al. 2002). These studies speculated that climate dictated the long-term fire regime by altering vegetational composition and fuel load. For example, a mid-Holocene increase in fire importance was attributed to the establishment of *Picea mariana* (Mill) BSP (black spruce) dominated communities following a climatic change to colder/wetter conditions in Alaska (Hu et al. 1996; Lynch et al. 2002). Similarly, Payette (1992) suggests that increased abundance of *P. mariana* and fire frequency also coincided with cooler climatic conditions during the past 2000 years in north-eastern Canada. In contrast, others (e.g. Larsen 1996; Carcaill et al. 2001) showed that climate directly influenced the fire regime through changing moisture, temperature and ignition frequency in Canada. For example, warmer and drier climatic conditions than at present in western Alberta during the mid-Holocene led to an increase in fire frequency (Hallett & Walker 2000), and greater seasonality of precipitation in north-western Ontario was inferred to have caused a decrease in fire frequency after 3000 calendar years before 1950 (bp) (Carcaill & Richard 2000; Carcaill et al. 2001).

In this paper, we used pollen, charcoal, macrofossils and lithological characteristics in sediment cores from two lakes in south-central Alaska (Copper River Valley) to reconstruct vegetation, mean fire interval (MFI) and effective moisture during the past 7000 bp. These reconstructions provided the basis to elucidate how climatic conditions and vegetational composition affected the regional fire regime.

**Study sites**

Moose (61°22.45′ N, 143°35.93′ W, 434 m a.s.l.) and Chokasna (61°22.40′ N, 143°6.14′ W, 537 m a.s.l.) lakes are topographically closed basins located between the Wrangell Mountains and the Chitina River in Wrangell-St Elias National Park and Preserve in south-central Alaska (Fig. 1). These lakes are c. 20 km apart. The surface area and maximum depth are 21.5 ha and 4.5 m for Moose Lake, respectively. For Chokasna Lake, surface area is 11.0 ha and maximum depth is 2.0 m. These lakes were chosen for their small surface areas and absence of inlets and outlets in order to minimize regional pollen deposition and charcoal transport through surface flow.

The annual mean temperature at McCarthy, Alaska, located 32 km east of Moose Lake, is −2.1 °C, with mean summer and winter temperatures of 10.5 °C and −14.5 °C, respectively (http://www.wrcc.dri.edu). The mean annual precipitation is 45.0 cm. The modern vegetation of the region is typical of closed boreal forests in Alaska (Viereck et al. 1992). Mixed forests of *Picea glauca* Moench (white spruce), *Populus tremuloides* Michaux (quaking aspen) and *Betula papyrifera* Marsh (paper birch) occupy well-drained upland sites, whereas *P. mariana*-dominated forests occur at poorly drained lowland sites. *Alnus crispa* (Ait.) Pursh (green alder) forms thickets on mountain slopes above 500 m.

Local vegetation differs at the two lakes as a result of variation in topography and soil moisture. Moose Lake lies in a rolling landscape where soils are moderately to well-drained Typic Cryorthods (Rieger et al. 1979). A mixture of deciduous (*P. tremuloides* and *B. papyrifera*) and coniferous species (*P. glauca*) dominates this landscape. *Picea mariana* is occasionally intermixed within the upland *P. glauca* forests. A moderately dense, tall shrub layer occurs in the understory that is comprised mostly of *A. crispa* with some *Salix* spp. *Picea mariana* dominates only a small proportion of the watershed at the western end of the lake where soils are poorly drained. In contrast, Chokasna Lake lies in a broad flat plane with poorly drained Histic Pergelic Cryaquepts soils that contain permafrost (Rieger et al. 1979). A relatively homogeneous *P. mariana* forest occupies most of the watershed. The only exception is a small area south-west of the lake where broadleaf species and *P. glauca* are common. A low-density layer of *Alnus* and *Salix* shrubs occurs in the understory. The shoreline vegetation includes *Alnus rugosa* (speckled alder), *Salix* spp., and a mix of Cyperaceae and Poaceae species. A few patches of *Nuphar* spp. (pond lily) occur in the lake.

The present-day fire regime of south-central Alaska is characterized by low fire frequencies compared with that of interior Alaska. Since 1950 only 24 fires burning a total area of 4058 ha have occurred in this region (Alaskan Forest Service 1999). The last fire within the watersheds of the study lakes occurred in 1915 based on
Historic accounts (Lutz 1956) and tree ages surrounding the lakes. Today people are the main ignition source in this region, and lightning strikes are rare (Gabreil & Tande 1983). Before the first American explorers arrived in the late 1800s, Tlingit, Yupik and Dena’ina (the Athabaskan tribe) people occupied south-central Alaska (Gibson 1976). Unfortunately, little is known about the use of fire by the Athabaskan tribe, but their lifestyle suggests that they did not substantially modify the natural fire regime. Tlingit, Yupik and Dena’ina people were fisherman and elk hunters. Household use and food preparation were probably their primary use of fire. Elk hunting occurred in the winter months when fire spread was ineffective. The small size and low population density of the tribe would have further limited its effect on the natural fire regime.

The fire season lasts from April to September, with fires occurring more frequently in *P. mariana*-dominated forests than in other vegetational types. *Picea mariana* tends to grow in densely packed stands with lichen-draped lower branches and a well-developed carpet of feather mosses (*Pleurozium schreberi, Hylocomium*...
splendens, Ptilium crista-castrensis), which facilitates fire ignition and spread (Viereck 1973). Thin bark and shallow roots make P. mariana prone to fire mortality, but their semiserotinous cones enable reseeding following fires. In contrast, P. glauca-dominated stands are typically less dense with a less flammable understory (Nienstaedt & Zasada 1990). Frequent fires eliminate P. glauca because of its low seed production until 30 years old. Deciduous species (e.g. B. papyrifera and P. tremuloides) are also less flammable, and their canopies intercept more sunlight than conifers (Brown & Davis 1973), creating cool and moist conditions in the understory (Van Wager 1983), which reduce the likelihood of fire ignition and spread.

Materials and methods

We obtained two stratigraphically overlapping sediment cores from Moose Lake and a single core starting at 40 cm below the sediment–water interface from Chokasna Lake. These cores were taken near the deepest location of each lake with a Streif corer equipped with an electric hammer. An intact surface–water interface and the upper 53 cm of sediment were sampled with a polycarbonate tube at Moose Lake. The Moose Lake cores were correlated on the basis of numerous sediment layers commonly present in all cores.

Core chronologies were based on $^{14}$C and $^{210}$Pb dates for Moose Lake and on $^{14}$C dates for Chokasna Lake. Seven and five AMS $^{14}$C dates on wood and Picea needles were obtained for Moose and Chokasna lakes, respectively (Table 1). Radiocarbon ages were converted to calibrated years before 1950 ($\text{BP}$) with Calib 4.3 (Stuiver et al. 1998). The median probability age was taken to be the calibrated age. Lead-210 analysis (Eakins & Morrison 1978) was done on contiguous 1-cm sediment samples from the surface core of Moose Lake until background values were reached. A constant rate of supply (c.r.s.) model was used to calculate ages with first-order errors (Binford 1990).

For Moose Lake, multiple approaches were used to calculate sample ages according to sediment characteristics (Fig. 2). A linear regression model fitted through 6834, 6559 and 6493 $\text{BP}$ was used to determine the ages of samples > 6493 years old. The ages of 6493 and 6004 $\text{BP}$ were from the top and bottom boundaries of a peat layer, respectively; we used linear interpolation to assign ages throughout this peat layer. A locally weighted regression (loess) function was used to assign ages for samples < 6004 years old.

At Chokasna Lake, a loess function was used to assign sample ages (Fig. 2). The calibrated ages of 5162 $\text{BP}$ at 144 cm and 5166 $\text{BP}$ at 184 cm are statistically the same. We assumed that the date of 5162 $\text{BP}$ on a large wood fragment was too old because of the greater tendency for wood than for needles to have long residence times in soils before depositing in lake sediments in high-latitude regions (Oswald 2002).

For Moose Lake, multiple approaches were used to calculate sample ages according to sediment character-istics (Fig. 2). A linear regression model fitted through 6834, 6559 and 6493 $\text{BP}$ was used to determine the ages of samples > 6493 years old. The ages of 6493 and 6004 $\text{BP}$ were from the top and bottom boundaries of a peat layer, respectively; we used linear interpolation to assign ages throughout this peat layer. A locally weighted regression (loess) function was used to assign ages for samples < 6004 years old.

For Moose Lake, multiple approaches were used to calculate sample ages according to sediment character-istics (Fig. 2). A linear regression model fitted through 6834, 6559 and 6493 $\text{BP}$ was used to determine the ages of samples > 6493 years old. The ages of 6493 and 6004 $\text{BP}$ were from the top and bottom boundaries of a peat layer, respectively; we used linear interpolation to assign ages throughout this peat layer. A locally weighted regression (loess) function was used to assign ages for samples < 6004 years old.

At Chokasna Lake, a loess function was used to assign sample ages (Fig. 2). The calibrated ages of 5162 $\text{BP}$ at 144 cm and 5166 $\text{BP}$ at 184 cm are statistically the same. We assumed that the date of 5162 $\text{BP}$ on a large wood fragment was too old because of the greater tendency for wood than for needles to have long residence times in soils before depositing in lake sediments in high-latitude regions (Oswald 2002). In addition, there was

<table>
<thead>
<tr>
<th>Laboratory number</th>
<th>Depth (cm)</th>
<th>$^{14}$C age</th>
<th>Calibrated age (2 Sigma) $\text{BP}$</th>
<th>Material dated</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moose Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CAMS 80803</td>
<td>55–56</td>
<td>1730 ± 40</td>
<td>1636 (1534–1722)</td>
<td>Wood</td>
</tr>
<tr>
<td>CAMS 84950</td>
<td>110–110</td>
<td>3485 ± 40</td>
<td>3756 (3640–3854)</td>
<td>Needle</td>
</tr>
<tr>
<td>CAMS 80799</td>
<td>130–131</td>
<td>4005 ± 45</td>
<td>4478 (4301–4779)</td>
<td>Wood</td>
</tr>
<tr>
<td>CAMS 80800</td>
<td>168–169</td>
<td>5250 ± 40</td>
<td>6004 (5924–6170)</td>
<td>Wood</td>
</tr>
<tr>
<td>CAMS 69530</td>
<td>236–238</td>
<td>5710 ± 80</td>
<td>6493 (6407–6634)</td>
<td>Needle</td>
</tr>
<tr>
<td>CAMS 80802</td>
<td>279–280</td>
<td>6005 ± 40</td>
<td>6834 (6728–6914)</td>
<td>Needle</td>
</tr>
<tr>
<td>CAMS 80801</td>
<td>310–311</td>
<td>5760 ± 40</td>
<td>6559 (6448–6644)</td>
<td>Needle</td>
</tr>
<tr>
<td>Chokasna Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CAMS 81962</td>
<td>80–81</td>
<td>1265 ± 40</td>
<td>1205 (1075–1284)</td>
<td>Needle</td>
</tr>
<tr>
<td>CAMS 84951</td>
<td>117–118</td>
<td>1825 ± 40</td>
<td>1760 (1500–2017)</td>
<td>Wood</td>
</tr>
<tr>
<td>CAMS 81963</td>
<td>144–145</td>
<td>4515 ± 40</td>
<td>5162 (5001–5309)</td>
<td>Wood</td>
</tr>
<tr>
<td>CAMS 84952</td>
<td>184–185</td>
<td>4530 ± 320</td>
<td>5166 (4318–5919)</td>
<td>Needle</td>
</tr>
<tr>
<td>CAMS 81964</td>
<td>230–231</td>
<td>6335 ± 40</td>
<td>7268 (7102–7414)</td>
<td>Wood</td>
</tr>
</tbody>
</table>
no lithological indication of a sediment hiatus between 144 cm and 177 cm of the core to suggest that the calibrated age of 5162 bp might be correct. This date was excluded from the calculation of the Chokasna Lake age model. The large uncertainty of the age at 184 cm is related to the low amount of carbon from a small Picea-needle fragment available for 14C analysis.

Percentage organic matter (OM) was determined using loss-on-ignition method. Subsamples were removed every 5 cm of the sediment cores from both lakes, dried at 100 °C, and ashed at 550 °C.

MACROFOSSIL, POLLEN AND CHARCOAL ANALYSES

The sediment core from Moose Lake was sampled at every 10 cm for macrofossils and then at 1 cm where abundant macrofossils were found during the initial sampling. To concentrate macrofossils, core sections of 1-cm depth interval were washed through a 125-μm mesh sieve. Macrofossils were identified under a 10× dissecting scope. Mollusc identification followed Clarke (1981). Macrofossil results were recorded as presence or absence; no quantification was made.

Sediment samples of 1–2 cm³ were removed at every 3–10 cm for pollen preparation by the methods of Faegri et al. (1989) and Cwynar et al. (1979). At least 300 pollen grains were counted per level. Eighty to 100 Picea pollen grains per level were separated into P. glauca and P. mariana types based on saccus shape, attachment, reticulum and saccus area/corpus area (Hansen & Engstrom 1985). The percentage of each species obtained was applied to the total number of Picea grains counted to estimate the proportions of P. glauca and P. mariana.

Charcoal quantification followed the methods of Clark & Hussey (1996). Sediment samples of 1–3 cm³ were taken at contiguous 1-cm intervals. The samples were treated with 10% KOH and sieved through a 180-μm mesh screen. The residue was dispersed across a Petri dish for visual identification of charcoal particles under a 20× stereoscopy. Particles that were black and opaque with angular edges were assumed to be charcoal (Clark 1988). The surface area of each charcoal particle was measured with a CoolSnap digital camera and Metaview image-analysis software. Charcoal accumulation rates (CHAR) are expressed in the unit of mm² cm⁻² year⁻¹.

ESTIMATION OF FIRE RETURN INTERVALS

Sedimentary charcoal peaks primarily represent fires that burned the area adjacent to a lake within the previous 0–5 years. However, charcoal continues to accumulate during non-fire years. This charcoal comes from distant fires and from re-deposition within the basin (Clark & Royall 1996; Whitlock & Millsapough 1996). To identify individual fire events in the CHAR record, we separated charcoal series into ‘peaks’ (i.e. charcoal deposited during local fires) and ‘background’ accumulation (i.e. charcoal deposited during non-fire years) for each sediment record (Clark & Royall 1996). The background charcoal component was estimated using a kernel smoother (e.g. Silverman 1986) with a bandwidth of c. 100 years, which best represented the data. The background accumulation was subtracted from the raw charcoal series to isolate positive residual peaks. Fires were estimated from the positive residual peaks by a method similar to Clark et al. (1996), where a sensitivity analysis was used to select a threshold value of CHAR, above which the residual charcoal peaks were distinctly different from charcoal deposition during non-fire years. The threshold value was taken as the CHAR where the mean interval between peaks in years was least sensitive to changes in threshold itself. Peaks above the threshold value were assumed to be actual fires. Consecutive charcoal peaks were taken to represent the same fire event, the age of which was the average of those for consecutive peaks. The fire return interval was calculated as the time (year) between two adjacent peaks.

RESULTS

MOOSE LAKE

The sediment is clay before 6500 BP and peat between 6500 and 6000 BP (Fig. 3). Silty gyttja occurs from 6000 to 4900 BP, followed by fibrous silts until 3800 BP. The sediment is primarily organic gyttja from 3800 BP to present, with a few silt layers. Percentage OM varies greatly with the maximal and minimal values of c. 60% and < 5% in the peat and clay, respectively (Fig. 3). Bivalves, gastropods, needles and wood are concentrated in two periods, from 6700 to 5800 BP and from 4900 to 4000 BP (Fig. 4). In addition, shells of Gyraulus parvus and Candona candida, Chara oogonia, calcified Chara stems, and abundant mosses are present from 6700 to 5800 BP.

The pollen diagram can be divided into three zones: 6600–6200 BP, 6200–2400 BP, and 2400 BP to present based on visual inspection (Fig. 3). Pollen spectra prior to 6200 BP are dominated by Cyperaceae (c. 10–60%), Picea (12–44%) and Betula (10–30%). From 6200 to 2400 BP, Alnus (c. 30–60%) and Betula (c. 10–35%) are the dominant pollen taxa, while P. mariana and P. glauca pollen percentages remain low (c. 10% each). After 2400 BP, P. mariana pollen increases to 28% at 1800 BP and then declines to < 20% for the remainder of the record. Alnus pollen declines from 2400 to 1850 BP but then increases to c. 60% at present.

Charcoal accumulation rates are highest (> 1 mm² cm⁻² year⁻¹) from 6550 to 6000 BP when CHAR values decline sharply to < 0.15 mm² cm⁻² year⁻¹ (Fig. 3). At least two fires burned the sediment surface as evidenced by the abundance of charred moss fragments at 6450 and 6500 BP. However, the lack of clear separation between peaks prevented the calculation of MFI during this period. The periods from 5800 to 5000 BP and after...
3800 BP contain charcoal peaks above 0.018 mm² cm⁻² year⁻¹ that are distinct from the background values based on the sensitivity analysis and distribution of CHAR (Fig. 5). Between 5000 and 3800 BP, CHAR values are < 0.01 mm² cm⁻² year⁻¹, except for the peak of 0.037 mm² cm⁻² year⁻¹ at 4450 BP.

CHOKASNA LAKE

The sediment is grey clay before 7000 BP, followed by peat with varying amounts of mosses, organic debris and clay from 7000 to 3600 BP (Fig. 3). The peat is highly decomposed from 7000 to 6100 BP, better preserved from 6100 to 5050 BP, and again highly decomposed but intermixed with clay from 5050 to 3600 BP. A core gap occurs from 3600 to 3150 BP. The peat is replaced by brown fibrous silts (i.e. rootlets) from 3150 to 2350 BP, followed by a fibrous gyttja from 2350 to 1750 BP. After 1750 BP, the sediment is primarily organic gyttja. Percentage OM is high (50–70%) before 4800 BP and low (c. 20%) between 4800 and 2550 BP with the exception of a peak (c. 50%) at 4200 BP. After 2550 BP, OM gradually rises to a maximum of 75% at 1425 BP and then declines to 15–25% after 1000 BP.

The pollen diagram can be divided into three zones based on visual inspection: 7000–5100 BP, 5100–2000 BP, and 2000–500 BP (Fig. 5). Pollen spectra prior to 5100 BP are dominated by Cyperaceae (c. 8–42%), P. glauca (c. 40–60%), and P. mariana (c. 20–30%). At 5100 BP, Alnus and Betula pollen percentages increase to c. 10%. From 2000 to 500 BP P. mariana becomes a dominant pollen type (up to 50%) while P. glauca pollen decreases...
Climatic history

The absence of lake sediment older than 7200 BP at Moose and Chokasna lakes probably reflects that the region was extremely dry and that the basins contained no water prior to that time. This interpretation is consistent with palaeoclimatic records from other lakes in interior Alaska, where lake levels rose markedly around 6800 BP at Birch Lake based on stratigraphic and seismic profiles (Abbott et al. 2000) and around 7500 BP at Farewell Lake based on trace-element geochemistry (Hu et al. 1998). Similarly, in the central Brooks Range, a number of lakes did not begin to accumulate lake sediments until the mid-Holocene, when increased effective moisture probably resulted in the rise of lake levels (F.S. Hu, unpublished data).

At both Moose and Chokasna lakes, alternating sediment types in conjunction with macrofossils suggest pronounced variations in water depth after 7000 BP (Figs 3 and 4). These variations probably reflect changes in effective moisture as a function of precipitation and evapotranspiration, because neither lake has inlets or outlets (Digerfeldt 1986). We use this assumption to...
infer variations in the regional moisture regime over the past 7000 years.

Prior to 3800 BP, effective moisture was likely to be lower than at present with substantial fluctuations. The peat deposits from 6500 to 5800 BP at Moose Lake and from 6800 to 3600 BP at Chokasna Lake indicate a shallow-water, wetland environment (Fig. 3). In addition, the presence of shallow-water species *Gyrula parvus*, *Chara*, and *Candona candida* (Delorme 1970; Clarke 1981), and the abundance of fibrous rootlets, suggest low lake levels at Moose Lake from 4900 to 4000 BP. At this site, a period of higher effective moisture interrupted the overall dry conditions, as evidenced by silty gyttja and the disappearance of shallow-water organisms between 5800 and 4900 BP (Figs 3 and 4). For the same period at Chokasna Lake, peat continued to accumulate, but the peat constituents are considerably less decomposed and intermixed with clay, probably reflecting decreased decomposition under prolonged seasonal inundation and with clay influx perhaps from landscape processes such as freeze-thaw cycles. The difference in sediment composition between the two sites during this period is consistent with the fact that Chokasna Lake is shallower than Moose Lake.

Effective moisture increased and lake levels rose between about 3800 BP and 1600 BP, as inferred from changes in a variety of indicators. At Moose Lake, shallow-water indicators (e.g. gastropods, macrophytes and fine roots) disappeared, and organic gyttja replaced fibrous silts around 3800 BP (Figs 3 and 4), and at Chokasna Lake fibrous silts replaced degraded peat around 3600 BP. At Chokasna Lake, shifts from fibrous silts to fibrous gyttja and then to organic gyttja suggest further water-level rises around 2550 BP and 1600 BP (Fig. 3). A deeper lake would have reduced detrital-mineral inputs to the centre of the lake, resulting in increased percentage OM at these times (Fig. 3). Lake levels at Moose Lake were probably sufficiently high after 3800 BP that any further rises would not change the sediment type.

Our data do not allow us to estimate temperature variations that may have accompanied changes in effective moisture. However, glacio-geological evidence of neoglacialation suggests that the late-Holocene was generally cooler than the mid-Holocene. For example, alpine glaciers advanced in various areas of Alaska, including the Wrangell Mountains, the westernmost St Elias Mountains and the northern Alaska Range, about 3000–2000 BP (Calkin 1988; Wiles et al. 2002). The late-Holocene rises in lake level and effective moisture, as inferred from our data from Moose and Chokasna lakes, probably resulted from increased precipitation as well as decreased evaporation because of climatic cooling.

**Vegetational history**

The pollen records from Moose and Chokasna lakes suggest that closed boreal forests with *P. glauca*, *P. mariana*, and *Betula* as the dominant taxa existed throughout the past 7000 years (Fig. 3). At present, *B. papyrifera* probably formed mixed stands with *P. glauca* on uplands near Moose Lake, especially before 2400 BP, as suggested by the abundance of *Betula* and *P. glauca* in the pollen record. The high abundance of *Alnus* pollen probably reflects that *A. crispa* dominated the shrub layer of the forests near Moose Lake. Compared with that of Moose Lake, the pollen diagram from Chokasna Lake has greater amounts of *P. glauca* and *P. mariana* associated with lower abundances of *Betula* and *Alnus*, suggesting that *Picea* was more common and hardwood species less common in the forests near Chokasna Lake. In addition, Cyperaceae pollen percentages are much higher at Chokasna Lake than at Moose Lake, probably reflecting the prominence of wetlands at Chokasna Lake before 2400 BP.

*Picea mariana* replaced *P. glauca* as the dominant *Picea* species around 2400 BP at Moose Lake and around 2000 BP at Chokasna Lake (Fig. 3). This shift appeared more prominent at Chokasna Lake than at Moose Lake. It probably reflected the establishment of the modern boreal-forest mosaic near Chokasna Lake, with *P. mariana* communities on poorly drained lowland sites and *P. glauca* communities on well-drained upland sites. The subtle rise of *P. mariana* pollen at Moose Lake around 2400 BP could probably be attributed to an increase in the deposition of regional *P. mariana* pollen. The larger surface area (21.5 ha) of Moose Lake than that of Chokasna Lake (11 ha) would have permitted greater regional pollen deposition. A modest expansion of the local populations of *P. mariana* may have also occurred. However, *P. mariana* was probably never substantially more abundant than *P. glauca* near Moose Lake; *P. glauca* and *B. papyrifera* continue to be the dominant forest species today at this site.

The extensive flat, low-lying terrain that surrounds Chokasna Lake probably led to more extensive soil paludification as effective moisture increased during the late-Holocene. These soils would have favoured the development of dense *P. mariana* stands (Viereck et al. 1992) that continue to be present today. Although an increase in effective moisture probably occurred around 3800 BP, as discussed above, this climatic change was probably insufficient to impede drainage extensively and cause paludification until a further moisture increase about 2000 BP. In contrast to Chokasna Lake, Moose Lake is surrounded by rolling hills and upland terraces that today maintain overall dry and warm soil conditions. On these soils, a mosaic of *P. glauca*, *Betula* and *Alnus* dominated stands prevailed over much of the watershed. Thus, despite the fact that the two sites probably experienced similar climatic changes, their vegetational histories differ because of differences in topographic features and associated soil conditions.

No previously published pollen diagram exists from the Copper River region, preventing us from verifying our pollen data with other results. However, numerous palynological studies in other regions of Alaska have documented the establishment of closed coniferous...
Climatic control of boreal forest fire by the early to mid-Holocene in central and eastern Alaska (Anderson et al. 1994). The existence of boreal forests at our sites throughout the past 7000 years is broadly consistent with this pattern. The shift from *P. glauca* to *P. mariana* as a dominant species in our study area around 2000 BP occurred much later than a similar vegetational change in interior Alaska during the mid-Holocene (Brubaker et al. 2001). However, this discrepancy is not unexpected, as previous studies have shown that the timing of this important vegetational shift, which marked the establishment of modern boreal forests at many sites, differed greatly in different regions of Alaska (e.g. Anderson et al. 1994; Hu et al. 1996; Brubaker et al. 2001; Lynch et al. 2002).

**Fire history**

Our charcoal profiles from Moose and Chokasna lakes are likely to represent local charcoal deposition. Recent studies using experimental burns have shown that 99% of airborne, large (> 180 µm) charcoal particles are deposited within 100 m from the burn edge (Clark et al. 1998; Lynch 2001). Sensitivity analyses of our CHAR data further suggest that peaks above 0.018 mm² cm⁻² year⁻¹ and 0.085 mm² cm⁻² year⁻¹ at Moose and Chokasna lakes, respectively, best represent local fires (Fig. 5). All CHAR peaks above these threshold values in the records are at least 1.7 times greater than the background values. This pattern agrees with modern studies that show that historic fires in Alaska and Canada produced charcoal peaks in lakes that were on average 1.5 times greater than background (Lynch 2001; Lynch et al. 2002). In addition, the CHAR value of 0.023 mm² cm⁻² year⁻¹ from the 1915 fire that burned at Moose Lake is clearly identifiable as a local fire. Thus, the CHAR peaks above the threshold values appear to reliably record local fires.

The estimated MFI values of 190 ± 20 years from 5800 to 5000 BP and 210 ± 80 years from 3800 BP to present at Moose Lake are near the upper limit of MFI values for modern *P. glauca* stands (60–210 years) in Alaska (Nienstaedt & Zasada 1990). The MFI of 220 ± 60 years from 5600 to 4550 BP at Chokasna Lake is very similar. The MFI of 150 ± 80 years after 2000 BP for Chokasna Lake reaches the upper limit of the MFI range (50–150 years) for modern *P. mariana*-dominated communities (Viereck 1973; Heinselman 1981). For the remainder of the records at both sites, MFI values are much greater (> 500 years) than those for modern *Picea*-dominated forests. Because our charcoal sampling resolution (i.e. 1 cm) exceeds 40 years per sample, our CHAR peaks may represent multiple fires, which would have resulted in overestimates of the actual MFI values. However, this potential problem is not restricted to periods with high estimated MFI values, and it is unlikely to be responsible for the marked MFI differences between dry and wet climatic conditions.

At both Moose and Chokasna lakes, MFI differed greatly between dry and wet climatic conditions. Higher fire frequencies occurred during wetter periods. For example, fires were more common from 5800 to 5000 and from 3800 BP to present at Moose Lake when the regional climate was wetter than during the remainder of the record (Fig. 6). Similarly at Chokasna Lake, fires were more important from 5600 to 4600 and from 3150 BP to present under wetter conditions than during the remainder of the record. Despite this broad similarity between the two records, some differences exist (Fig. 6).

---

**Fig. 6** Major pollen taxa (*Alnus, Betula, P. glauca* and *P. marina*) and fire return intervals. Horizontal lines across plots indicate transitions between interpreted dry and wet climate stages.
In particular, fires occurred more frequently at Chokasna Lake (MFI 150 ± 80 years) than at Moose Lake (MFI 210 ± 80 years) during the past 2000 years. In contrast, the fire frequency was lower at Chokasna Lake (MFI 295 ± 20 years) than at Moose Lake (MFI 210 ± 80 years) between 3150 and 2000 yr.

A potentially important factor that may have compromised the reliability of our MFI estimates is variation in charcoal deposition resulting from lake-level fluctuations. This possibility is supported by the fact that CHAR varies with sediment type at both Moose and Chokasna lakes. Macroscopic charcoal particles should behave similarly to other plant macrofossils (e.g. needles) in response to changes in the sedimentological regime. In addition, larger charcoal particles should have been preferentially deposited in the sediment when lake levels were lower and lake diameters smaller. However, our results show that CHAR is lower when the presence of other macrofossils indicates shallower water (e.g. from 5000 to 3800 m) and higher when their absence indicates deeper water (e.g. from 3800 m to present at Moose Lake).

This observation strongly argues against the possibility that CHAR variation was driven by lake-level fluctuations. An exception to this discussion is the high CHAR values in the peat of 6500–6000 B.P. at Moose Lake, which probably resulted from burning of the peatland surface.

Other potential factors that could have affected our MFI estimates are variation in the amount of surface runoff from the watershed to the lake, sediment mixing, and re-deposition of charcoal within the lakes. For example, the higher charcoal accumulation in the Bor Island Lake sediment than in the air during the Bor Island experimental fire (Clark et al. 1998) demonstrated the importance of surface flow for charcoal deposition in lakes. An increased amount of surface run-off during wetter climatic conditions thus could have increased the influx of charcoal particles that resided in watershed soils to the lake, resulting in elevated CHAR values during non-fire periods. Such a process should have affected other terrestrial macrofossils as well. However, at our sites, Picea needles and wood fragments are less abundant in the sediments that were presumably deposited during the periods of wetter conditions (e.g. after 3800 B.P. at Moose Lake).

In addition, surface flow is reduced following most fires and fine-fuel accumulation. These factors should have shortened the fire season and lowered the probability of occurrence. An increase in the fuel load on the landscape can enhance fire occurrence, particularly if coincident with an increase in flammable species (Heinselman 1973). For example, P. mariana tends to grow in densely packed stands, making them prone to burning (Viereck 1973). At Dune Lake in interior Alaska, an increase in P. mariana at 5500 B.P. resulted in more frequent fires (Lynch et al. 2002). However, this explanation seems...
unlikely at Moose and Chokasna lakes prior to 2000 bp because *P. glauca* forests dominated during both dry and wet periods (Fig. 5). A typical *P. glauca* forest contains enough coarse fuels to support a MFI of 100 years. Thus, fuels before 3800 bp, when effective moisture was low, were probably adequate for frequent burns at our sites.

In the boreal forests of Alaska, areas with frequent cloud-to-ground lightning strikes also have the highest incidence of fires (Gabrel & Tande 1983; Kasischke et al. 2002; Dissing & Verbyla 2003). An increase in ignition may account for lower MFI during periods of wetter climatic conditions. Lighting, while uncommon in the Copper and Chitina River valleys (Dissing & Verbyla 2003; Reap 1991), occurs frequently enough during the fire season (May–October) to provide an occasional ignition. From 1986 to 1999 the Copper and Chitina River valleys had a lightning density of $5 \times 10^{-3}$ strikes km$^{-2}$ during the fire season (Dissing & Verbyla 2003). The occurrence of fires in Alaska is closely linked to large-scale atmospheric circulation patterns (Hess et al. 2001; Kasischke et al. 2002), which controlled summer thunderstorms and lightning patterns. Using satellite data to infer the spatial pattern of lightning strikes, studies (Biswas & Jayaweera 1976; Reap 1991) have demonstrated the importance of large-scale frontal systems and instability for the formation of thunderstorms over Alaska. Thus, a change in large-scale climatic patterns may have resulted in more lightning strikes to increase fire ignitions and more thunderstorms to increase effective moisture.

Dry surface fuels are also necessary for fire ignition, and increased moisture may reduce biomass flammability under wetter climatic conditions. Surface fuel moisture is related to summer precipitation and freeze-thaw cycles of permafrost soils in Alaska (Kasischke et al. 2000). Summer precipitation, in turn, is negatively correlated with the patterns of fire in Alaska (Hess et al. 2001; Kasischke et al. 2002). However, an overall wet climatic regime at centennial to millennial time-scales does not preclude variable weather conditions, as well as a particularly dry fire season where fuel moisture would be low enough to allow ignition. For example, during the dry summer of 1915, fires burned large areas of the Chitina River valley (Jennifer Allen, personal communication; Lutz 1956). Thus a combination of increased ignition and interannual moisture variability may have provided the conditions for more frequent fires under the generally wetter conditions at our sites. Unfortunately, interannual moisture variation cannot be inferred from our palaeoclimatic reconstruction. Additional studies that reconstruct seasonal precipitation (Nesje et al. 2002) at high-temporal resolution are needed to improve our understanding of fire dynamics in relation to long-term climatic patterns.

The close proximity (20 km apart) of Chokasna and Moose lakes eliminates ignition frequency as the cause of their different MFI values after 2000 bp. Variations in local factors, such as soil moisture, topography and stand-level vegetation, probably exerted secondary controls over fire occurrence. Specifically, the extensive lowlands surrounding Chokasna Lake probably facilitated soil paludification when effective moisture increased in the region around 2000 bp. This soil change would have favoured the expansion of *P. mariana*, a fire-prone species (Viereck 1973), resulting in a decrease of MFI, as suggested by our pollen and charcoal data. In contrast, the rolling uplands surrounding Moose Lake were less susceptible to soil paludification. Fires occurred less frequently at this site than at Chokasna Lake, because the mixed forests dominated by *P. glauca* and deciduous species (*Betula* and *Alnus*) are less flammable (Brown & Davis 1973). These results support the idea that differences in vegetational composition and fuel type associated with local factors can result in diverge fire regimes under the same regional climate.

Our results, along with those of Carcaillet et al. (2001) and Lynch et al. (2002), demonstrate that warmer and drier climatic conditions do not necessarily induce greater fire importance. These results contradict the current understanding of modern fire–climate relationships. It is also inconsistent with model predictions that a drier and warmer climate, as a result of greenhouse warming, will lead to increased fire activity in boreal systems (Flannigan et al. 2001). Regions such as south-central Alaska may experience a reduction in fire activity with climatic warming if altered atmospheric circulation results in a decrease in lightning frequency. Among the impacts of lower fire importance on the regional ecosystems are that late-successional *Picea* forests would be favoured and that permafrost melting would be retarded, which could potentially increase or maintain terrestrial carbon storage. The effects of interactions among climate, topography and fuel type on fire frequency could also have an impact on the magnitude and direction of future changes in communities (species distribution, migration and extinction) and ecosystem processes (soil and nutrient dynamics) (Weber & Flannigan 1997). For example, a mixed broadleaf and conifer stand on well-drained uplands may be more resistant to shifts in critical fire weather and more likely to retain stored carbon in the soils. Therefore, forecasting future boreal ecosystem changes requires knowledge of the links among processes operating at both long temporal scales, such as precipitation–ignition relationships and fuel accumulation, and at fine spatial scales, such as topography, soil types and forest composition.

**Acknowledgements**

This work was supported by National Science Foundation grant OPP 01–08702 and a Packard Fellowship in Science and Engineering to FSH. We thank Willy Tinner and Brandon Curry for coring assistance, Tom Brown for $^{14}$C dating, and Andrea Hui for charcoal counting. Comments from Paul Henne, Ben Clegg, Dan Gavin, Peter Moore, Dave Nelson, Jian Tian, Willy Tinner and two anonymous reviewers improved the manuscript.
References


Climatic control of boreal forest fire


