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A multi-proxy holocene record of environmental change from the sediments of Skinny Lake, Iskut region, northern British Columbia, Canada

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Abstract

A stratigraphic record from a lake in the Central Plateau Region of northern British Columbia reveals changes in environment and inferred climate during the Holocene. Upon deglaciation (ca. 11500 BP), Skinny Lake became an embayment of an ice-dammed lake. High clastic sedimentation rates, an unstable landscape, and cool, possibly wet conditions likely persisted until the early Holocene (ca. 9000 BP). From ca. 9000–8300 BP declining lake levels coupled with warm and dry conditions resulted in the formation of a prominent marl bed. A colonizing shrub and herb assemblage persisted from 9000 BP until about 8300 BP when it was replaced by a spruce (*Picea*) and subalpine fir (*Abies lasiocarpa*) forest under slightly cooler and moister conditions. The middle Holocene was warmer-than-present, however, decreasing temperature and increasing precipitation trends characterize the period from ca. 6000 BP–3000 BP. The transition to modern climate at 3000 BP is evident primarily in the lithostratigraphic record and corresponds with the initiation of the Tiedemann glacial advance (ca. 3300 BP) in the south-coastal mountains of British Columbia. A significant change in fossil pollen occurs at ca. 2400 BP and is characterised by an increase in pine pollen accompanied by decreases in alder (*Alnus*), spruce and fir. This also coincides with an increase in west-sourced exotic western hemlock (*Tsuga heterophylla*) and cedar type (Cupressaceae) pollen possibly transported by regional changes in air mass circulation patterns associated with Aleutian Low dynamics. This study demonstrates that both lithostratigraphic and biotic proxies are helpful in reconstructing the timing and nature of climate change and that each may have varying sensitivities to a particular type of change.

Introduction

To date relatively few studies in the North and Central Plateau and Mountain regions of British Columbia (Meidinger and Pojar 1991) have focused on reconstructing Holocene climatic change (Hebda (1995); Figure 1). The central portion of northern British Columbia is particularly devoid of long records of vegetational and climatic history. Published studies in northwestern British Columbia (Miller and Anderson (1974), Cwynar (1988, 1993), Spooner et al. (1997), Mazzucchi (2000); Figure 1) indicate temporal variations in moisture and temperature states and trends. Cwynar (1993), Spooner et al. (1997), Mazzucchi (2000) suggest that late Holocene climate change was accompanied by air-mass circulation changes with substantial environmental effects. Studies to the south (Banner et al. (1983), Gottesfeld et al. (1991); Figure

1) and on the eastern margin of British Columbia (White and Mathewes (1982), MacDonald (1984, 1987), MacDonald and Cwynar (1985); Figure 1) also document late Holocene climatic change but with no correlation to changes in regional atmospheric circulation. In this study we investigate paleoclimatic change in the North Plateau region of northern British Columbia using both sedimentological and fossil pollen records. It was anticipated that a paleoenvironmental record from this site would aid in the correlation between northwestern coastal records, those from the eastern Rocky Mountains margin, and those to the south.

Study site

Skinny Lake (informal name) is 1 km long, with a maximum measured depth of 7 m and is located at an elevation of 910 m (3000 ft.) above sea level (Figure 1). The study site is near the centre of the Northern and Central Plateaus and Mountains physiographic

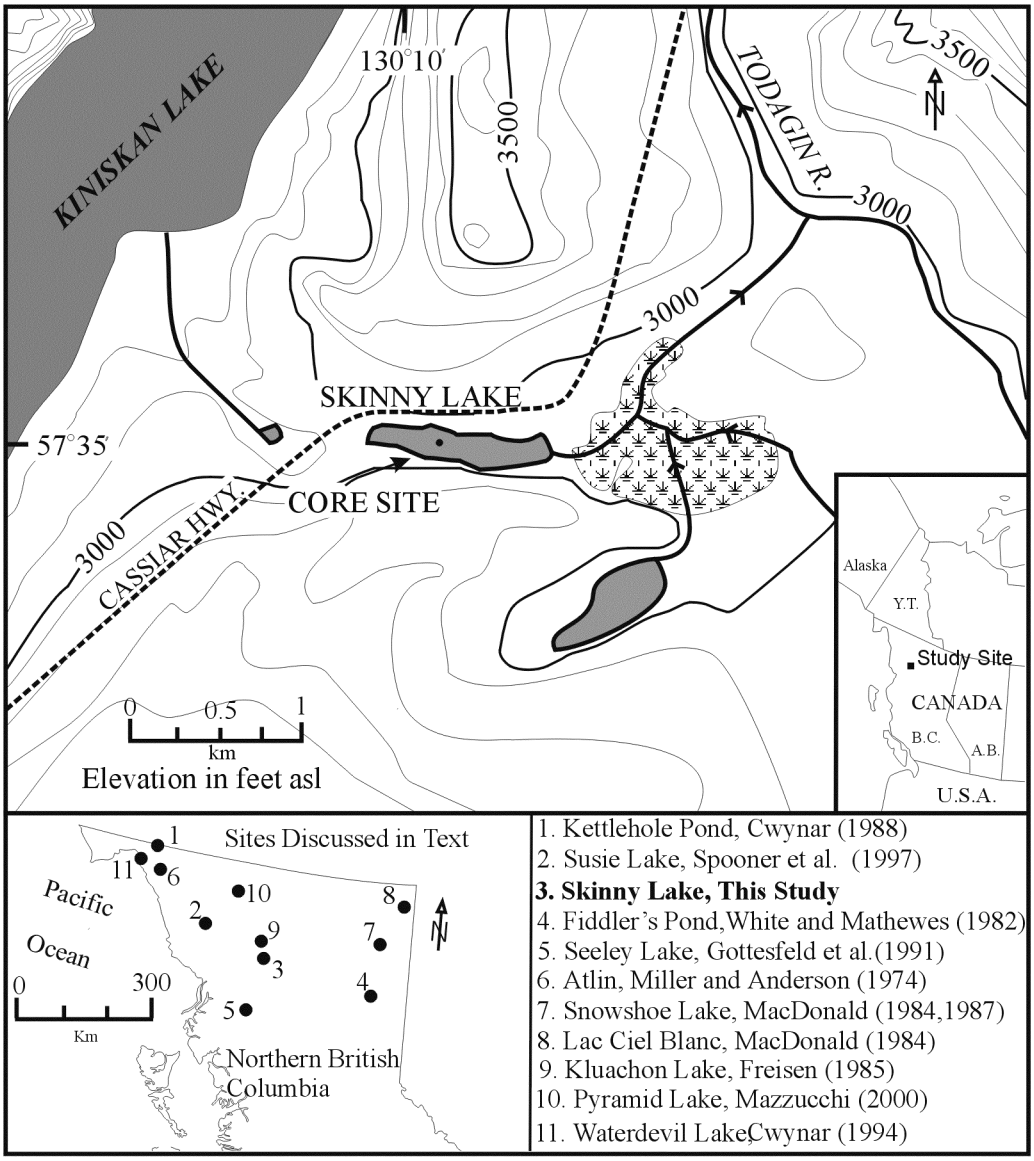

Figure 1. Location of Skinny Lake and other sites referred to in the text. Skinny Lake is a small, shallow, organic lake with minimum inflow and outflow.

region (Meidinger and Pojar 1991) and is bounded on the west by the Klastline Plateau (1800 m) and on the east by the Spatsizi Plateau (2000 m). Skinny Lake is located at a local expansion in the Todagin River valley just east of the Iskut River valley (Kluachon Lake, Figure 1), a major north-south trending physiographic feature in the region. The site was chosen because the lake is easily accessed and its small size and minimal inflow suggest low rates of clastic sedimentation. Small shallow organic lakes may also be particularly sensitive to changes in air temperature (Spooner 1998). A short duration (ca. 2000 yr.) paleoecological record (Friesen 1985) obtained from Kluachon Lake, 20 km north of Skinny Lake (Figure 1) serves as a useful reference.

Bedrock in the region is dominated by a complex suite of Mesozoic and Cenozoic volcanic and sedimentary rocks that include limestone and coal (Souther 1992). A dormant volcanic complex (Mt. Edziza) is located about 20 km to the west of the study site and is thought to have erupted at least 30 times during the Holocene (Fladmark 1985; Souther 1992). Relatively little is known about Quaternary glaciation in the Iskut River valley. Ryder and Maynard (1991) indicate that at the climax of Late Wisconsinan (Fraser) glaciation a major ice divide existed close to the study site. Deglaciation of the plateau regions may have preceded that of valleys, which were occupied by downwasting stagnant ice (Gabrielse and Souther 1962; Clague 1987; Ryder and Maynard 1991). The timing of deglaciation is uncertain (Kerr 1948; Ryder and Maynard 1991). Available basal dates suggest that large valleys were ice free by 9000 BP (uncalibrated age; Spooner et al. (1997), Spooner and Osborn (2000)) and alpine sites may have been ice free as early as 9700 BP (Mazzucchi 2000) Several authors have carried out ethnoarchaeological studies (Emmons 1911; Helm 1956; Sheppard 1983; Albright 1984) and geobotanical studies (Anderson 1970; Cathey 1974; Krajina et al. 1982) in the region.

The study site is located within the Northern Boreal Mountains Ecoprovince and the Dry Cool Boreal White and Black Spruce subzone of the BWBS zone (BWBSdk biogeoclimatic ecosystems classification; Meidinger and Pojar (1991)). Precipitation is evenly distributed throughout the year with summertime heating resulting in convective showers; however, rain-shadow effects can cause some areas to be very dry (Demarchi 1996). The mean annual temperature for this zone is –2.9°C to 2°C and monthly averages remain below 0°C for 5–7 months of the year (Meidinger and Pojar 1991). Outbreaks of arctic air occur frequently in the winter and spring and below freezing temperatures prevail from October to April with January temperatures averaging −20°C (Pojar 1985). Lakes commonly freeze in late October and are often not ice free until June. Temperatures in the summer (June to August) average 10°C (Pojar 1985).

White spruce (*Picea glauca*) is common in the BWBS zone as is lodgepole pine (*Pinus contorta*) a species common on well-drained terraces as well as recently burned areas. Black spruce (*Picea mariana*) dominates poorly drained soils and trembling aspen (*Populus tremuloides*) occurs on south facing slopes (Utzig and Walmsley 1982). Other common species include alder (*Alnus*), birch (*Betula*), and willow (*Salix*). Both the Spruce-Willow-Birch BEC zone and the Engelmann Spruce-Subalpine Fir BEC zone occur above the BWBS in the study region (Meidinger and Pojar 1991).

Methods

Skinny Lake was cored through the ice using a portable percussion coring system (Gilbert and Glew 1985; Reasoner 1993) to obtain three 7.5 cm diameter cores. Cores were frozen and transported to Calgary, Alberta where they were split with a high-speed diamond rock saw, photographed, and examined. All analyses were carried out on core SK99-2.

Pollen analysis was conducted on 1 cm^3 sediment sub-samples from the center of the core, following the procedure of Faegri and Iverson (1975). Pollen concentration was facilitated by the addition of *Lycopodium* marker grains (Stockmarr 1971). Palynomorphs were identified under 400x magnification using pollen reference slides and various reference texts (Bassett et al. 1978; Faegri and Iverson 1975; McAndrews et al. 1973). Pine pollen was not separated into the Haploxylon and Diploxylon subgenera as the Haploxylon/Diploxylon ratios were consistently less than 0.1. *Picea glauca* and *Picea mariana* were separated using grain size differentiation techniques described in Birks and Peglar (1980), Cushing (1961) as these two species have different habitat tolerances. Pollen data were plotted using Canplot (Campbell and McAndrews 1992) and zonation and cluster analysis was accomplished using CONISS (Grimm 1987). Terrestrial pollen only were plotted The pollen count for each taxon is reported as percentages of the pollen sum.

Carbon/nitrogen (C/N) data were collected at the Isotope Science Laboratory, University of Calgary, for the organic portion of the core (230 cm to top of core) using a Finnigan Mat 'TRACERMAT' which comprises a Carlo Erba NA 1500 elemental analyzer interfaced to a magnetic-sector mass analyser. These data were collected in order to determine the origin of sedimentary organic matter. Meyers (1994) has shown that the sedimentary C/N ratio reflects the sources of organic matter and undergoes little change once deposited. Phytoplankton have low C/N ratios (4–10), whereas vascular land plants have C/N ratios of 20 or greater (Meyers and Lallier-Vergés 1999). Allochthonous lake sediments generally produce relatively high C/N ratios, whereas low C/N ratios are usually associated with lake eutrophication. (Huttunen and Meriläinen 1983).

Sub-samples of sediment were analysed for magnetic susceptibility (MS) using a Saphire Instruments SI-2B magnetic susceptibility meter. A modified syringe was used to retrieve 1 cm^3 sub-samples from the core at 10 cm intervals. Magnetic susceptibility is a ratio of the induced magnetic field upon a sample to the intensity of the magnetising field and, in general, within highly organic sediment, it indicates fluctuations in the clastic content of the core. Loss on Ignition (LOI) analyses were carried out at 550°C (total organic carbon) for four hours and 1000°C (total inorganic carbon) for two hours following the procedures of Dean (1974), Heiri et al. (2001). LOI was carried out on samples of known volume and cross-sectional area thus facilitating the calculation of sediment influx rates ($g/cm^2/yr$).

Analyses of the coarse mineral grains in lake sediment can assist in the interpretation of the complex relationship between water-column productivity and landscape stability (Spooner 1998; Beierle et al. 2001). The small fetch (max. 1050 m) and moderate depth of Skinny Lake limit the common processes capable of transporting coarse sediment ($>$ 0.5 mm D_0) into the centre of the lake basin to avalanche/debris flow run-out onto ice and, rarely, eolian transport onto ice (Luckman 1975; Cole 1983; Evans 1994; Lewis et al. (in press)). These clasts (ice rafted debris, IRD) melt through the ice in-situ, accelerated by their dark colour (Lewis et al. (in press)). A relative assessment of coarse sediment grain size analysis was accomplished on lacustrine sediment (sediment units 2–7; 295 cm to top of core) by cutting the remainder of the core (after pollen and macrofossil analyses) into 5 cm portions which were impregnated with low viscosity embedding media. Thin-sections were produced and then scanned using a Polaroid SprintScan 35 plus slide scanner. The area of individual grains was determined using the image analysis program Imagetool®. Clast area was then converted to equivalent disk diameter (D_0; Francus (1998)). Grain counts of D_0 values greater than 0.5 mm were recorded. The data serve as a useful indication of relative changes in frequency of coarse grained clasts referred to in the text as ice rafted debris (IRD). Chronological control for the core was established from 5 AMS dates (Table 1).

Table 1. Radiocarbon dates from Skinny Lake sediment core SK99-2, dates are uncalibrated.

Depth bct[1] cm	Age (^{14}C yrs BP)	Lab. No.	Material
30	1939±46	AA-41878	Alder twig
67	3130±60	TO-8544	Alder twig
125	5552±50	AA-38582	Seed
196	8140±50	TO-8545	Spruce cone
270	11433±97	AA-41879	Twig frag. (willow?)

TO = Isotrace Laboratories, Toronto, ON, CND. AA = Arizona AMS Facility, Tucson AZ, USA.

below core top

Results and interpretation

Core lithostratigraphy

Core SK99-2, chosen for detailed study, is longest (330 cm) and exhibits complex stratigraphy which is summarised in Figure 2. The core was visually subdivided into 7 distinct sediment units.

Sediment units

The lowermost unit (Unit 1; 330 cm–295 cm; Figure 2) is a dark grey, well indurated and coarse-grained diamicton with large (>3 cm) angular polymictic clasts; some clasts are striated. Unit 1 is interpreted as a lodgement till and is in sharp contact with a massive dark grey clay layer (Unit 2; 295 cm–275 cm; Figure 2) also with occasional striated cobbles and pebbles. Unit 2 is interpreted to represent initial infill of the

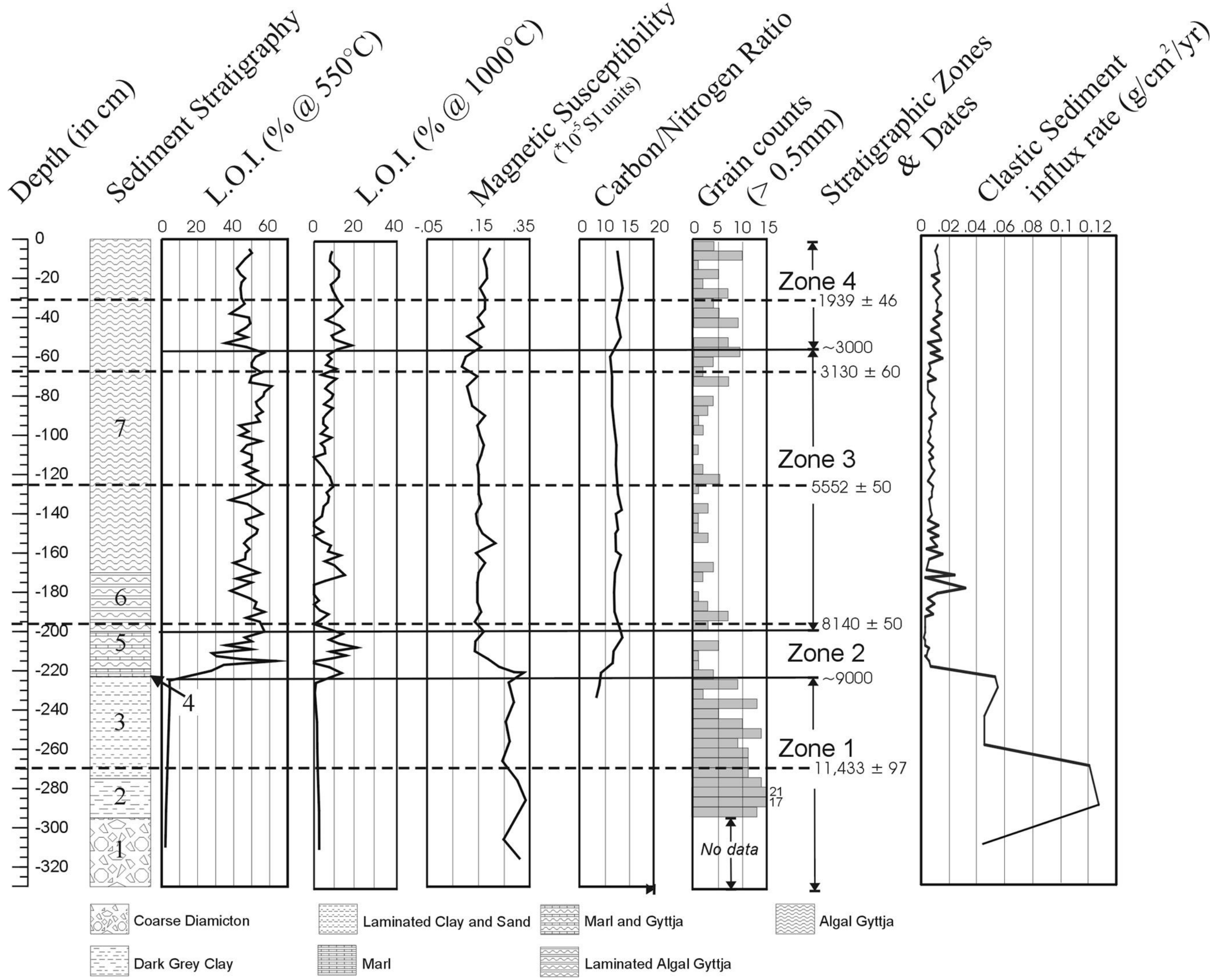

Figure 2. Skinny Lake sediment core lithostatigraphy. Units 1–3 are primarily clastic, have very low pollen concentrations and were deposited during and shortly following regional deglaciation. Units 4–6 are primarily organic and record the development of both aquatic and terrestrial productivity. LOI at 550°C values are uniformly low (< 10%) for units 1–3 but increase and vary for units 4–6 (30–60%). A notable decline in LOI (550°C) and increase in magnetic susceptibility occurs from 60 cm to the top of the core. Concentrations of IRD vary greatly in the core with an increase in upper unit 7 mimicking changes observed in the other stratigraphic proxies.

lake basin upon deglaciation. IRD is common in this unit. Laminated clay and sand (Unit 3; 275 cm–222 cm; Figure 2) rest in sharp contact on the dark grey clay. Contacts between the laminae are abrupt; sand layers fine upwards, are mildly calcareous, and average 0.5–0.7 cm thick. Clay layers are somewhat thicker (1 cm–3 cm) and contain articulated bivalves (*Pisidium sp.*) and IRD. Organic detritus becomes more common towards the top of this unit. A date of 11433 ± 97 BP was obtained from twig fragments at 270 cm (Table 1). The sand layers may represent turbidity current deposits generated as valley slope slumps transformed into lake-marginal debris flows (Spooner and Osborn 2000). The clay layers were probably deposited during periods of quiescence.

Marl dominates unit 4 (222.5–219 cm; Figure 2) which is composed of a few yellow and brown highly calcareous sediment layers with abrupt boundaries. No *Chara sp*. fossils were noted, suggesting a non-biogenic source for the carbonate. The layers are themselves laminated. This unit contains numerous articulated and disarticulated *Pisidium sp*. shells concentrated near the base. The sharply bounded alternating dark brown and light brown layers of unit 5 (219–200 cm; Figure 2) contain shells, especially in the light brown layers. Both units 4 and 5 exhibit increased LOI at 1000°C compared to unit 3. Units 4 and 5 record variations in the marl content of the lake sediment. The marl is likely a consequence of generally warm lake water conditions that resulted in periodic supersaturation with respect to $CaCO_3$. IRD is less common in both of these units.

Unit 6 (200–170 cm; Figure 2) consists of dark brown and strongly laminated algal gyttja. Unit 7 (170 cm to top of core; Figure 2) also consists of dark brown gyttja with occasional light brown marl layers that correspond to increases in LOI at 1000°C. Occasional IRD was noted within these units and become more common towards the top of the core. These two units are interpreted as gyttja deposited as the landscape was locally stabilized and lake productivity increased. During this interval, terrestrial and aquatic flora and fauna became important sediment sources. Coarse sediments in these units are probably IRD transported by debris flow run-out onto lake ice.

Lithostratigraphic zones

Four lithostratigraphic zones were established by visually determining points of maximum change in observed and measured parameters (*see* Figure 2).

Stratigraphic zone 1 (SZ-1, 330–222.5 cm*, units 1,2,3; ca. 11,500–9000 BP)*

SZ-1 consists of a diamicton with striated cobble clasts which are in abrupt contact with massive and laminated fine grained clastic. We interpret this zone to represent till (unit 1) overlain by retreat phase glacial lake sediment (units 2, 3). Glaciolacustrine sediments are common in the region indicating that numerous lakes existed during deglaciation. At this time Skinny Lake may have been an embayment in a much larger proglacial lake that occupied the north-south trending Iskut River Valley (including Kiniskan Lake, Figure 1) located to the west of Skinny Lake (Ryder and Maynard 1991; Hanson and McNaughton 1936). High clastic sedimentation rates, and numerous IRD characterize this zone and indicate an unstable landscape at the site (Figure 2). Two low C/N values (<10, Figure 2) at the top of this zone indicate that the organic sediment that is present has a relatively high aquatic component. However, twig fragments were found within this zone (11433 ± 97 BP; Table 1) indicating that initial terrestrial colonization of the study area had begun soon after deglaciation.

Stratigraphic zone 2 (SZ-2, units 4, 5; 222–203 cm*, ca. 9000–8200 BP)*

A sharp transition exists between SZ-1 and overlying SZ-2 which is predominantly organic sediment characterized by increasing LOI and C/N values and decreasing magnetic susceptibility and grain size counts. This zone records the rapid increase in aquatic productivity coupled with a decrease in clastic sedimentation rates as the lake basin stabilized. The marl found at the base of this zone is non-biogenic and is likely the result of a rapid increase in lake (air) temperature. This zone likely records a warming and drying trend.

Stratigraphic zone 3 (SZ-3, units 6, 7; 203–62 cm*, ca. 8200–3000 BP)*

SZ-3 is characterized by consistent LOI (550°C), MS, and C/N ratios. In this zone IRD are relatively rare. This zone records the establishment of a stable trophic state in the lake basin. Marl is still present though in decreasing amounts up the core. Stable C/N values around 13 indicate that autochthonous organisms became an important source of organic carbon at this time.

Stratigraphic zone 4 (SZ-4, unit 7; 62 cm *to top of core, ca. 3000 BP to Present)*

SZ-4 consists of organic sediment that exhibits a decrease in LOI at 550°C accompanied by a slight increase in LOI at 1000°C, an increase in magnetic susceptibility, an increase in IRD counts, and slightly higher sediment influx rates. These data are indicative of a relative decline in the organic content of the core. The slightly higher C/N values in this interval suggest a decrease in lake productivity. The relative abundance of IRD in SZ-4 is an indication of mass wasting activity at this time. The lithostratigraphic changes in this zone reflect a trend towards greater landscape instability associated with cooler and possibly moister conditions at the site.

Pollen analysis

No pollen was recognized in SZ-1. Pollen concentrations in SZ-2 (295 cm–222.5 cm, > 9000 BP) were not sufficiently high enough for counting. Above 222.5 cm three local pollen assemblage zones have been identified (Figure 3).

Pollen zone 1 (PZ-1, alder-willow-aspen, 222 cm*–200* cm*, ca. 9000–8200 BP)*

PZ-1 is dominated by a shrub and herb assemblage; alder (A*lnus*) dominates the pollen spectrum with values at the base of SZ-3 of greater than 60%. Soapberry (*Shepherdia canadensis*), aspen (*Populus tremuloides*), and sedge (Cyperaceae) attain their highest values in this zone, and willow (*Salix*) and birch (*Betula*) are also relatively abundant. Total pollen concentrations average about 100×10^5 grains/cm^3. Percentages for white spruce (*Picea cf. glauca,* 3%) are relatively low in this zone. Pine (*Pinus contorta*) and fir (A*bies*) pollen grains are rare.

This zone records colonisation of the site by trees, shrubs, and herbs. Both soapberry and alder are common in landscapes that have been recently exposed. Aspen stands were probably more abundant than the pollen frequencies suggest as aspen pollen is often poorly preserved. White spruce pollen percentages (6%, 205 cm) and the presence of a spruce cone in the core (196 cm, 8140 ± 50 BP) indicate that spruce was an element of the vegetation at the site (Hebda and Allen 1983). The paucity of black spruce, pine, and fir pollen in most of this zone this zone probably indicates that these trees were absent from the region.

Pollen zone 2 (PZ-2, spruce-fir-alder, 200 cm*–44* cm*, ca. 8200–2400 BP)*

Increases in pine, white spruce, black spruce, and fir and a decline in alder values mark the beginning of the zone. Of note is variability of pollen percentages for pine, spruce, and fir which make distinct trends difficult to recognise. Alder and birch percentages tend to decrease in this zone. Black spruce pollen percentages (6%, 180 cm) suggest that it likely became a component of the local vegetation based on modern analogues (Hebda and Allen 1983). Pollen concentrations are variable with a high value of 500×10^3 cm^3. The lower portion of this zone (222–135 cm, ca. 8200–6000 BP) differs from the upper portion (135–44 cm; ca. 6000–2400 BP) in that it contains a more diverse herb assemblage and higher alder and birch percentages, perhaps indicating a more open forest developed under drier-than-present conditions.

Total pollen concentrations are highest at about 6000 BP. Fir pollen averages about 3% indicating that this species was near the site at this time (Hebda and Allen 1983). Pine pollen percentages of > 30% at 100 cm (ca. 4700 BP) indicate that pine became a significant component in nearby forests. The absence of *Artemisia* and Ericaceae in upper PZ-2 may indicate increasing moisture and the development of a mature coniferous forest surrounding the lake. PZ-2 is interpreted as a mixed forest of spruce and fir with successional stands of alder and birch. Fires may have been common in this zone as indicated by erratic pollen percentages and the abundance of seral species. The vegetation inferred for lower PZ-2 (ca. 8200–6000 BP) implies that the climate at Skinny Lake was warmer than present. The upper zone (ca. 6000–2400 BP) was also warm but may represent a trend towards increased moisture beginning at 6000 BP.

Pollen zone 3 (PZ-3, pine-western hemlock, 44 cm*–0* cm*, ca. 2400 BP–present)*

This zone is characterised by the rapid increase in pine pollen percentage to values greater than 50% suggesting that pine migrated to the site at this time (Hebda and Allen 1983). Spruce, fir, and alder pollen tend to decrease through this zone. Of note is the significant increase in Western hemlock (*Tsuga heterophylla*) and cedar type (Cupressaceae) near the

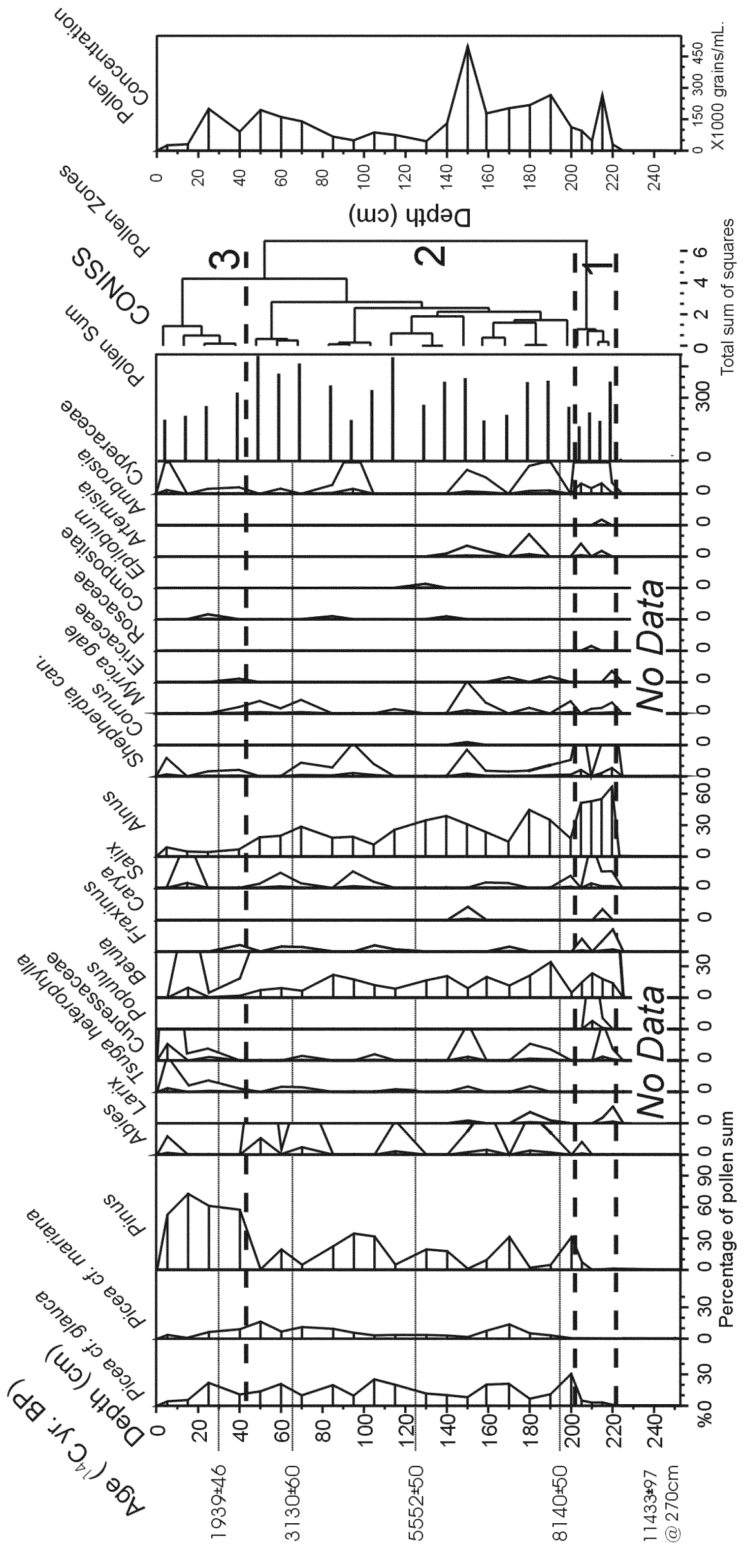

Figure 3. Relative pollen diagram. Dashed line indicates zone boundaries.

top of this zone (ca. 1000 BP). Western hemlock and cedar are presently found 100 km west of the site (Friesen 1985; Hosie 1990). The observed increase may be a response to 1) a change in the range limit of these species (movement inland), 2) an increase in the productivity of these species within their range limit, and/or 3) less pollen productivity locally and a stronger regional pollen. The first two scenarios are indicative of increased moisture at coastal sites. The third scenario may occur in response to the development of a more open forest as modern cool and moist climate is established. All three scenarios are indicative of environmental change. Pollen concentrations tend to decrease up through this zone, most likely a result of decreasing sediment density.

Discussion and external comparisons

The lithostratigraphic record (Figure 2) indicates that a prolonged period of high clastic sedimentation and low productivity (ca. 11500–9000 BP; Units 1, 2, 3) persisted following deglaciation. High IRD counts are likely associated with landscape instability (associated with recent deglaciation) which would result in an increase in the frequency of sediment transfer onto lake ice. The lack of pollen in this interval may not be an indication of a barren landscape. More likely poor preservation coupled with rapid clastic accumulation rates acted to dilute the pollen record. Other records (Cwynar (1988), Stuart et al. (1989); Figure 1) indicate early deglaciation as well but do not contain the significant lag in biotic response observed at Skinny Lake. Nearby alpine studies (Spooner et al. 1997; Mazzucchi 2000) indicate later deglaciation (ca. 10000 BP) suggesting that in the Skinny Lake region major trunk valleys may have become ice free well in advance of alpine areas. Ryder and Maynard (1991) have indicated that rapid discharge in major river valleys in northern British Columbia may have resulted in early deglaciation of these valleys. We suggest that the late Pleistocene and earliest Holocene at Skinny Lake was characterised by cool and moist conditions and an unstable landscape generated by complex temporal patterns of deglaciation.

The palynology of Skinny Lake indicates that early Holocene vegetation colonisation (9000–8200 BP) was followed by an extended period (8200–2400 BP) with only subtle change. Prominent marl deposits (units 4, 5; c.a. 9000–8000 BP) indicate that warmer-than-present conditions may have existed during the early Holocene. Similar indications of early Holocene warming have been observed to the south at Seeley Lake (Gottesfeld et al. (1991); Figure 1), west at Susie Lake (Spooner et al. (1997): Figure 1) and to the north at Pyramid Lake (Mazzucchi et al. (2000); Figure 1). The middle Holocene is characterized as a period of enhanced productivity under warmer and drier conditions than present. During the early-middle Holocene a mixed seral forest dominated. Highly varying LOI values and large fluctuations in pollen percentages, especially of seral species may indicate that fires were common during this period and that drier-than-present conditions likely existed (Mazzucchi 2000). Most records from the region (*see* Figure 4) indicate a trend towards slightly cooler and possibly wetter climate from ca. 6000–4000 BP. At Skinny Lake the low resolution of the pollen data and the lack of a treeline response due to the low elevation of the site probably preclude recognition of these trends (or the lack there of). The difficulty in interpreting moisture trends within this region may also, in part, be due to the establishment of local moisture regimes in response to local/regional physiographic differences (*see* Figure 4). Demarchi (1996) has indicated that the rugged relief typical of the area leads to complex patterns of surface heating and cold air drainage.

Significant differences exist in the timing of the transition to the modern state in the lithostratigraphic and palynological records from Skinny Lake (Figure 3,4). Other sites in northern British Columbia (Figure 4) record a transition to the modern state from 4000 yr. BP (western sites) to 3000 BP (eastern sites) mainly on the basis of transitions in pollen and macrofossil records. At all of these sites local conditions such as proximity to treeline (Spooner et al. (1997); Figure 4) or sensitivity to moisture input (White and Mathewes 1982) result in an enhanced sensitivity to climate change. At Skinny Lake the apparent lack of a pollen response at 3000 BP may be due the lack of a strong local forcing mechanism that would induce a vegetative response to a subtle but persistent change in regional climate.

In the lithostratigraphic record, decreasing LOI (550°C) and a trend of increasing magnetic susceptibility compliment an increase in IRD during the Late Holocene. The increase in IRD (and associated proxies) in this interval may be an indication of an increase in the frequency of avalanches generated during early spring thaw before the lake had become ice free. The

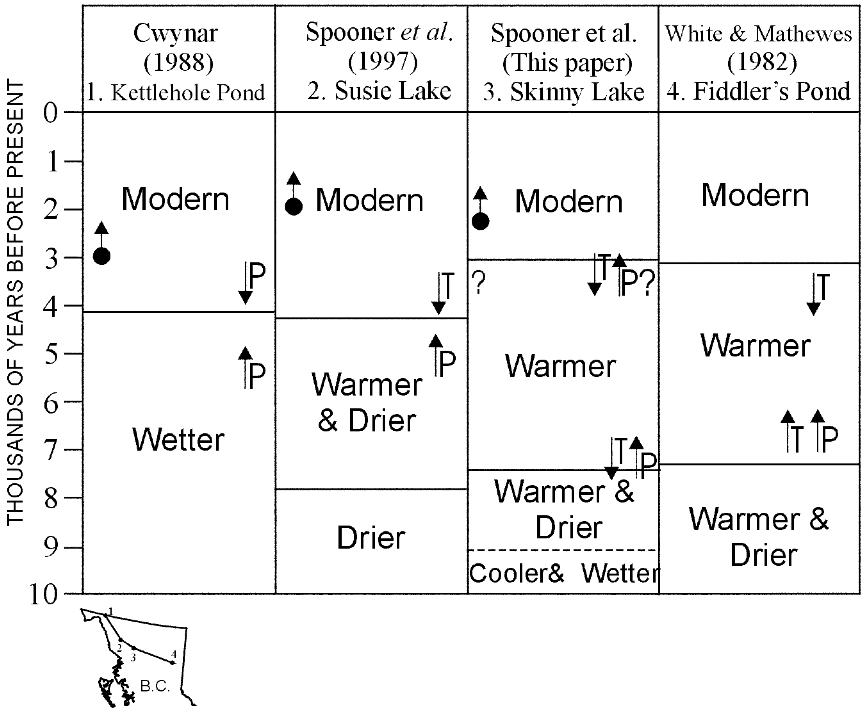

Figure 4. Summary of Holocene climatic states and trends for select sites in northern British Columbia. P = Precipitation, T = Temperature. Solid boundaries denote times of marked change, dashed boundaries denote gradual or uncertain times of change. Circle with arrow indicates approximate timing of airmass circulation change.

IRD could then be generated by debris flow run-out over the lake ice (Gardner 1983) (Figure 1). Full-depth wet snow avalanches are most common in the early spring as melt-freeze conditions result in an isothermal snow pack. Debris flows associated with the avalanche occur as the downslope areas traversed by the avalanche are often ice free and partially thawed (Gardner 1983). These conditions are most common on south facing slopes and occur before any lake ice has weakened. The increase in IRD may also be partially related to increase in the duration and thickness of ice cover. Spooner (1998) has suggested that shallow, high surface area to volume ratio lakes like Skinny Lake are hydrodynamical unstable and characteristically polymictic. The thermal inertia of these lakes is low and therefore these lakes are likely susceptible to changes in the phenology of lake ice initiation and break-up.

These data, in concert with higher C/N ratios and a subtle increase in clastic sediment influx rates suggest that both increased transfer of sediment onto lake ice and decreased lake productivity characterised this time. Decreased lake productivity can be the result of prolonged ice cover, cooler lake water temperatures, increased turbidity, and decreased insolation. Thus, the lithostratigraphic indicators, taken together, suggest that the transition to modern climate state occurred at 3000 BP. The timing of the transition to modern state in the lithostratigraphic record correlates closely with the initiation of the Tiedemann glacial advance (ca. 3300 BP) in southwestern British Columbia (Ryder and Thomson 1986) and the Peyto and Robson advances (ca 3300 BP) in the Canadian Rockies (Luckman et al. 1993). Elsewhere in northwestern British Columbia Pellat and Mathewes (1994), Spooner et al. (1997) recognized coincident declines in tree line at alpine sites in the Queen Charlotte Islands and the northern Coast Mountains respectively.

Pollen trends indicate a significant change at about 2400 BP. The dramatic increase in pine (60%) in this interval has been noted elsewhere (Gottesfeld et al. 1991) but is difficult to explain, given the modern cool, relatively dry climate that exists at the site. At Skinny Lake the increase in pine may be partially in response to migration to the site (Hebda and Allen 1983). As well, Gottesfeld et al. (1991) have postulated that an increase in anthropogenic fire would favour pine. Friesen (1985) argued that fire was common to the region during this time and also suggested an anthropogenic source. Cupessaceae and hemlock increase in significance though neither of these species is currently found at the site. A local reduction in pollen productivity in response to establishment of modern climate at the site may be partially responsible. This increase may also be in response to migration of these species up the Iskut Valley at least as far as their present position, within 100 km of Skinny Lake (Friesen 1985) and/or an increase in productivity at their range limit. All mechanisms indicate changing climate conditions.

An increase in the frequency of long-distance transported pollen has been noted at other north-western sites (Cwynar (1993), Spooner et al. (1997); *see* Figure 4) and has been attributed to a inland shift in the penetration of the moist Pacific air mass (Cwynar 1993). The timing of this change is not well constrained but in all cases it occurs significantly later than the transition to the modern temperature and precipitation state (Figure 4). The Aleutian Low (AL) pressure system located in the Gulf of Alaska may have a significant impact on regional air mass circulation. In a study of modern climate records in northwestern British Columbia Baltzer et al. (2001) have noted that most sites in this region are sensitive to temperature and moisture change associate with AL dynamics. A deepened AL (which becomes dominant in the late fall and winter) would result in enhanced cyclonic (counter clockwise) flow of warm moist P*acific* air inland and would provide a viable mechanism for the transfer of exotic pollen inland. A consistently deepened AL has been correlated with a positive Pacific Decadal Oscillation (PDO) state (Mantua et al. 1997).

Conclusions

The results from Skinny Lake provide a link between coastal paleoclimate records and those located on the eastern margins of northern British Columbia. This record indicates that the site became ice free about 11500 BP and that initial regional warming did not take place until 9000 BP. The middle Holocene was slightly warmer than present, however, the moisture state remains uncertain. The transition to modern state is not synchronous in the lithostratigraphic and fossil pollen datasets indicating either a significant lag in response time to a single event or, more likely, that each dataset may document a response to a different environmental event. We suggest that regional cooling occurred at ca. 3000 BP and was followed at ca. 2400 BP by a significant change in air mass circulation perhaps resulting from a consistently deepened Aleutian Low.

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