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Lateglacial and early Holocene climate oscillations in the Matanuska Valley, south-central Alaska

Zicheng Yu[a,*], Karina N. Walker[a], Edward B. Evenson[a], Irka Hajdas[b]

[a]*Department of Earth and Environmental Sciences, Lehigh University, 31 Williams Drive, Bethlehem, PA 18015, USA*
[b]*AMS ^{14}C Lab ETH/PSI, ETH Hoenggerberg HPK H27, CH-8093 Zurich, Switzerland*

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Abstract

Here we present multi-proxy data from two cores taken from Hundred Mile Lake in the Matanuska Valley of south-central Alaska to investigate the climate, vegetation and deglaciation history of the last 14,000 years. The chronology of the cores was controlled by five AMS dates. Sediment lithology changes from clay at ~14–13 ka (1 ka = 1000 cal BP), through marl at 13–8 ka, to gyttja at 8–0 ka. The transition from clay to marl probably represents increased productivity of the lake in a stabilizing watershed, induced by initial Allerød climate warming after ice retreat, as calcite precipitation of marl was facilitated by *Chara* photosynthesis under high water temperature in summer. The $\delta^{18}O$ record obtained from bivalve *Pisidium* mollusc shells shows several large shifts between 13 and 8 ka. A negative excursion of ~2‰ in $\delta^{18}O$ at 12.4–11.5 ka is suggestive of a regional expression of the Younger Dryas (GS-1) cooling event. A 4.5‰ negative shift from −10.5‰ at 13–11 ka to −15‰ at 10.5–8 ka occurred during the peak carbonate interval around 10.5 ka. This surprisingly large negative shift in $\delta^{18}O$ values during the early Holocene thermal maximum has not been documented elsewhere in the region. This shift suggests a major change in atmospheric circulation patterns, possibly through more frequent easterly flow of warm and dry air masses that are also depleted in ^{18}O. Pollen results from marl sediments indicate vegetation change from a herb tundra, through shrub birch-dominated tundra, to an alder forest, which follows closely with other regional pollen sequences in south-central Alaska. The results from this study suggest that the climatic shift during the early Holocene was of greater magnitude than the Younger Dryas event, implying the importance of regional feedback processes in high latitudes in controlling regional climate response to large-scale forcing.

1. Introduction

Several broad-scale climate changes have been documented in Alaska during the postglacial period. During the Last Glacial–Interglacial transition, climate reversals to cold conditions punctuated deglacial warming trends. These have been documented in Glacier Bay (Engstrom et al., 1990), southwestern Alaska (Hu et al., 2002), and Kodiak Island (Peteet and Mann, 1994). Kaufman et al. (2004) reviewed paleoclimate evidence for the western Arctic (including Alaska) and found that timing and magnitude of early Holocene warming period, referred to as the Holocene thermal maximum (HTM), vary in different regions. Also, the spatial pattern of warming during the HTM is similar to the warming patterns currently observed in the Arctic during the last few decades (Kaufman et al., 2004). Several pollen diagrams have been published in south-central Alaska, mostly around Cook Inlet and on Kenai Peninsula (Ager and Brubaker, 1985; Ager, 1999, 2001). However, the Lateglacial climate oscillations and Holocene climate changes from multiple proxy data have been poorly documented, especially in the Matanuska Valley.

The goal of this study is to document and understand postglacial environmental change of the Matanuska Valley in south-central Alaska. Specific objectives of the study were (1) to derive new multiple proxy records of climate and vegetation change from lake sediments from the upper Matanuska Valley; (2) to elucidate and understand climatic oscillations during the last deglaciation and early Holocene; and (3) to provide additional constraint on the deglaciation history of the region. In this paper, we will focus on the analysis results from authigenic carbonates formed during the Lateglacial and early Holocene period, for which we have good AMS-dating

*Corresponding author. Tel.: +1 610 758 6751; fax: +1 610 758 3677.
E-mail address: ziy2@lehigh.edu (Z. Yu).

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control. We found lithological and possibly oxygen-isotopic evidence for the Younger Dryas (YD) climate event and also a previously unrecognized climate shift during the early Holocene, indicated by 4.5‰ negative shift in $\delta^{18}O$.

2. Study region and site

The Matanuska Valley is located in south-central Alaska between the Talkeetna Mountains in the north and Chugach Mountains in the south (Fig. 1(B)). The study region is within the physiographic region of the Pacific mountain system, with a transitional maritime-continental climate (Anderson and Brubaker, 1993). The local climate of the region is influenced by the orographic effect of the Chugach Mountains and its proximity to the Matanuska glacier (Fig. 1(C)). The temperatures range from −15 to −6 °C in January, and from 7 to 19 °C in July. Mean annual precipitation is about 400 mm at Matanuska (Fig. 2(B)). The region is located within the interior boreal forest ecozone (Viereck and Little, 1975), dominated by white spruce (*Picea glauca*) on the uplands and black spruce (*Picea mariana*) in the lowlands. The forest also includes birch (*Betula*), alder (*Alnus*), and various grasses (Poaceae).

The bedrock is comprised of the Cretaceous Matanuska Formation, which includes diverse fossiliferous marine shales and calcareous concretions, volcanic–lithic siltsone, sandstone, and conglomerates (Winkler, 1992). The surficial geology of the region consists largely of Quaternary deposits, which include moraines, thin tills, loess, and outwash. Deglaciation of the study region is dated at ~13,000 radiocarbon yr BP (Fig. 1(C); Williams and Ferrians, 1961; Larson et al., 2003).

Hundred Mile Lake (HML) is located at 61°48′40″N and 147°50′10″W at an elevation of 506.3 m above sea level, approximately 5 km from the terminus of the Matanuska Glacier (Fig. 1(C)). The lake is situated between two discontinuous moraines (Fig. 1(C); Larson et al., 2003). The lake is small (~500 × 200 m^2) and located in an isolated watershed of about 1 km^2 in size (Fig. 1(C)). Its maximum water depth is more than 2.4 m deep as measured during lake coring. Calcite precipitation is active under shallow water in the summer (S. Johnson, 2004, personal communication). Its hardwater nature is likely due to groundwater input through the calcareous bedrock in the watershed. Oxygen and hydrogen isotope values of a water sample collected on 18 August 2007 were −15.4‰ and −139‰ relative to SMOW, respectively.

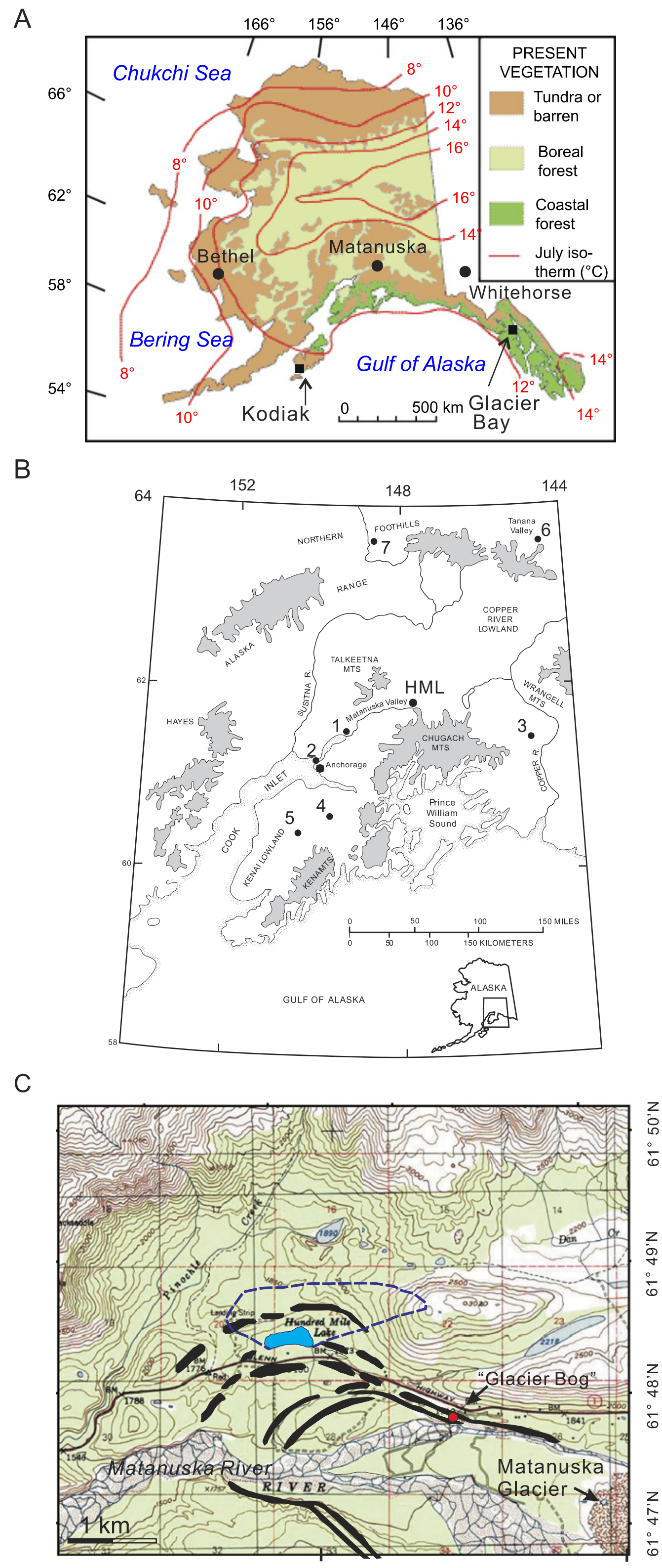

Fig. 1. Location maps: (A) map of vegetation zones and July temperature distribution in Alaska, also showing the locations of climate stations Bethel, Matanuska (Alaska) and Whitehorse (Yukon). July temperatures are higher in the interior than the coast regions. Also shown are Kodiak Island and Glacier Bay (■) as discussed in the text. (B) Map of south-central Alaska showing Hundred Mile Lake (HML) and other published study sites (map modified from Ager, 1999): 1. Kepler Lake (Forester et al., 1989); 2. Point Woronzof (Ager, 1983); 3. Seventy Mile Lake (Ager and Brubaker, 1985); 4. Hidden Lake (Ager, 1983); 5. Tera Lake (Ager, 2001); 6. Tangle Lakes (Schweger, 1981); 7. Windmill Lake (Bigelow and Edwards, 2001). (C) Topographic map showing topographic features and catchment area (dashed line) around Hundred Mile Lake. The bold black features are moraines of the Matanuska Glacier (modified from Williams and Ferrians, 1961). "Glacial Bog" (red dot) is the peat deposits from which Williams (1986) obtained the two basal ^{14}C dates discussed in the text.

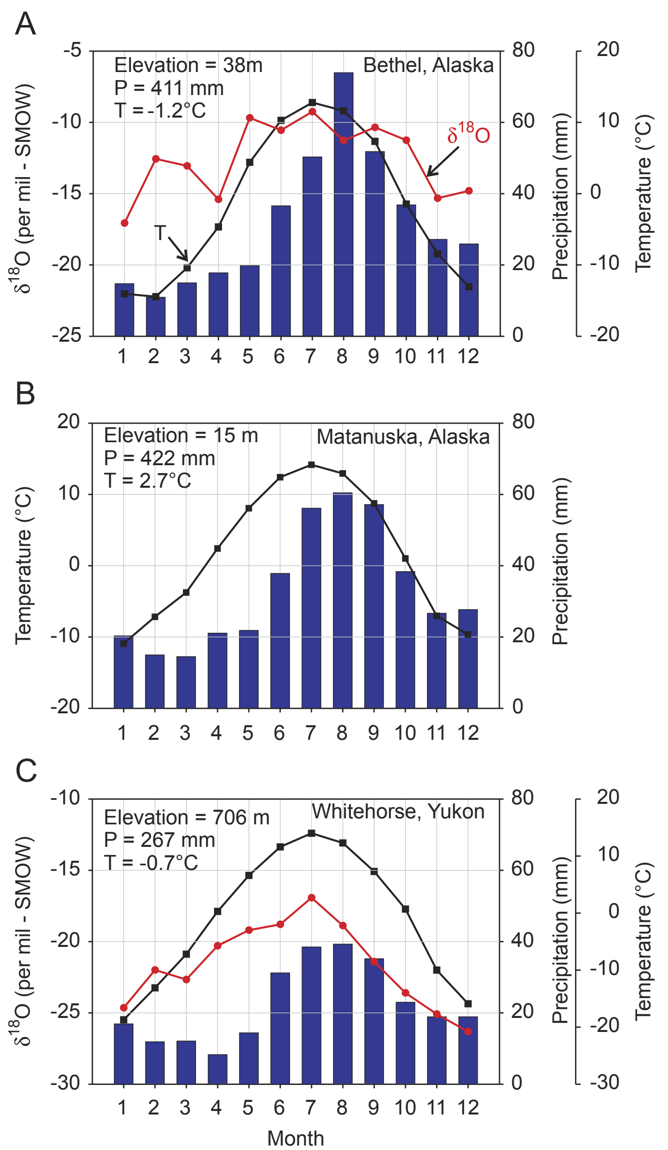

Fig. 2. Climate and precipitation isotope plots: (A) monthly temperature (squares), precipitation (bars) and isotopes in precipitation (dots) at Bethel, Alaska; (B) monthly temperature (squares) and precipitation (bars) at Matanuska, Alaska; (C) monthly temperature (squares), precipitation (bars) and isotopes in precipitation (dots) at Whitehorse, Yukon. P = mean annual precipitation, and T = mean annual temperature (climate normals 1971–2000). All isotope data were from IAEA/WMO Global Network of Isotopes in Precipitation (GNIP) database (http://isohis.iaea.org).

3. Methods

3.1. Field sediment coring

Two sediment cores were recovered from HML in early January 2004 with a Livingstone–Wright piston corer (Wright et al., 1967). Core 1 (under 2.3 m of water) is 265 cm long to cover the entire sediment sequence. To strengthen the duplicability of the records from Core 1, we also collected Core 2 (under 2.4 m of water) about 16 m away, from which we only kept the lowest 2 m of sediments. One meter long core segments were extruded in the field and wrapped in plastic wrap and aluminum foil in halved PVC pipes for transportation back to the laboratory. Sediment characteristics were noted in the field and then described in detail in the laboratory. The cores have been stored in a cold room at 4 °C.

3.2. Loss-on-ignition (LOI) analysis

LOI analysis was performed at 1 cm intervals for the two HML cores. Weight loss after 1 h combustion at 550 °C was used to estimate organic matter content, and weight loss after 1 h at 1000 °C for calcium carbonate content. Silicates are the rest of the dry sediments.

3.3. Macrofossil analysis and AMS radiocarbon dating

Subsamples of 2-cm sections were taken at 10 intervals and sieved for terrestrial plant macrofossils, but only one sample produced adequate material for dating. One terrestrial plant macrofossil from Core 2 at 462–464 cm and one bulk organic-rich sediment at 373 cm from Core 1 were submitted for AMS ^{14}C dating at the University of Arizona. Four bivalve *Pisidium* shell samples were picked, cleaned with distilled water, and dissolved in concentrated phosphoric acid. The resulting CO_2 was then purified and reduced to filamentous graphite (Hajdas et al., 2004) for analysis at the AMS ^{14}C facility at ETH Zurich, Switzerland. The ^{14}C dates on *Pisidium* shells were corrected for old carbon effect of 1250 years based on measurements of terrestrial macrofossil and shell samples deposited only 2 cm apart. All radiocarbon dates were calibrated by the calibration program CALIB Rev. 5.0.1 using INTCAL 04 data set (Reimer et al., 2004) (Table 1).

3.4. Stable-isotope analysis

Stable-isotope (O, C) analysis was performed on three types of carbonates: *Pisidium* mollusc shells (aragonite), ostracode (a *Candona* species) shells (low Mg-calcite), and *Chara* encrustations (calcite) (Fig. 3(C)). Sieved and freeze-dried marl samples were examined under a stereo-microscope, and mollusc/ostracode shells or calcite crystals (encrustations) were picked and cleaned. Most analysis was from *Pisidium* shells (Table 2), which were abundant and provided adequate materials at most levels.

The analytical procedure and methodology are similar to that used for deep-sea sediment (McCrea, 1950). Each carbonate sample was reacted with 100% phosphoric acid (H_3PO_4) at a constant temperature of 25 °C overnight to produce CO_2. The released CO_2 was analyzed for its $^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios with a Finnigan MAT 252 mass spectrometer in the stable-isotope lab at Lehigh University. Results were presented as conventional delta notation (δ), defined as $[(R_{sample}–R_{standard})/R_{standard}] \times 1000$ (where R is the absolute ratio of $^{18}O/^{16}O$ or $^{13}C/^{12}C$, and the Vienna-PDB (Pee Dee belemnite) is the standard). The analytical precision is ± 0.1‰ for both $\delta^{18}O$ and $\delta^{13}C$.

Table 1
AMS ^{14}C dates from Cores 1 and 2 at Hundred Mile Lake, Alaska

HML core	Depth (cm)	Material dated	AMS lab. no.[a]	δ^{13}C (‰—VPDB)	^{14}C date (±SE)	Hardwater effect corrected ^{14}C date[b]	Calibrated age (2σ range in cal BP)[c]	Calendar age (cal yr BP)	Note
1	373	Bulk organic	AA-59593	−26.9	6630±40	NA	7456−7575	7515	398 cm of equivalent depth in core 2
2	462–464	Terrestrial plant macrofossil	AA-59592	−25.0	9650±50	NA	10,785−11,198	10,991	4.1 mg organic matter
2	464–466	*Pisidium* mollusc shells	ETH-29987	1.3±1.2	10,990±100	9740±100	10,746−11,343	11,045	1 mg C
2	474–476	*Pisidium* mollusk shells	ETH-29988	0.8±1.2	10,660±80	9410±80	10,411−10,868	Rejected	1.9 mg C
2	500–502	*Pisidium* mollusk shells	ETH-29989	0.3±1.2	11,670±80	10,420±80	12,054−12,656	12,355	2.6 mg C
2	523–525	*Pisidium* mollusc shells	ETH-29990	1.0±1.2	12,570±85	11,320±85	13,049−13,361	13,205	2.6 mg C

[a]AA—University of Arizona, NSF—funded AMS lab.; ETH—Eidegenossische Technishe Hochschule, Institut fur Teilechenphsik.
[b]Correction of 1250 years based on difference between macrofossil date of 9650 and shell date of 10,990 ^{14}C BP and the 2-cm depth difference (~50 years/cm sediment-accumulation rate between 398 (373 in Core 1) and 463 cm).
[c]Calibrated by the program CALIB Rev 5.0.1 (Stuiver and Reimer, 1993) using INTCAL 04 data set (Reimer et al., 2004).

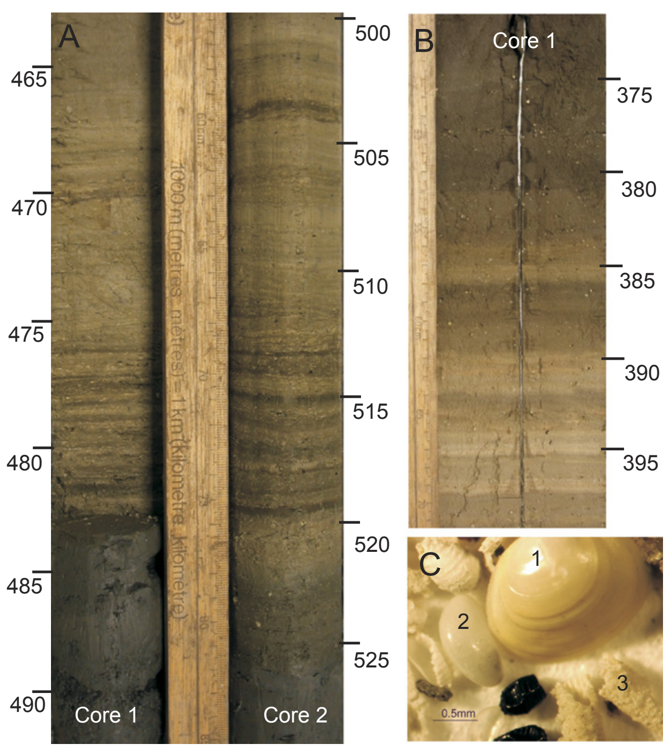

Fig. 3. Photos of cores and carbonate fractions from Hundred Mile Lake, Alaska: (A) photos of basal sections of Cores 1 and 2 showing the transition from silty clay to marl; (B) photo of Core 1 showing the transition from marl to gyttja. Depth scales are centimeters below the lake surface; (C) micrograph of carbonate materials used for isotopic analysis: 1. Bivalve mollusc shells, 2. Ostracode (*Candona*) shells, and 3. *Chara* encrustations.

3.5. Pollen analysis

Pollen analysis was carried out at 18 levels from the marl section of HML Core 2. Pollen samples were prepared using standard preparation procedure (Fægri and Iversen, 1989) as modified and described in Yu (2000), including treatments of HCl, KOH, HF, and acetolysis. Pollen was identified under a compound microscope at 400× magnification aided with the use of published pollen keys and reference slides. Pollen concentration was calculated with an added known number of *Lycopodium* spores (Maher, 1981). Pollen diagram was generated using Tilia and TGView programs by Eric Grimm of Illinois State Museum. CONISS was carried out using pollen taxa having >2% of the total pollen sum.

4. Results

4.1. Lithology and LOI results

Lithology of the two cores from HML is based on visual sediment inspection and LOI analysis results. Lithology of the 265 cm long Core 1 consists of a basal section of 12 cm mineral material (gravels, clays, and silts) from 495 to 483 cm below lake surface (Fig. 3(A)). Above that is 93 cm marl sediment (from 483 to 390 cm), which then changes to gyttja with some bandings between 390 and 383 cm (Fig. 3(B)). The section of the core from 383 to 377 cm is the transitional gyttja, with the presence of mollusc shells, until the total disappearance of mollusc shells at 375 cm (Fig. 3(B)). Gyttja then completely dominates from 377 to 230 cm, with several light-colored bands located at 366, 330, and 310–305 cm.

Lithology of the 200-cm-long HML Core 2 shows the same sequence of mineral materials, marl, and gyttja as Core 1. A mineral material layer (gravels, clays, and silts) comprises the lower 25 cm (550–525 cm) of HML Core 2 (Fig. 3(A)). Above that is 120 cm of marl sediments that change into gyttja at 405 cm. Visibly distinct bands

Table 2
The $\delta^{18}O$ and $\delta^{13}C$ values from carbonate of Core II at Hundred Mile Lake, Matanuska Valley, Alaska

Depth (cm)	Age (cal BP)	*Pisidium* shells		Ostracode shells		*Chara* encrustations	
		$\delta^{18}O$	$\delta^{13}C$	$\delta^{18}O$	$\delta^{13}C$	$\delta^{18}O$	$\delta^{13}C$
405	7889	−14.1	−4.9				
409	8103	−14.4	−4.7				
413	8317	−15.9	−6.3				
417	8531	−14.5	−5.9				
421	8745	−14.8	−6.1				
425	8959	−15.7	−8.0				
429	9173	−13.6	−4.9				
433	9387	−14.3	−6.8				
437	9601	−14.6	−5.0				
438	9654	−14.1	−5.2				
441	9815	−14.9	−5.9				
445	10,028	−14.6	−6.5				
449	10,242	−14.9	−5.8				
453	10,456	−13.4	−4.0				
454	10,510	−12.2	−4.5				
458	10,724	−11.7	−5.3				
460	10,831	−11.9	−5.8				
461	10,884	−11.5	−5.7	−12.2	−2.2		
462	10,938	−10.4	−6.3				
463	10,991	−10.7	−5.1				
465	11,045	−11.8	−5.0				
466	11,081	−10.3	−3.6				
469	11,191	−10.5	−4.8				
475	11,409	−10.7	−3.6				
476	11,445	−10.8	−2.6				
477	11,482	−12.8	−4.9				
479	11,554	−11.0	−3.2				
481	11,627	−12.2	−3.3				
482	11,664	−10.6	−3.6				
484	11,736	−11.2	−2.9				
485	11,773	−12.8	−4.5				
487	11,846	−12.0	−3.3				
489	11,918	−12.2	−3.1				
491	11,991	−12.2	−4.2				
493	12,064	−11.8	−3.1				
495	12,137	−11.1	−2.4				
497	12,209	−12.0	−2.6				
499	12,282	−11.0	−2.7				
500	12,319	−10.8	−3.2				
501	12,355	−10.5	−3.5	−11.2	−0.6		
503	12,429	−9.9	−3.4	−10.2	−0.9		
505	12,503	−11.2	−3.2	−9.2	−0.6		
509	12,651	−11.2	−3.5				
514	12,835	−10.2	−2.9				
516	12,909	−10.6	−2.2			−12.5	3.1
517	12,946	−11.6	−2.5				
521	13,094	−10.3	−2.8				
524	13,205	−10.3	−1.9				
525	13,242	−10.8	−2.1				

throughout HML Core 1 are also present in HML Core 2 (Fig. 3(A)).

Both cores show similar stratigraphy of sediment composition from LOI analysis (Fig. 4). The basal clay contains >90% silicates. Marl contains ~20–80% carbonate, and 10–30% organic matter. Gyttja contains ~30–40% organic matter and <10% carbonate.

4.2. ^{14}C dates and chronology

Six ^{14}C AMS dates from HML were obtained, corrected for old carbon effect of shell dates and calibrated to calendar ages (Table 1). The date of 10,660 ^{14}C BP (ETH-29988) was rejected due to the dating reversal. An age model was developed based on linear interpolation of calibrated ages of five accepted dates and the age of the sediment surface (2004 AD = −54 cal BP) (Fig. 5). Extrapolation of the age–depth relationship to the base of the core (550 cm) provides an estimated age of ~14 ka (1 ka = 1000 cal yr BP). This tentative chronology of the sediment core will be used for the following discussion.

We realized that our age model has relatively large dating uncertainty during the Lateglacial period, due mostly to the constant old carbon correction factor of 1250 years that we used. The correction factors could be variable during that 3000-year period. Also, even assuming a constant old carbon effect, an alternative old carbon correction would be 370 years, if we reject the first shell date (ETH-29987). That alternative age model would put the initial warming (i.e., the sharp increase in carbonate at 525 cm) at 14.1 ka and the isotopic shift at the YD at 13.2–12.1 ka. In any case, the estimated uncertainty is around 500 years. This correction procedure would not affect the age model during the early Holocene, as it is relatively well constrained by the AMS age of 10,991 cal BP on terrestrial macrofossils. Also, pollen stratigraphy at HML appears to support the chronology in general terms, especially when comparing with well-dated pollen sequence from Windmill Lake in central Alaska (Bigelow and Edwards, 2001).

4.3. Stable-isotope analysis

The $\delta^{18}O$ and $\delta^{13}C$ values obtained from bivalve *Pisidium* shells range from about −10‰ to −16‰ and from −1.8‰ to −8‰, respectively (Fig. 6). The few analysis from *Chara* encrustations and ostracode shells show mostly negative offsets for $\delta^{18}O$ but positive offsets for $\delta^{13}C$. We focus the following description only on the results from bivalve *Pisidium* shells (curves in Fig. 6).

The lower marl sediments (13.3–12.3 ka) have the highest $\delta^{18}O$ values of −10.5‰. A decrease in $\delta^{18}O$ values of ~2‰ occurs from 12.3 to 11.5 ka, with some large-magnitude fluctuations. A rapid increase to the high value of −10.5‰ occurs at 11.5–10.9 ka, which is followed by the large decrease of $\delta^{18}O$ values of −4.5‰ between 10.9 and 10.2 ka. From 10.2 to 7.9 ka there is considerable fluctuation in values from −13.6‰ to the lowest of the $\delta^{18}O$ values at −16‰.

The $\delta^{13}C$ values show a general decline trend from −1.8‰ at 13.3 ka to −8‰ at 9 ka. Superimposed on this declining trend, it appears that variability increases from a fluctuating range of 1–1.5‰ at 13.3–11.5 ka to 2–3‰ at 11.5–8 ka. During the Lateglacial period from 13.3 to

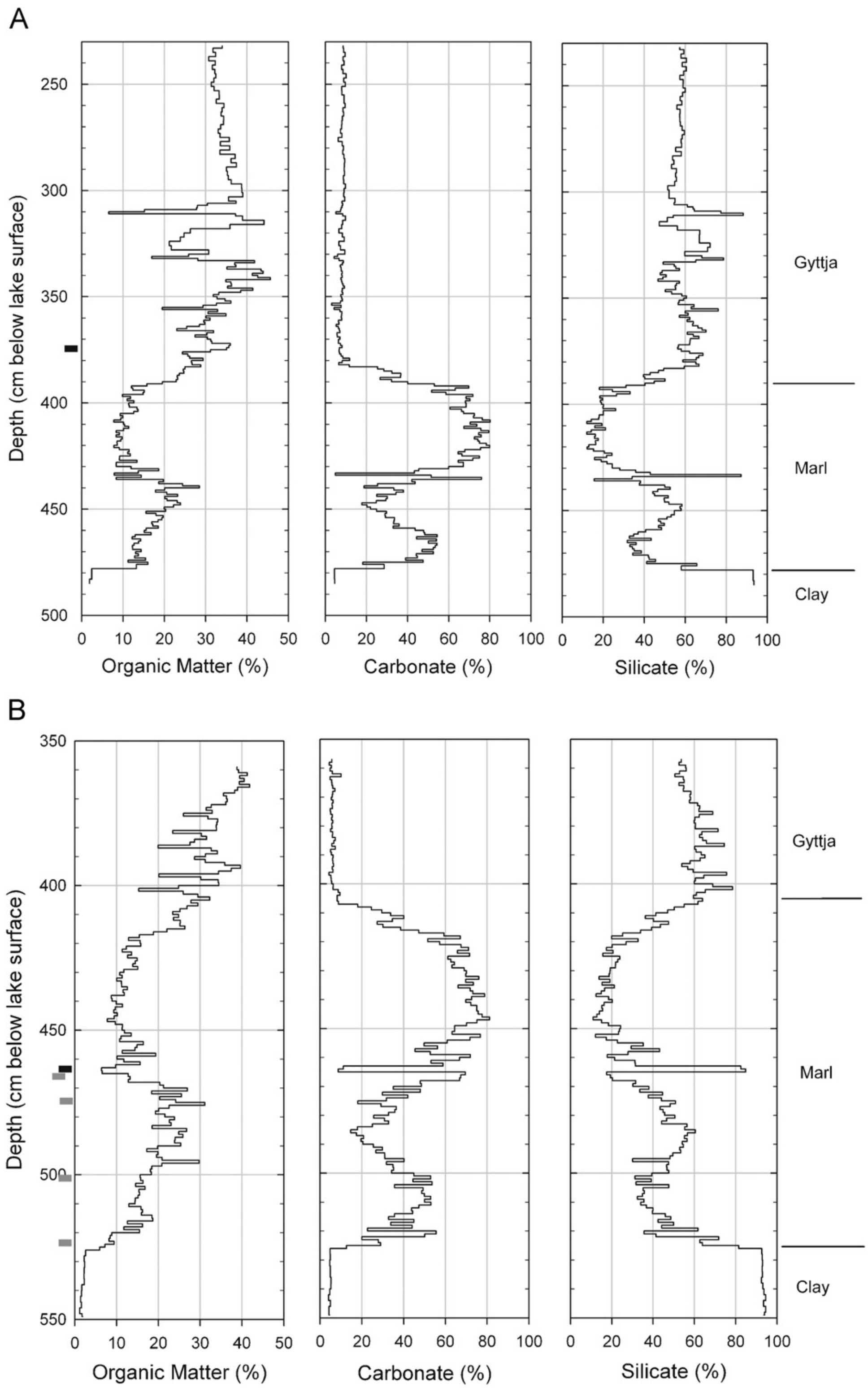

Fig. 4. Loss-on-ignition results at Hundred Mile Lake, Alaska: (A) Core 1 and (B) Core 2. Rectangles on the left side show the location of AMS dates: black = terrestrial plant macrofossils; dark gray = bulk organic sediment; and light gray = mollusc shells.

11.5 ka and the early Holocene at 10.2–8 ka $\delta^{18}O$ shows positive relation with $\delta^{13}C$. However, during the transition in the first millennium of the Holocene, decreasing $\delta^{18}O$ appears to correlate with increasing $\delta^{13}C$.

4.4. Pollen analysis results

The pollen diagram at HML was divided into three pollen zones based on CONISS cluster analysis (Fig. 7).

Zone HML-1 (525–515 cm; 13.3–12.9 ka): This lowest zone is characterized by *Artemisia* (30%), Cyperaceae (up to 20%) and *Salix* (around 20%). At the end of this first zone a steady increase in *Betula* is concomitant with decreases in *Artemisia*, Cyperaceae, and *Salix*.

Zone HML-2a (515–467 cm; 12.9–11.1 ka): This subzone is characterized by high *Betula* (80%) with variable amounts of *Artemisia* (10%), Cyperaceae (<10%), *Salix* (<10%), and *Alnus* (10%). Within this subzone there is a period of increased *Alnus*. *Zone HML*-2b (467–430 cm; 11.1–9.2 ka). This subzone is still characterized by high *Betula* (~70%),

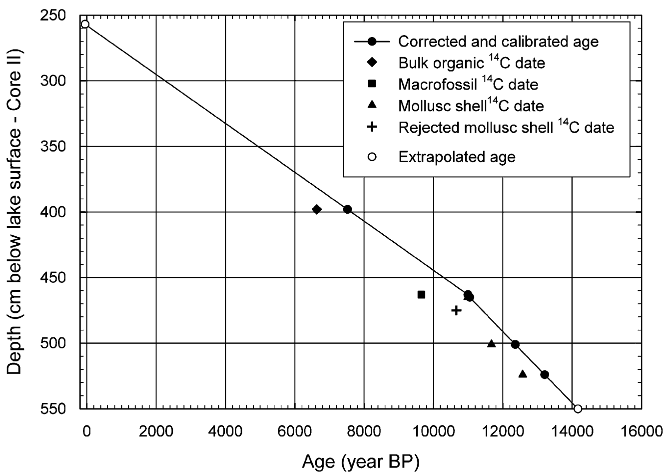

Fig. 5. Age–depth plot (age model) for composite core using Core 2 depth scale at Hundred Mile Lake, Alaska. The line is the age–depth model used in all the discussion in the text. Also shown are individual uncorrected and uncalibrated ^{14}C dates.

but with increased amounts of *Salix* (~10%) and Cyperaceae (<15%). Also, this subzone represents the time period when Polypodiaceae reached maximum value (10%). At the end of this subzone *Alnus* begins to increase while most other taxa begin a gradual decline.

Zone HML-3 (430–409 cm; 9.2–8 ka): This uppermost zone is characterized by abundant *Alnus* (up to 60%) and decreases in *Betula* (~40%). Also, the zone has a steady increase in *Picea* (up to 20%). *Artemisia*, Cyperaceae, and *Salix* are still present but at low amounts.

5. Discussion

5.1. Postglacial development of the lake and watershed

Lithology can be used as an indicator of lake productivity and watershed stability. The basal mineral (clay and silt) sediments likely represent glaciolacustrine deposits when the glacier was still within the watershed or immediately after ice retreat prior to hillslope stabilization by vegetation. The presence of marl sediments in HML is likely the result of hardwater chemistry derived from groundwater in the surrounding bedrock on the north side of the Matanuska River, which contains calcareous concretions within the Matanuska Formation (Winkler, 1992). Shallow lake water tends to favor calcite precipitation during the summers (Kelts and Hsu, 1978), which at HML has been facilitated by *Chara* photosynthesis. At HML, the sharp increase in carbonates around 13.3 ka (Fig. 8) suggests the initial warming after ice retreat locally and an increase in lake productivity. This initial marl was deposited during

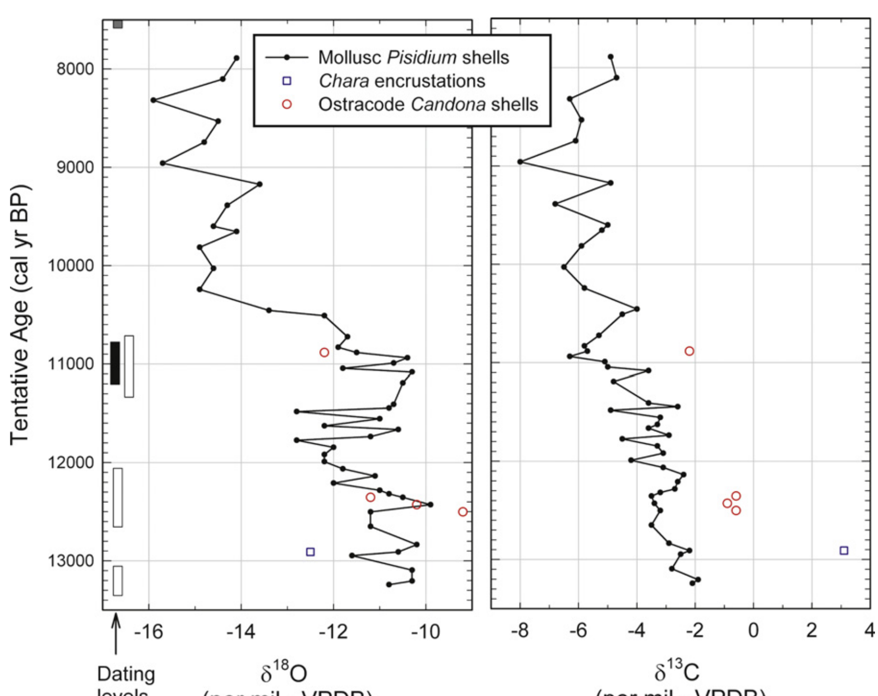

Fig. 6. Stable oxygen and carbon isotopes of Core 2 at Hundred Mile Lake, Alaska. Rectangles on the left side show the location of AMS dates: black = terrestrial plant macrofossils; gray = bulk organic sediment; and open = mollusc shells.

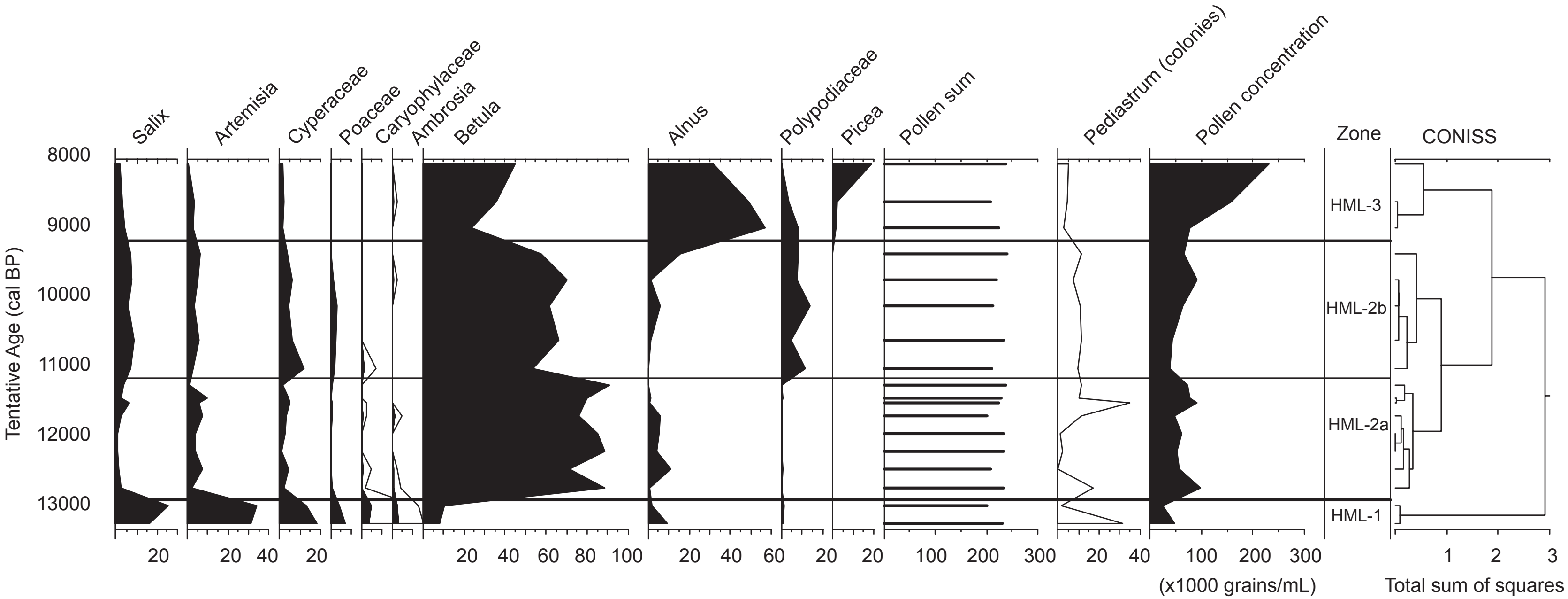

Fig. 7. Summary percentage pollen diagram from Hundred Mile Lake, Alaska.

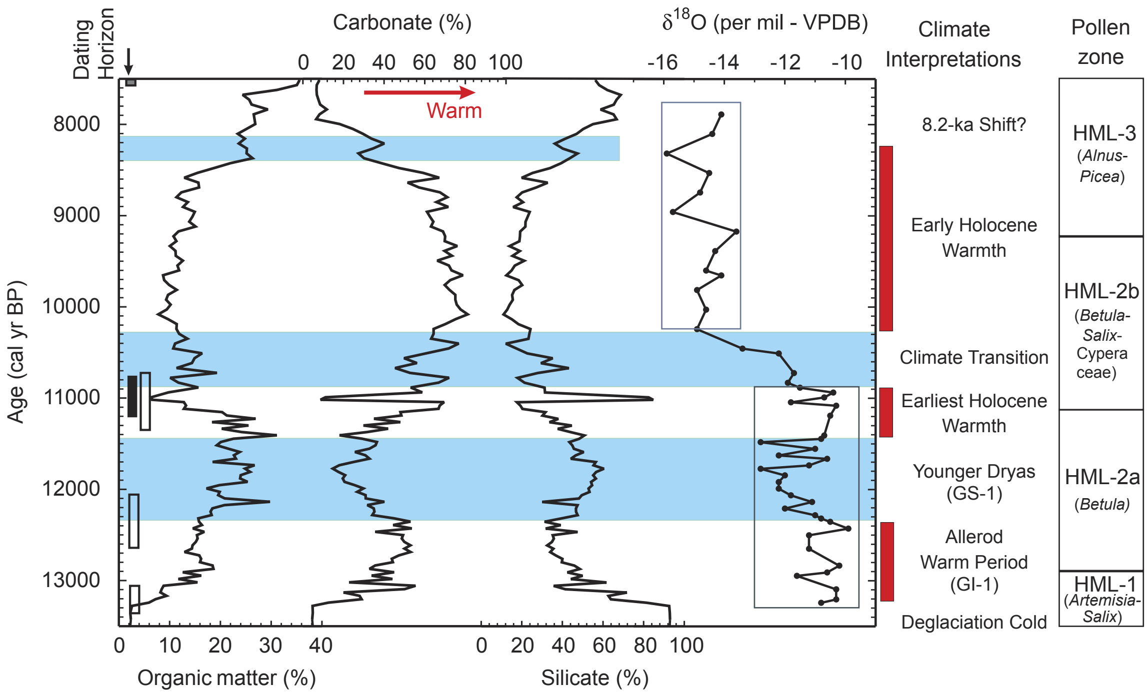

Fig. 8. Correlation of multiple proxies during the Lateglacial and early Holocene from Hundred Mile Lake and their climatic interpretations. Blue horizontal bands show the three major climate oscillations based on $\delta^{18}O$ shifts: the Younger Dryas (GS-1), early Holocene climate transition, and mid-Holocene climatic transition around 8.2 ka. Also shown are Allerød warm period (GI-1) and Early Holocene Thermal Maximum. Rectangles on the left side show the location of AMS dates: black = terrestrial plant macrofossils; gray = bulk organic sediment; and open = mollusc shells. INTIMATE event stratigraphy was based on Björck et al. (1998): Greenland Interstadial (GI)-1 (Bølling–Allerød) and Greenland Stadial (GS)-1 (Younger Dryas).

the Allerød warm period. The low carbonates and high silicates from 12.4 to 11.2 ka may represent the lithological response to the YD cooling event (GS-1 event; Björck et al., 1998), induced by reduced lake productivity and elevated erosion from the watershed under cold conditions. The increase in carbonate and decrease in silicates at the end of the YD (11.5 ka) indicate Holocene warming, and the subsequent peak carbonate between 10.5 and 8.5 ka may correspond with HTM in Alaska (Kaufman et al., 2004).

The major lithology shift from marl to gyttja between 8.5 and 8.2 ka indicates change in lake chemistry and/or lake level, possibly responding to 8.2 ka cooling event in the Northern Hemisphere (Alley et al., 1997). However, the shift at HML appears to cause permanent changes in lake conditions rather than short-lived oscillations documented elsewhere. The occurrence of gyttja throughout the upper core after 8 ka is likely the result of the lake deepening.

5.2. Paleoclimatic shifts inferred from stable isotopes in lacustrine carbonates

The $^{18}O/^{16}O$ ratios in freshwater carbonates, when in equilibrium, depend on the isotopic composition of lake water and on water temperature. In high latitudes, the isotopic signal of the water reflects atmospheric temperature (Siegenthaler and Eicher, 1986; Yu, 2000). So $\delta^{18}O$ values indicate air temperatures, precipitation sources and/or changes in local hydrology, potentially providing an independent proxy of climate change (Kelts and Hsu, 1978). The stable-isotope record from HML demonstrates considerable shifts in both $\delta^{18}O$ and $\delta^{13}C$ values. A negative excursion in $\delta^{18}O$ of ~2‰ occurred at 12.3 ka (Fig. 8), likely representing cooling corresponding to the YD climate reversal as documented in Alaska and elsewhere (Yu and Wright, 2001). The interval of high $\delta^{18}O$ values before the YD indicates warm climate during the Allerød. The absence of the Bølling warm period might be caused by the late retreat of local glacial ice and then delayed lake formation. The surprisingly large and dramatic shift of −4.5‰ in $\delta^{18}O$ values in the early Holocene (between 11 and 10.3 ka) suggests a major change in climate, which has not been documented elsewhere in the region (see below). Shortly after ice retreat HML was isolated hydrologically from Matanuska Glacier, as glacial moraines sit between the glacier and the lake and the lake is in a small and isolated catchment, about 50 m above the Matanuska River bed (lake surface elevation of 1661 feet vs. <1500 feet for the river). So the lake's hydrology and isotopes are very unlikely affected by glacial melt water after the deposition of these moraines.

The $\delta^{13}C$ values are a function of aquatic production, organic matter input, and decomposition. The $\delta^{13}C$ values of the HML core show a decrease from ca −2‰ to −8‰ (from 13.3 to 9 ka; Fig. 6), perhaps indicating an increased input of organic matter from the watershed as vegetation succeeded from treeless shrub *Salix*-herb tundra to *Betula*-dominated dense shrub tundra and *Alnus*-dominated forest (Fig. 7). Increased organic matter input would increase decomposition in the lake, causing the decline in $\delta^{13}C$ values. The lack of covariance before 10 ka suggests that the local hydrology did not play a significant role in driving the $\delta^{18}O$ shifts in carbonates (Talbot, 1990). Instead, the $\delta^{18}O$ shifts most likely reflected temperature changes. After 10 ka the stronger covariance suggests possible closed-basin conditions during lower lake level, which appears to be consistent with peak carbonate content during the HTM (Kaufman et al., 2004).

5.3. Regional vegetation history

There are several paleoecological studies in south-central Alaska, mostly from fossil pollen analysis of lake sediments and peat sections (see Fig. 1(B) for site locations). The only published paleo-record in the Matanuska Valley is from Kepler Lake, near the mouth of the Matanuska River. At Kepler Lake, Forester et al. (1989) derived a record of pollen and ostracode assemblages from 240 cm long calcareous sediments deposited during the last 2500 years and found a cold and dry climate period, possibly corresponding to the Little Ice Age. Outside the Matanuska Valley in south-central Alaska, Point Woronzorf (PW) peat section near Anchorage and Seventy Mile Lake northeast of the Chugach Mountains in the Copper River Basin are two other sites closest to the study region. During the last 12,000 radiocarbon years BP (equivalent to 13.8 ka) at PW, vegetation changed from birch–sedge herb tundra, through birch shrub tundra and mixed poplar–willow–alder vegetation zone, to birch–alder–spruce zone (Ager and Brubaker, 1985). At Seventy Mile Lake, vegetation changed from birch–sedge, through birch–willow–poplar, to spruce–alder-dominated forest (Ager and Brubaker, 1985). The longest pollen record from south-central Alaska was from Hidden Lake on the Kenai Peninsula, where the 14,500 radiocarbon year (17.5 ka) record shows vegetation succession from herb tundra, through birch–alder shrub tundra, to spruce-dominated forest (Ager, 1983).

At HML vegetation changed from herb tundra (>13.2–12.9 ka), through *Betula* shrub tundra (12.9–9.2 ka), to *Alnus* forest (9.2–8 ka). Before 13 ka, the vegetation around HML was largely dominated by herbaceous and shrub taxa. High percentages of *Salix*, Caryophyllaceae, *Artemisia*, *Ambrosia*, Cyperaceae, and Poaceae were present at 13 ka likely indicating sparse vegetation cover. Vegetation history at this site was similar to other sites in south-central Alaska following glacial ice retreat. This pollen assemblage includes pollen types of varying ecological preference indicating a tundra environment, or a varied environment from mesic to dry conditions. This landscape is similar to that found at Hidden Lake, Point Woronzof, and Seventy Mile Lake shortly after deglaciation (Ager and Brubaker, 1985). HML documents warm-loving plant types and increase in *Betula* after 13 ka. Ager and Brubaker (1985) also note this *Betula*-dominated zone at Hidden Lake, Point Woronzof, and Seventy Mile Lake.

The increase of Polypodiaceae after 11 ka is a widespread phenomenon in the region as similar increases in this taxon have been documented by Ager and Brubaker (1985) at Point Woronzof. By 9.2 ka vegetation composition at HML begins to shift from shrub–tundra to an *Alnus* forest landscape, with the arrival of *Picea*, which is in accordance with other regional records (Ager and Brubaker, 1985). Ager and Brubaker (1985) indicate that Hidden Lake, Point Woronzof, and Seventy Mile Lake all recorded a *Populus–Salix* zone prior to the appearance of the *Alnus–Betula–Picea* forest. However, HML does not contain a *Populus* zone in the pollen record.

5.4. Responses of watershed, lake, and vegetation to climate change

If we use $\delta^{18}O$ as a proxy of climate changes, then lithology and pollen can be used to indicate responses of lake, watershed, and vegetation to these climate changes (Wright, 1984). The highest $\delta^{18}O$ values (−10‰ to −11‰) between 13.3 and 12.4 ka possibly reflect initial warming after ice retreat. During this period the landscape and watershed begin to stabilize. The landscape consisted largely of shrub *Salix*–herb tundra. An increase in carbonate (close to 60%) also reflects stabilization in the watershed and lake. Around 12.5 ka a decrease in carbonate and $\delta^{18}O$ values represents the YD. Beginning around 11.4 ka carbonate begins increasing as does the $\delta^{18}O$ value, but they reach peak values at different times (Fig. 8). There is a possible delay of several hundred years in lake response to Holocene warming.

The marl to gyttja transition around 8 ka suggests an increase in lake level, due possibly to a climate shift to cool and wet conditions, which is consistent with other regional records (e.g., as reviewed in Bigelow and Edwards, 2001; Kaufman et al., 2004). For example, lake level increased around 8 ka at Birch Lake near Fairbanks in interior Alaska (Abbott et al., 2000). The final flooding of continental shelves might have brought moisture source close to the study region, which, together with reduced evaporation after the HTM, would have contributed to the increase in effective moisture. Vegetation responds similarly to a wetter environment, as an *Alnus–Betula*-dominated forest interspersed with *Picea* and herbaceous taxa dominate the landscape surrounding HML.

5.5. The Younger Dryas (GS-1) climate event and ice margin in south-central Alaska

Evidence for a YD-aged climate event has been accumulating in Alaska, but mostly restricted to the near coast regions. Engstrom et al. (1990) was the first study to demonstrate the presence of the YD in Alaska and on the Pacific Northwest coast. On Pleasant Island in the Glacier Bay area of southeastern Alaska, Engstrom et al. (1990) and Hansen and Engstrom (1996) described a pollen sequence implying that an established lodgepole pine

(*Pinus contorta*) parkland or forest reverted to shrub- and herb-dominated tundra (Ericaceae, Poaceae, Cyperaceae, and *Artemisia*) from 10,800 to 9800 ^{14}C BP. This vegetational change is matched by geochemical evidence of reduction of organic matter from catchment soils and increased mineral erosion (Engstrom et al., 1990). They interpreted these vegetational and geochemical changes as caused by a possible YD cooling event, because no known readvance of glaciers occurred at that time in the area. On Kodiak Island in southwestern Alaska, Peteet and Mann (1994) found distinct lithologic oscillations and a dramatic reversal in vegetation involving the near disappearance of Polypodiaceae (''Fern Gap'') from 10,800 to 10,000 ^{14}C BP, suggesting colder and drier conditions. Hajdas et al. (1998) performed high-resolution radiocarbon analysis from this site and found that the dramatic rise in atmospheric radiocarbon at the beginning of the YD interval coincided with the onset of 'Fern Gap'. They interpreted this reversal as a high-latitude Pacific expression of the YD event. Hu et al. (2002) presented a multi-proxy record from a lake in southwestern Alaska and found that the YD is indicated by vegetation change from shrub tundra to herb tundra at 13–11.6 ka, among other sedimentary evidence.

At HML, the YD is suggested by changes in lithology (decrease in carbonate, and increases in silicate and organic matter) and by a negative excursion of oxygen isotopes. The lithology change probably reflects changes in lake productivity and watershed stability. The 2‰ negative excursion in δ^{18}O at 12.4–11.5 ka may represent a decrease in air temperature of ~5 °C, if we use a simple relation of 0.36‰ per °C (e.g., Yu and Eicher, 1998) and assume all isotopic change caused by temperature change. Both lithology and oxygen isotopes show simultaneous shifts at the beginning of the YD, but lithology apparently lags behind the isotopic shift at the end of the YD, the beginning of the Holocene (Fig. 8). Also, another interesting aspect is that the YD appears to be a highly variable interval climatically, as has been increasingly indicated elsewhere (e.g., von Grafenstein et al., 1999; Ebbeson and Hald, 2004). The lack of pollen change associated with the YD may be due to either the coarse sampling resolution of pollen analysis or insensitivity of upland vegetation at that time.

HML lies just inside the outermost of a set of nested Matanuska Glacier moraines mapped by Williams and Ferrians (1961) and is located just 5 km down valley of the modern Matanuska Glacier terminus (Fig. 1(C)). Extrapolation of our age–depth model to the base of the lacustrine sediment provides a minimum deglaciation age of 14 ka for this site. In addition, ages of 14.1 and 15.5 ka (calibrated) from a peat and pond deposit at ''Glacier Bog'' (GB), located midway between HML and the modern Matanuska terminus (Fig. 1(C)), were reported by Williams (1986). These dates mean that the Matanuska Glacier had retreated from its Last Glacial Maximum position near Anchorage to very near its present position by ~15.5–14 ka. This also requires that the YD ice margin and end moraine, if there is one, is located up valley of HML and GB. If these ages are correct the YD ice margin must lie between these dated sections and the modern Matanuska margin, or up-glacier of the modern margin (Evenson et al., 2005). The ''up-glacier'' scenario is consistent with the emerging picture of early postglacial warmth in high northern latitudes far from the melting ice sheets, including peak temperatures before YD in parts of Alaska (Kaufman et al., 2004), in response to peak summer insolation near YD time. Any YD readvance of Matanuska Glacier may have been muted both because the YD signal was small so far from the North Atlantic center of action and because YD cooling in widespread regions was concentrated in wintertime and glaciers respond primarily to summer temperatures (Evenson et al., 2005).

5.6. *Possible causes of early Holocene climate shift around 10,500 cal BP*

The mechanisms that could possibly cause the 4.5‰ decline in δ^{18}O around 10.5 ka in the early Holocene include changes in air/water temperatures, meteoric precipitation seasonality, and a change in moisture source regions or trajectories. A positive correlation of +0.6‰ per °C exists between δ^{18}O values in meteoric precipitation and air temperature at middle and high latitudes (Dansgaard, 1964). If the shift in δ^{18}O values is caused by a decrease in mean annual air temperature, it would represent 7–8 °C cooling. If using −0.24‰ as the negative temperature-dependent fractionation between calcite precipitation and water temperature (Friedman and O'Neil, 1977), then a −4.5‰ shift in δ^{18}O would represent ~20 °C increase in water temperature. However, such a decrease in air temperature or an increase in water temperature are unrealistic and are not supported by any records in the region. If assuming water temperature closely follows the air temperature, then it would represent ~12 °C decrease in temperature. Also, the fact that the isotopic decline corresponds with an increase in carbonate content to its maximum value of 80% at ~10 ka rules out the possibility of a large decrease in temperature. So the 4.5‰ decrease in δ^{18}O values was unlikely caused entirely by air and water temperature changes.

The two nearest climate stations that have isotopic measurements of meteoric precipitations are Bethel in southwestern Alaska and Whitehorse in southern Yukon (Fig. 1(A) for location). Both stations show large seasonal change in oxygen isotope values, and winter precipitation has lower isotopic values than summer precipitation, with a difference of up to 10‰ in δ^{18}O between summer and winter months (Fig. 2). So a shift to more winter precipitation would have decreased δ^{18}O values in meteoric input water, lake water, and precipitated bivalve shells. Also, any moisture sources from the east would have much lower δ^{18}O values than from the west, with an average of 9‰ difference between Bethel and Whitehorse. Thus, a change in atmospheric circulation patterns to that effect would change the isotopic values in inflowing water to the lake. Anderson et al. (2005) attributed the observed carbonate isotope changes at a lake in southern Yukon during the middle and late Holocene to changes in location and intensity of the Aleutian low-pressure system.

The climate of south-central Alaska today is affected by large-scale seasonal atmospheric circulation. In winter months, a low-pressure system (Aleutian low) in the Gulf of Alaska is a dominant climate control (Mock et al., 1998). Counter-clockwise circulation around the cyclones associated with the Aleutian low-pressure system tends to deliver abundant moisture to southern coast of Alaska from the Gulf of Alaska. Due to the presence of coastal mountain ranges, limited amount of moisture reaches interior Alaska, and the Matanuska Valley is situated in a rain shadow on the northern side of Chugach Mountains. During summer months, cyclone activity associated with the Aleutian low weakens (Fleming et al., 2000), while the position of East Asian trough at the upper atmosphere (500 mb) and Pacific substropical high system at sea level are the important climate controls of Alaskan climate (Mock et al., 1998).

With present-day climate controls in mind, what would have caused the depletion in ^{18}O during warm and probably dry early Holocene around 10.5 ka? This warm and dry climate has been documented in arctic-wide (Kaufman et al., 2004) and regional synthesis of eastern interior Alaska (Edwards et al., 2001). Several circulation anomalies identified by Mock et al. (1998) seem to be able to explain the warm and dry climate with a possible decreasing δ^{18}O. During these anomalies, a weaker Pacific subtropical high in northeast Pacific Ocean would cause less westerly flow to south-central Alaska. An associated high-level positive anomalies north of Alaska would increase easterly flow around an anticyclone to the north. The easterly flow has high temperature in summer, owing to intensive heating of the land surface and distance from oceanic influence (see Fig. 1(A) for July temperature isotherm). Edwards et al. (2001) applied these circulation anomalies (increased ridging north of Alaska and weakened westerly circulation) to explain warm and dry climate around 9 ^{14}C ka (~10.2 ka) in eastern interior Alaska around Fairbanks as documented by pollen and lake-level records (e.g., Abbott et al., 2000). This change in atmospheric circulation appears to be supported by GCM simulation (Kutzbach et al., 1993). A weakened Pacific subtropical high would bring more frequent warmer and drier air from the east to the Matanuska Valley and we know the moisture from the east (e.g., from Whitehorse), despite limited amount, has much lower δ^{18}O values. Also, even though the dominant moisture was still from the Gulf of Alaska via the westerly flow, the more frequent extension of the dry easterly flow would shift the trajectories of water vapor coming from the Pacific Ocean or force the moist air to condense at higher altitude, inducing precipitation depleted in ^{18}O (e.g., Smith and Hollander, 1999). Thus, such a change in circulation pattern would explain the observed warm (high carbonate content) and lower δ^{18}O values around 10.5 ka.

A possible cause for the change in atmospheric circulation could be change in solar activity. Björck et al. (2001) identified a short cooling event at 10.3 ka from lacustrine, tree-ring, ice-core, and marine records around the north-eastern North Atlantic and suggested that a decrease in solar activity could be an important trigger for this broad and climate cooling event based on close correlation with solar proxies. If that is the case, then a decrease in solar forcing that caused cool North Atlantic could have triggered a shift in atmospheric circulation in the North Pacific region.

6. Conclusions and implication

(1) The multiple proxy data from HML in the Matanuska Valley of south-central Alaska provide complementary information for changes in vegetation and climate during the Lateglacial and early Holocene. The δ^{18}O values from bivalve *Pisidium* shells provide independent records for climate changes. A negative excursion of 2‰ in δ^{18}O values at 12.4–11.5 ka represents a regional expression of the YD climate reversal after the warm Allerød period during the last deglaciation. The magnitude and timing of the YD are similar to other records in Alaska as well as in the North Atlantic region. Our data suggest that Matanuska Glacier might have been insensitive to the YD climate event, with no apparent advance.

(2) During the early Holocene when carbonate content reached its maximum, δ^{18}O values shift dramatically from −10.5‰ to −15‰ at ~10.5 ka, which has not been documented elsewhere in Alaska. This large decline in δ^{18}O during this warm and probably dry period suggests a major shift in atmospheric circulation patterns. In particular, a weaker Pacific subtropical high and associated high pressure system north of Alaska would induce less westerly moisture transport and increased easterly air flow, causing warm and dry climate in south-central Alaska. Also, the moisture from the east has much lower δ^{18}O values than that from the west.

(3) The lithology change from clay to marl at 13.3 ka suggests a delayed response of the aquatic system to deglacial warming after ice retreat. Pollen data show vegetation changes from herb tundra, through *Betula* shrub tundra, to an *Alnus* forest. The increase in *Betula* at ~13 ka appears to lag behind climate warming by a few centuries. During the mid-Holocene around 8 ka a dramatic shift takes place in lithology from carbonate-rich marl to organic-rich gyttja sediments, indicating a climatic shift to cooler and wetter conditions, which are consistent with other regional records.

(4) The data from this study suggest that the climatic shift in the region during the early Holocene was in greater magnitude than the YD event, implying a differential

regional response in south-central Alaska to large-scale climate forcing, possibly caused by the stronger regional feedback processes in high-latitude regions.

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