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A biogenic-silica $\delta^{18}O$ record of climatic change during the last glacial–interglacial transition in southwestern Alaska

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Abstract

Despite growing evidence for environmental oscillations during the last glacial–interglacial transition from high latitude, terrestrial sites of the North Pacific rim, oxygen-isotopic records of these oscillations remain sparse. The lack of data is due partially to the paucity of lakes that contain carbonate sediment suitable for oxygen-isotopic analysis. We report here the first record of oxygen-isotopic composition in diatom silica ($\delta^{18}O_{Si}$) from a lake in that region. $\delta^{18}O_{Si}$ increases gradually from 19.0 to 23.5‰ between 12,340 and 11,000 ^{14}C yr B.P., reflecting marked climatic warming at the end of the last glaciation. Around 11,000 ^{14}C yr B.P., $\delta^{18}O_{Si}$ decreases by 1.7‰, suggesting a temperature decrease of 3.5–8.9 °C at the onset of the Younger Dryas (YD) in southwestern Alaska. Climatic recovery began ca. 10,740 ^{14}C yr B.P., as inferred from the increase of $\delta^{18}O_{Si}$ to a maximum of 23.9‰ near the end of the YD. Our data reveal that a YD climatic reversal in southwestern coastal areas of Alaska occurred, but the YD climate did not return to full-glacial conditions.

Keywords: Younger Dryas; Paleoclimate; Oxygen Isotopes; Lake Sediment; Alaska; North Pacific

Introduction

Abundant evidence exists from the circum-North Atlantic region that climatic warming at the end of the last glaciation was punctuated by abrupt reversals to glacial conditions, including the Younger Dryas (YD) (Mayle and Cwynar, 1995; Alley, 2000; Ammann et al., 2000). Major efforts have been made to quantify these climatic changes in that region on the basis of proxy records from ice cores and marine and lake sediments (e.g., Cuffey et al., 1995; Levesque et al., 1997; Severinghaus et al., 1998; Shemesh and Peteet, 1998). In the North Pacific region, recent studies also revealed intriguing evidence for the YD (e.g., Engstrom et al., 1990; Peteet and Mann, 1994; Hu et al., 1995, 2002; Brubaker et al., 2001; Briner et al., 2002). However, the nature and magnitude of these climatic changes remain poorly understood, which can be partially attributed to reliance on proxy indicators (e.g., sediment characteristics, pollen assemblages that lack modern analogs) that do not offer quantitative climatic information. Consequently, comparison of these abrupt climatic events between the North Atlantic and North Pacific regions is difficult, hampering our understanding of the mechanisms that regulate the Earth's climate system.

We present here the results of the first study of oxygen-isotopic composition in freshwater-diatom biogenic silica ($\delta^{18}O_{Si}$) from the high-latitude region of the North Pacific, where continental paleoclimatic reconstruction has been limited by the lack of lacustrine carbonate material suitable for oxygen-isotope analyses. The Grandfather Lake (50° 48'N, 158° 31'W, 142 m altitude, informal name; Fig. 1) record spans the last glacial–interglacial transition. A previous palynological study at that site showed marked vegetational changes possibly related to the YD, but conclusions were greatly compromised by chronological problems (Hu et al., 1995). We report here a new chronology constrained by five AMS ^{14}C dates on terrestrial plant macrofossils and one conventional ^{14}C date on bulk sediment. The $\delta^{18}O_{Si}$ record allows for quantitative estimates of tempera-

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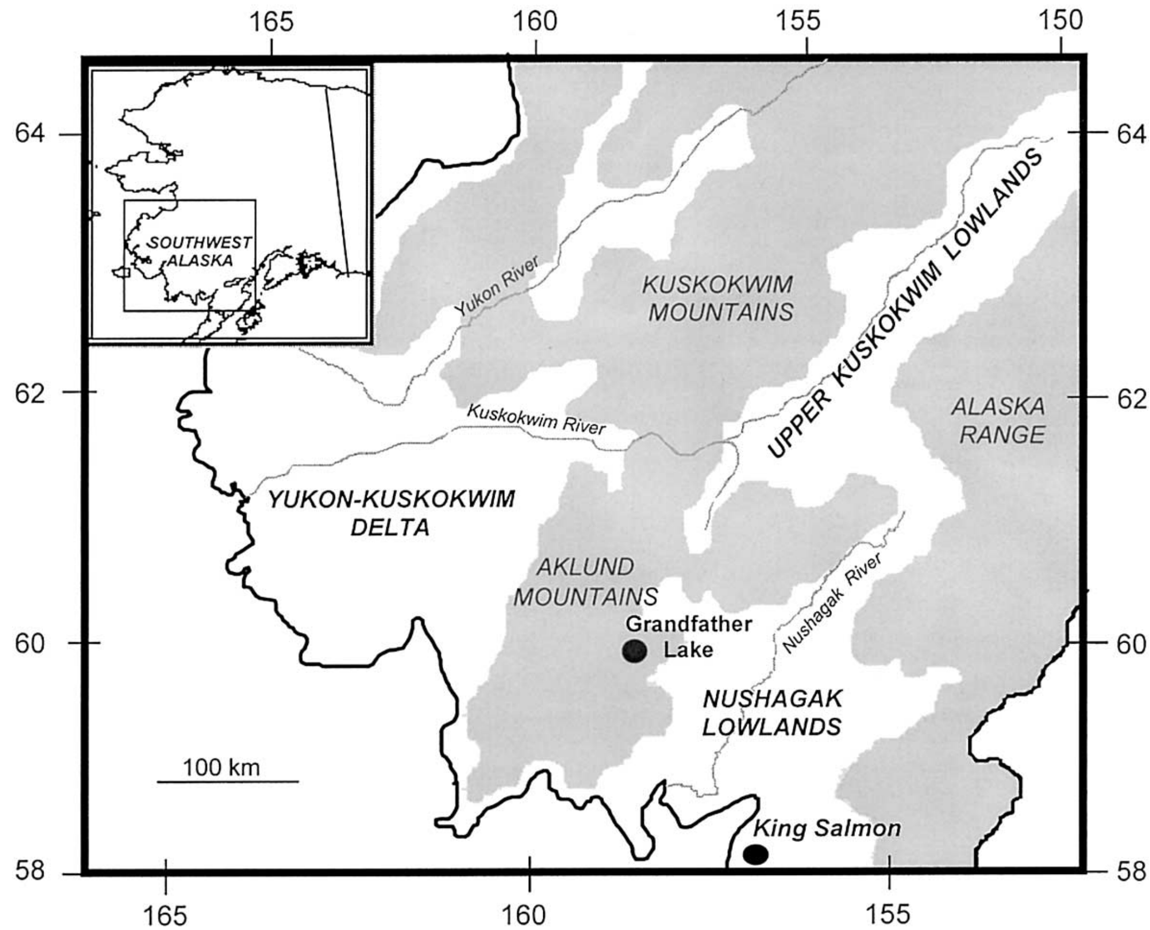

Fig. 1. Map of Alaska showing location of Grandfather Lake (modified from Brubaker et al., 2001).

ture changes, given certain assumptions, and offers new information on the characteristics of climatic oscillations, including those related to the YD, during the last glacial–interglacial transition.

Study site

Grandfather Lake has a surface area of ca. 0.35 km^2 and a maximum depth of 20 m. It is an upland lake located between the Ahklun Mountains and the western Nushagak Lowland (Fig. 1). During the late Wisconsinan, the area surrounding Grandfather Lake was covered by the eastern flank of the Ahlkun Mountains ice cap, with deglaciation probably by 15,000 ^{14}C yr B.P. (Briner et al., 2001; Manley et al., 2001; Levy, 2002). At present, Grandfather Lake appears to be a through-flow system with two inlet streams and one large outlet. Lakes in this region are typically ice-covered from October through May.

The regional climate is transitional between maritime and continental conditions. Mean annual temperature at King Salmon is ca. +0.7°C, with mean January and July temperatures of −10.3 and +12.5°C, respectively. Mean annual precipitation in lowland areas ranges from ca. 45 to 65 cm, with an average of 130–180 cm of snow per year. Permafrost is discontinuous in the lowlands and absent in the uplands. Modern vegetation in the watershed is mixed forest-tundra with dense *Alnus crispa* (green alder) thickets on hillsides and locally abundant stands of *Picea glauca* (white spruce).

Materials and methods

Two overlapping sediment cores (~12 m each) were recovered from the deepest part (20 m deep) of Grandfather Lake using a modified Livingstone piston corer (Wright et al., 1984). The two cores showed identical lithological features, and they were easily matched with magnetic-susceptibility measurements. Subsamples for the ^{14}C and $\delta^{18}O_{Si}$ analyses in this study were taken from the same core that was previously used for paleoecological reconstructions (Hu et al., 1995, 2001).

We washed numerous sediment samples at depth intervals of 0.5–1.0 cm through a 150-μm mesh sieve to concentrate plant macrofossils for ^{14}C dating. Plant macrofos-

Table 1
Radiocarbon dates,[a] Grandfather Lake

Depth (cm)	^{14}C Date (^{14}C yr B.P.)	Calibrated date[b] (^{14}C cal yr B.P.)	Laboratory number	Dated material
523	7820 ± 40	8543 (8594) 8631	CAMS 66726	Small twig
570	10,290 ± 50	11,780 (11,998) 12,104	CAMS 61476	Small twig
588	10,540 ± 50	12,352 (12,635) 12,829	CAMS 60817	Small twig
669	11,030 ± 160	12,893 (13,010) 13,161	CAMS 61477	Small twig
671	10,970 ± 50	12,893 (12,989) 13,124	CAMS 61478	Small twig
830–840	12,870 ± 50	14,484 (15,503) 15,809	Beta 59071	Bulk sediment

[a] All dates except at 830–840 cm are AMS dates. The 830–840 cm date is from Hu et al. (1995).
[b] Midpoints in parentheses and 1-sigma ranges.

sils (all twigs of shrubs) from five stratigraphic levels were pretreated with an acid–base–acid protocol at the University of Illinois and then submitted to Lawrence Livermore National Laboratory for AMS ^{14}C analysis. Throughout the remainder of the text, we use ^{14}C ages for convenience of comparison with Hu et al. (1995). For reference, we provide the calibrated ages (before 1950) converted from the ^{14}C ages (Table 1) with CALIB 4.3 (http://depts.washington.edu/qil/calib/calib.html).

The Grandfather Lake sediment is void of carbonate. We analyzed diatom-derived biogenic silica (BSi) for oxygen-isotopic composition. The $\delta^{18}O_{Si}$ technique (Shemesh et al., 1992) originally applied in marine records has been used recently in continental paleoclimatic studies (Rietti-Shati et al., 1998; Shemesh and Peteet, 1998; Rosqvist et al., 1999; Barker et al., 2001; Rioual et al., 2001; Shemesh et al., 2001). $\delta^{18}O_{Si}$ analysis followed the procedure of Shemesh et al. (1995). The basic steps of sample preparation are as follows: (1) separation of a pure diatom fraction from the sediment, (2) oxidation of organic matter associated with diatomaceous silica, (3) controlled oxygen isotope exchange (Juillet-Leclerc and Labeyrie, 1987), and (4) BSi-oxygen extraction by fluorination. The extracted O_2 is converted to CO_2 for isotope analysis on a Finnigan MAT-250 mass spectrometer. The results are calibrated *versus* the NBS-28 quartz international standard and are reported in the δ-notation vs VSMOW with a long-term reproducibility of 0.14‰.

Chronology

The new ^{14}C dates from Grandfather Lake are all in stratigraphic order, and we use linear interpolation to derive ages of the samples between ^{14}C dates (Table 1; Fig. 2). The ages of 11,030 ± 160 and 10,970 ± 50 ^{14}C yr B.P. at 669 and 671 cm, respectively, are statistically the same with an average of 11,000 ± 105 ^{14}C yr B.P. These two dates provide a solid constraint for the onset of the YD. The ages of 10,290 ± 50 and 10,540 ± 50 ^{14}C yr B.P. at 570 and 588 cm, respectively, fall within the YD chronozone. No plant macrofossils were found between 523 (7820 ± 40 ^{14}C yr B.P.) and 570 cm, thereby limiting chronological control for the end of the YD. We also did not find plant macrofossils below 671 cm. Thus we use the bulk-sediment conventional ^{14}C date of 12,870 ± 210 ^{14}C yr B.P. at 830–840 cm to derive ages for samples below 671 cm. This date was considered reliable because of its consistency with ages of postglacial peaty paleosols and basal thaw-lake deposits in the region (Hu et al., 1995).

These new ^{14}C ages indicate that the tentative chronology of Hu et al. (1995) was generally correct for the onset of a pollen-inferred YD, but that chronology underestimated the ages of the subsequent samples that were older than 9200 ^{14}C yr B.P. (Fig. 2). This new chronology confirms the speculation that some of the bulk-sediment dates (which were removed from the calculation of the age model in Hu et al., 1995) were too old. However, no consistent offset exists between the AMS dates on plant macrofossils and conventional dates on bulk sediments (Fig. 2). For example,

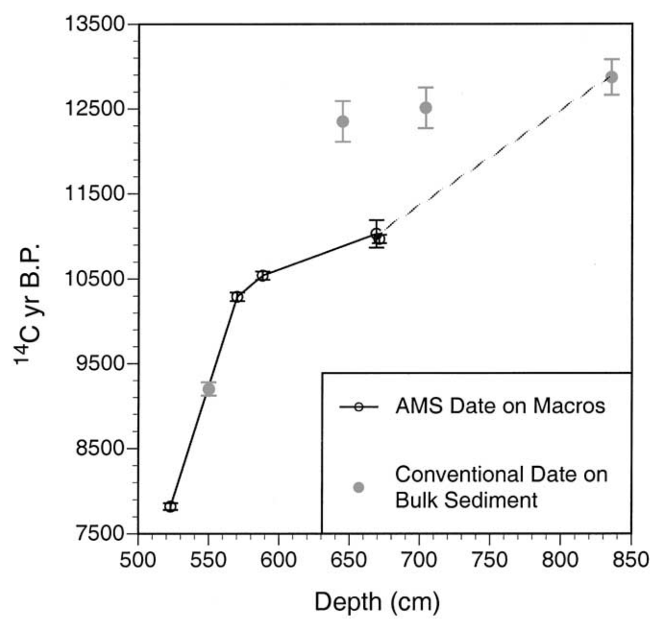

Fig. 2. Age–depth relationship of the late-glacial and early-Holocene section of the Grandfather Lake sediment. Open circles are new ^{14}C AMS dates on terrestrial plant macrofossils. Closed circles are conventional ^{14}C dates on bulk sediments from the same core reported in Hu et al. (1995).

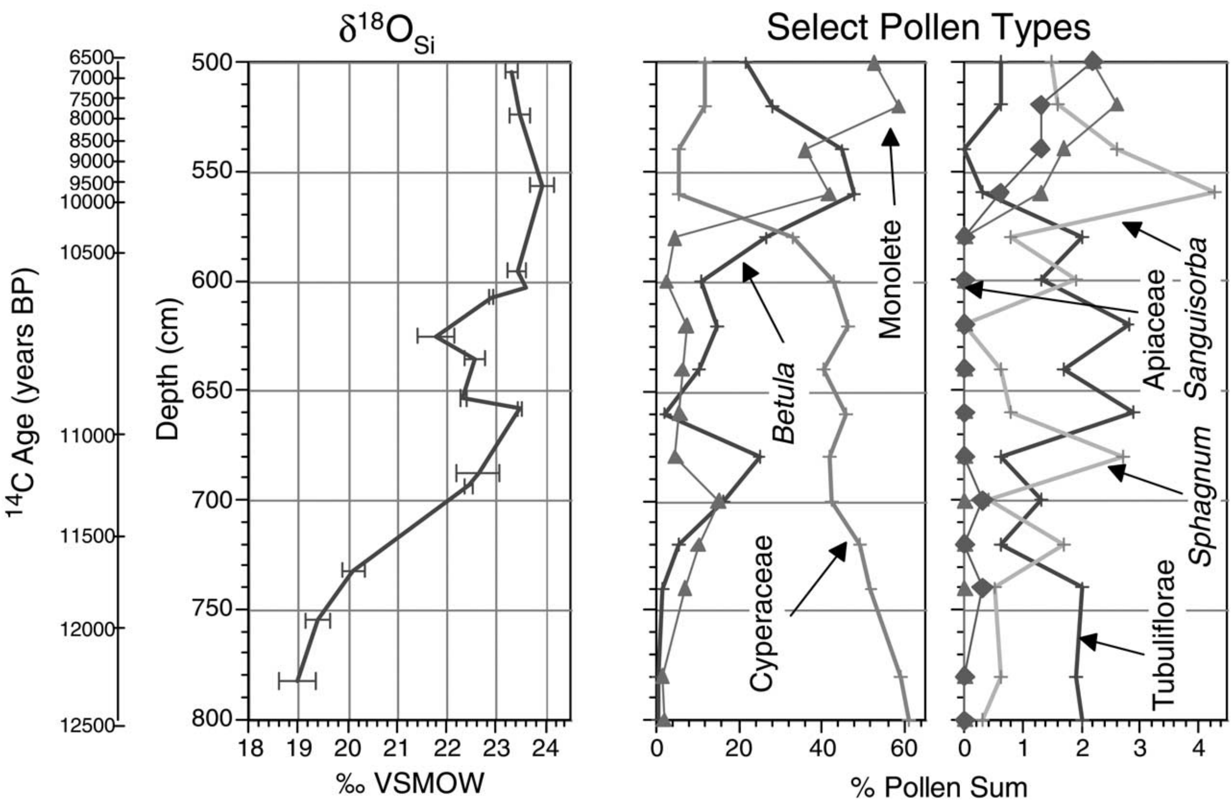

Fig. 3. Oxygen-isotopic composition of biogenic silica in comparison with select pollen profiles, Grandfather Lake. Error bars are based on duplicate analyses at each level.

in contrast to the large discrepancy near the onset of the YD, the bulk-sediment age of 9200 ± 80 ^{14}C yr B.P. falls on the line defined by the two adjacent AMS ages, suggesting that this bulk date is reasonable. We do not know why old carbon occurs in the Grandfather Lake sediment. The watershed bedrock is carbonate-free, making a "hard-water effect" an unlikely cause. One possible cause is the long-residence time of organic matter in the soils of high-latitude regions (Abbott and Stafford, 1996; Michaelson et al., 1996).

Climatic inferences from oxygen-isotopic composition of biogenic silica

The $\delta^{18}O_{Si}$ record from Grandfather Lake shows marked stratigraphic variations during the last glacial–interglacial transition (Fig. 3). $\delta^{18}O_{Si}$ increases gradually from 19.0 to 23.5‰ between 12,340 and 11,000 ^{14}C yr B.P. Around 11,000 ^{14}C yr B.P., $\delta^{18}O_{Si}$ decreases by 1.7‰ to reach a minimum of 21.8‰ at 10,740 ^{14}C yr B.P. $\delta^{18}O_{Si}$ then begins to increase to reach a maximum of 23.9‰ ca. 10,000 ^{14}C yr B.P. It exhibits a slight decreasing trend of 0.6‰ from 10,000 to 7500 ^{14}C yr B.P.

In high-latitude regions, diatom blooms that produce the biogenic silica for the $\delta^{18}O_{Si}$ analysis occur during summer, when lakes are ice-free with a nutrient supply from the watershed. Thus $\delta^{18}O_{si}$ captures summer environmental conditions, if the lake is a through-flow system with a short water residence time (Rosqvist et al., 1999; Shemesh et al., 2001). $\delta^{18}O_{Si}$ reflects both the temperature and the oxygen-isotopic composition ($\delta^{18}O_w$) of the lake water. $\delta^{18}O_w$ is determined by the regional temperature dependency of precipitation $\delta^{18}O$ and by the hydrological setting of the lake, the latter of which is, in turn, influenced by evaporation, moisture source, and mixing with regional ground water. Most of these factors are insufficiently known for the early-postglacial period of Grandfather Lake, making it difficult to tease out their effects on $\delta^{18}O_w$. Although we do not have isotopic data for the modern water of Grandfather Lake, the $\delta^{18}O$ and δ^2H values of water samples from six other lakes in southwestern Alaska all fall in the range of precipitation $\delta^{18}O$ and δ^2H values defining the local meteoric water line (Fig. 4; data are from the International Atomic Energy Agency–GNIP database for Bethel, Alaska). This pattern suggests the lack of substantial evaporative ^{18}O enrichment of lakes in that region, probably because summers are cool with a positive moisture balance under the near-coastal climate of the region. Similar observations have been made by recent studies at some high-latitude lakes in Swedish Lapland and South Georgia Island in the Southern Ocean (Rosqvist et al., 1999; Shemesh et al., 2001).

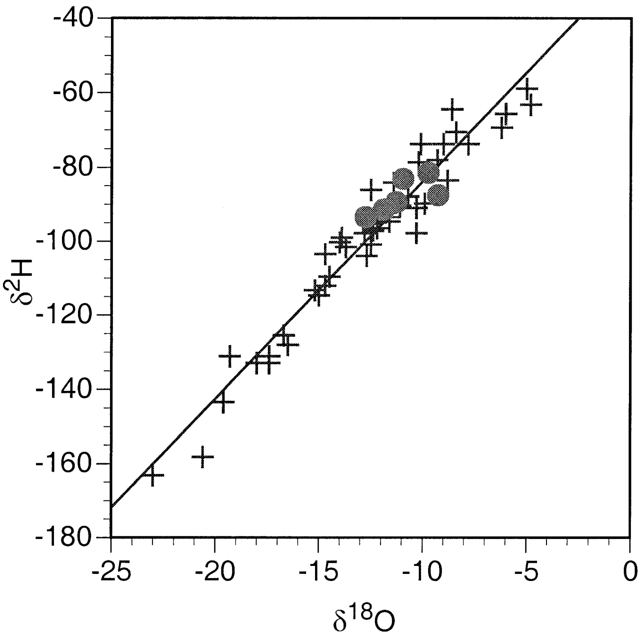

Fig. 4. δ^{18}O–δ^2H relationship of precipitation as defined by data (crosses) from Bethel, Alaska, from 1962 to 1970. These data are from the International Atomic Energy Agency—Global Network for Isotopes in Precipitation (http://isohis.iaea.org/GNIP.asp). Solid circles are isotopic composition of modern surface water samples from lakes in southwestern Alaska.

The precipitation δ^{18}O and instrumental weather data from Bethel show a relationship of 0.69‰/°C between atmospheric temperature and precipitation-$\delta^{18}O_{Si}$ ($r^2 = 0.29$, $n = 55$, $p < 0.005$), similar to that based on the global data set (0.65‰/°C; Rozanski et al., 1993). The water-temperature dependency of oxygen-isotopic fractionation is ca. -0.2 to -0.5‰/°C for the formation of diatom silica (Juillet-Leclerc and Labeyrie, 1987; Shemesh et al., 1992; Brandriss et al., 1998). The integration of these two factors results in a $\delta^{18}O_{si}$-temperature relationship of ca. 0.19–0.49‰/°C. By applying this relationship, we estimate a decrease of summer temperature by ~3.5–8.9 °C during the YD in the Grandfather Lake area. With the same assumptions, the steady $\delta^{18}O_{si}$ increase of 4.5‰ prior to 11,000 ^{14}C yr B.P. would be equivalent to an increase of ~9.2–23.7°C in summer temperature, and the decrease of 0.6‰ 10,000–7500 ^{14}C yr B.P. to a ~1.2–3.2°C decrease in summer temperature.

We caution that our temperature estimates are tentative, and the higher ends of these estimates are probably unrealistic. The great uncertainties of these estimates reflect the large range of water-temperature-dependent isotopic fractionation in diatom-silica formation. Currently, we do not have any means to better constrain the two temperature-dependent fractionation factors and to separate the isotopic effects of water temperature *versus* ambient air temperature. Furthermore, inadequate knowledge of local paleohydrology makes a rigorous assessment of our assumptions impossible. For example, paleoecological data (Hu et al., 2002) suggest that the YD was colder and drier in southwestern Alaska. Lake levels possibly were much lower than at present, making Grandfather Lake a hydrologically closed basin during the late glaciation and early Holocene. Under such conditions, our estimates of temperature fluctuations would be conservative, because drier conditions would have caused ^{18}O enrichment of lake water, dampening the effects of ^{18}O depletion induced by decreased temperature. However, no major change in $\delta^{18}O_{si}$ exists at the termination of the YD when the most pronounced effective-moisture increase occurred (Hu et al., 2002). This observation supports the assumption that hydrological changes were not an important factor controlling $\delta^{18}O_{si}$ at this particular site.

Other factors might also compromise the reliability of our temperature estimates. We assume that the $\delta^{18}O_{si}$ variations in the Grandfather Lake record were not caused by changes in diatom assemblages. We do not have diatom-assemblage data to directly test this assumption, although a species effect on $\delta^{18}O_{si}$ is not known to exist (Shemesh et al., 1995, 2001; Rioual et al., 2001). Variations in the amount of ^{18}O-depleted glacial meltwater draining to the lake could also have resulted in major $\delta^{18}O_{si}$ shifts. In particular, the post-YD $\delta^{18}O_{si}$ decrease may reflect the increased input of glacial meltwater to the lake. However, glaciers in the Grandfather Lake catchment should have long disappeared by the early Holocene (Briner et al., 2001; Manley et al., 2001). The seasonality of precipitation is another possible factor controlling $\delta^{18}O_{si}$. For example, the low $\delta^{18}O_{si}$ values during the YD could have resulted from an increased proportion of water input from spring snowmelt relative to summer rain. In addition, changes in moisture source and storm trajectory could have also exerted a major influence on $\delta^{18}O_w$. The flooding of the Bering Strait was probably an important factor in that context, but we cannot assess this factor until the sea level history of Bering Strait is better confined.

Nevertheless, our $\delta^{18}O_{si}$-inferred temperature changes are supported by pollen data from the same core (Hu et al., 1995). The replacement of herb tundra by *Betula* shrub tundra coincides with ^{18}O enrichment of local precipitation prior to 11,000 ^{14}C yr B.P. Recent greenhouse experiments and historic photographic documentation show that increased growing-season temperatures caused the expansion of *Betula* shrubs in arctic tundra (Chapin et al., 1995; Sturm et al., 2001). Thus, pollen assemblages confirm our interpretation of postglacial climatic warming prior to 11,000 ^{14}C yr B.P. On the basis of our new ^{14}C chronology, the reversal of *Betula* shrub tundra to herb tundra, as inferred from the decline of *Betula* pollen and increases of Cyperaceae and Tubuliflorae, indeed occurred at the onset of the YD ca. 11,000 ^{14}C yr B.P. (Hu et al., 1995; Fig. 3). This vegetational reversal suggests a climatic cooling around 11,000 ^{14}C yr B.P., supporting our climatic inference based on $\delta^{18}O_{si}$. The decline in $\delta^{18}O_{Si}$ appears to lag behind that of *Betula* percentages ca. 11,000 ^{14}C yr B.P., although the temporal resolution of our pollen and $\delta^{18}O_{Si}$ analyses is inadequate for a detailed comparison of the two proxies. Within the YD, low $\delta^{18}O_{Si}$ values broadly correspond to low *Betula* pollen percentages. Both *Betula* percentages and $\delta^{18}O_{Si}$ begin to increase within the YD, reaching maximum values near the end of the YD ca. 10,000 ^{14}C yr B.P. Pollen data indicate a subsequent decline in *Betula* shrubs, probably caused by the climatic cooling as inferred from the $\delta^{18}O_{Si}$ decrease of 0.6‰ ca. 10,000–7500 ^{14}C yr B.P. Furthermore, our temperature reconstructions at Grandfather Lake seem plausible in the context of other paleorecords from Alaska. For example, several recent studies have revealed new evidence for climatic oscillations during the last glacial–interglacial transition (e.g., Epstein, 1995; Bigelow and Edwards, 2001; Briner et al., 2002; Hu et al., 2002; Mann et al., 2002), some of which offer strong support for a YD-like climatic reversal.

An intriguing feature in the $\delta^{18}O_{si}$-inferred climatic record from Grandfather Lake is that the magnitude of our inferred temperature rise 12,340–11,000 ^{14}C yr B.P. appears much greater than those of temperature fluctuations during the YD chronozone. This pattern suggests that the regional climate during the YD did not return to full-glacial conditions in southwestern Alaska. Pollen data from the region (Brubaker et al., 2001; Hu et al., 2002) also suggest that the YD climatic cooling did not result in conditions as severe as during the last glaciation. Whereas *Betula* was typically absent in the full-glacial pollen spectra in Alaska (e.g., Anderson et al., 1994), there existed substantial fluctuations in the abundance of *Betula* shrubs within the YD, including values similar to those of the pre-YD warm period (Fig. 3; Hu et al., 2002). On the basis of reconstructed equilibrium-line altitudes in the Ahklun Mountains, southwestern Alaska, Briner et al. (2002) estimated that the extent of glacial advance during the YD was 25–40% of that during the full glaciation. These data stand in contrast with evidence from the North Atlantic region where the YD climate was characterized by near full-glacial conditions (e.g., Cuffey et al., 1995; Severinghaus et al., 1998). This contrast implies that global/hemispheric forcings induced heterogeneous responses on the regional scale. Quantitative documentation of such regional responses is essential to understanding the causal mechanisms of abrupt climatic events.

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