



Abrupt climatic events during the last glacial-interglacial transition in Alaska

Feng Sheng Hu,^{1,2} David M. Nelson,² Gina H. Clarke,¹ Kathleen M. Rühland,³ Yongsong Huang,⁴ Darrell S. Kaufman,⁵ and John P. Smol³

Received 16 June 2006; revised 10 August 2006; accepted 16 August 2006; published 28 September 2006.

[1] Evidence is mounting that abrupt climatic shifts occurred during the last glacial-interglacial transition (LGIT) in the North Atlantic and other regions. However, few high-resolution climatic records of the LGIT exist from the high latitudes of the North Pacific rim. We analyzed lake sediments from southwestern Alaska for biogenic silica, organic carbon, organic nitrogen, diatom assemblages, and compound-specific hydrogen isotopes. Results reveal climatic changes coincident with the Younger Dryas, Intra-Allerød Cold Period, and Pre-Boreal Oscillation. However, major discrepancies exist in the paleoclimate patterns of the Bølling-Allerød interstadial between our data and the GISP2 $\delta^{18}\text{O}$ record from Greenland, and causes are uncertain. These data suggest that the North Pacific and North Atlantic experienced similar reversals during climatic warming of the LGIT but that the Bølling-Allerød cooling trend in the GISP2 $\delta^{18}\text{O}$ record is probably not a hemispheric or global pattern.

Citation: Hu, F. S., D. M. Nelson, G. H. Clarke, K. M. Rühland, Y. Huang, D. S. Kaufman, and J. P. Smol (2006), Abrupt climatic events during the last glacial-interglacial transition in Alaska, *Geophys. Res. Lett.*, 33, L18708, doi:10.1029/2006GL027261.

1. Introduction

[2] A number of recent studies documented climatic fluctuations during the LGIT in the high latitudes of the North Pacific [e.g., *Briner et al.*, 2002; *Engstrom et al.*, 1990; *Hu et al.*, 2002; *Peteet and Mann*, 1994]. These studies unveiled compelling evidence that the region experienced a Younger Dryas (YD)-type climatic reversal. However, the specific climatic manifestations (e.g., temperature versus moisture) and geographic extent of this reversal remain uncertain. In addition, none of the published proxy-climate records have contiguous sampling, and many of them are compromised by the lack of secure chronologies [e.g., *Hu et al.*, 1995]. These problems, along with the slow sedimentation rates typical of high-latitude lakes, have made it impossible to compare the detailed structures of climatic dynamics during the LGIT between the North Pacific and North

Atlantic. For example, it is unknown if and how the Bølling-Allerød interstadial and the centennial-scale climatic reversals (e.g., Intra-Allerød Cold Period: IACP), which are well documented in the circum-North Atlantic, were manifest around the North Pacific. As a result, hemispheric connections and mechanistic implications of these climatic events are poorly understood.

[3] We report here a multiproxy record of climatic and ecosystem changes during the LGIT at two lakes (Arolik Lake: 59°28'N, 161°07'W, 145 m asl; Ongoke Lake: 59°15'N, 159°25'W, 200 m asl; Figure 1) in southwestern Alaska. We analyzed sediment cores at contiguous 1-cm intervals for biogenic silica (BSi) at both lakes and for organic carbon (OC) and nitrogen (ON) at Arolik Lake. Our chronologies, based on multiple AMS ^{14}C ages of terrestrial plant macrofossils at each site, indicate that the average resolutions of these data are ~ 40 and ~ 26 years for Arolik and Ongoke lakes, respectively. We also analyzed select samples from Arolik Lake for diatom assemblages and hydrogen-isotope ratios of bound palmitic acid to help constrain our climatic and ecological interpretations.

2. Materials and Methods

[4] Arolik Lake is in a moraine-dammed basin located along the southwestern flank of the Ahklun Mountains near Bristol Bay (Figure 1). The lake lies outside of the late-Wisconsin glacial limits [*Manley et al.*, 2001]. Sediment cores were recovered using a percussion corer. The data presented in this paper are all from core AL4, taken from a sub-basin with a water depth of 18 m. Ongoke Lake is located in the southwestern flank of the Wood River Mountains. The area was glaciated during the last glacial maximum [*Manley et al.*, 2001]. The lake had a maximum water depth of ~ 7 m in June 2004. Sediment cores were retrieved using a modified Livingstone corer from an extensive flat area at a water depth of 5 m.

[5] Sediment samples from both sites were washed with distilled water through a 150- μm mesh to concentrate terrestrial plant macrofossils for ^{14}C dating (Figure 2; see auxiliary material¹ for ^{14}C ages and related information). For Arolik Lake, we used the chronology of *Hu et al.* [2003]. For Ongoke Lake, five ^{14}C ages were converted to calibrated ages with CALIB 5.0.2, and a locally weighted regression model [*Cleveland and Devlin*, 1988] was fit through the calibrated ages. We emphasize that, as typical of LGIT studies, our chronologies suffer from ^{14}C plateaus and have

¹Departments of Plant Biology and Geology, University of Illinois at Urbana-Champaign, Urbana, Illinois, USA.

²Program in Ecology and Evolutionary Biology, University of Illinois at Urbana-Champaign, Urbana, Illinois, USA.

³Paleoecological Environmental Assessment and Research Laboratory, Department of Biology, Queen's University, Kingston, Ontario, Canada.

⁴Department of Geological Sciences, Brown University, Providence, Rhode Island, USA.

⁵Departments of Geology and Environmental Sciences, Northern Arizona University, Flagstaff, Arizona, USA.

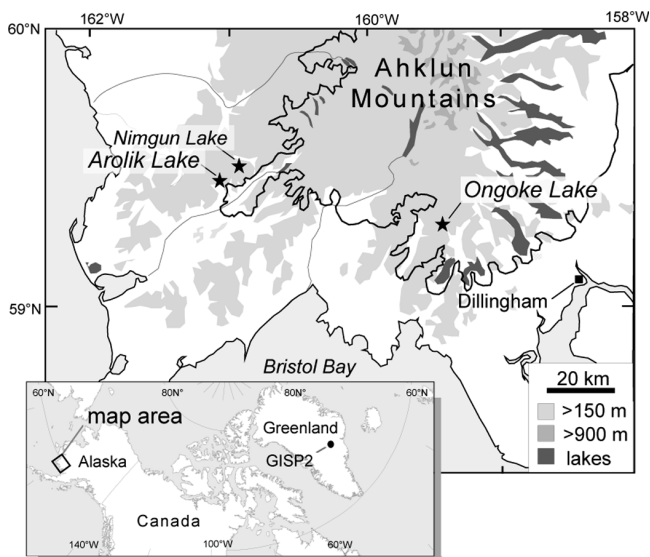


Figure 1. Map showing the locations of Arolik and Ongoke lakes as well as other sites mentioned in the main text. Solid line around Ahklun Mountains depicts the ice cap extent during the last glacial maximum on the basis of work by *Manley et al.* [2001].

error ranges of up to 1000 years in certain intervals (see Figure 2 for the 2-sigma error ranges of calibrated ages).

[6] BSi was extracted with 10% Na_2CO_3 and determined with a spectrophotometer (Spectronic Genesys5) as described by *Mortlock and Froelich* [1989]. Most of the Arolik BSi data were included by *Hu et al.* [2003], which focused on the Holocene section of the same core and contained a brief discussion on the LGIT. For OC and ON measurements, freeze-dried samples were analyzed with a Costech Instruments 4010 elemental combustion system.

[7] For diatom-assemblage analysis, ~ 0.2 g of wet sediment from each level was treated with H_2O_2 to remove organic material [*Battarbee et al.*, 2001]. Small aliquots of the sediment slurry were strewn onto glass coverslips, dried, and then mounted onto microscope slides using Naphrax[®]. A minimum of 400 diatom valves were counted for each sample using a Leica DMR light microscope. Principal component analysis was used to summarize the main floristic variation. The species scores were divided by the standard deviation with linear rescaling so as to most accurately display the relationship between species and the sample scores from

principal component axes 1 (PC1) and 2 (PC2). Diatom diversity was calculated using the N2 statistic [*Hill*, 1973], which estimates the effective number of abundant taxa in each sample.

[8] For hydrogen-isotopic analysis, palmitic acid was isolated from the bound lipid fractions of sediments by accelerated solvent extraction [*Huang et al.*, 2002]. Bound lipids were released by saponifying the solvent-extracted sediments with 0.5 N KOH/MeOH (with ~ 2 – 3% water) for 1 hr. The acid fraction was then isolated by solid phase extraction, followed by purification using silica gel column chromatography. The concentration of compounds was determined with GC-FID, and identification was made with GC-MS. An HP 6890+ GC-pyrolysis system interfaced to a Finnigan Delta+ XL stable isotope spectrometer through a high-temperature pyrolysis reactor was used for hydrogen isotopic analysis [*Hilkert et al.*, 1999; *Huang et al.*, 2002].

3. Results and Discussion

3.1. Climatic Reversals During the Last Glacial Termination

[9] The profiles of BSi concentrations from both sites show pronounced variations (Figure 3). Although some of these variations may be related to lake or watershed processes that may not be directly linked to climatic change, coherent patterns across sites can be used to decipher past climatic conditions, especially when verified with other proxy indicators. At Arolik Lake, BSi increases from 85 to 240 mg/g between 16000 and ~ 13500 cal BP. It remains high but with a decreasing trend until ~ 13200 cal BP when it drops abruptly to reach a minimum of 85 mg/g at ~ 13100 cal BP. BSi recovers to a peak value of 160 mg/g at ~ 12900 cal BP and then drops abruptly again to minimal values by ~ 12800 cal BP. It increases afterwards to reach peak values of ca. 280 mg/g at ~ 11000 cal BP with small reversals, including one at ca. 11400–11200 cal BP.

[10] The BSi profile from Ongoke Lake exhibits stratigraphic patterns generally similar to those at Arolik Lake. However, before 11500 cal BP, the major shifts appear to predate those at Arolik Lake by up to a few hundred years. In addition, some of the temporal variations are more muted at Ongoke Lake. For example, the abrupt BSi decrease around 13200 cal BP at Arolik Lake appears to occur around 13300 cal BP with a reduced magnitude at Ongoke Lake. The chronological differences of the major shifts between the two sites likely reflect ^{14}C dating errors instead of intra-regional

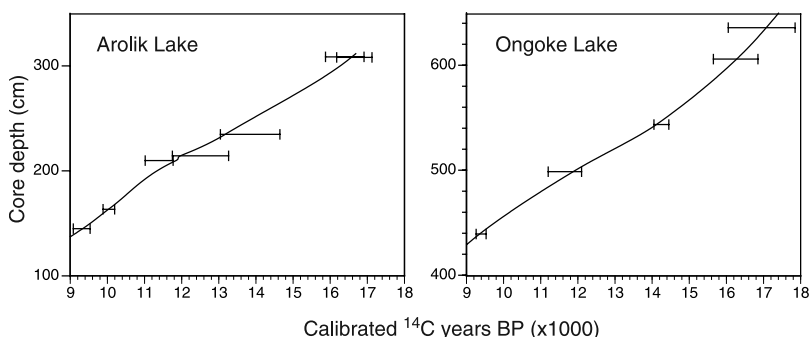


Figure 2. Age-depth models of the sediment sections spanning the last glacial-interglacial transition from Arolik and Ongoke lakes. Error bars indicate 2-sigma ranges (95% confidence intervals) of calibrated ^{14}C ages.

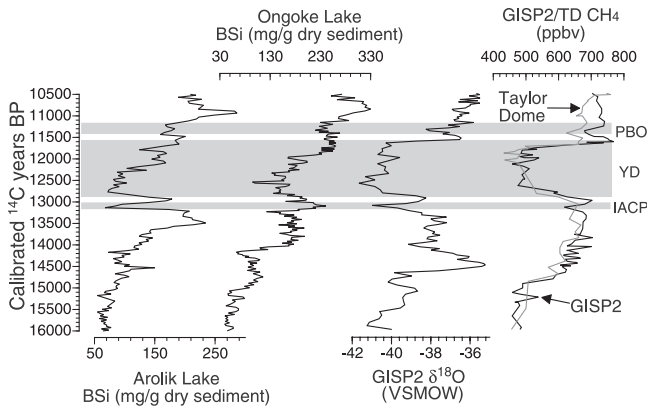


Figure 3. Biogenic-silica record of the last glacial-interglacial transition from Arolik and Ongoke lakes in comparison with $\delta^{18}\text{O}$ and CH_4 data from Greenland Ice Sheet Project 2 (GISP2) and CH_4 data from Taylor Dome (TD). BSi: Biogenic silica (mg SiO_2 per g dry bulk sediment); PBO: Pre-Boreal Oscillation; YD: Younger Dryas; IACP: Intra-Allerød Cold Period.

climatic heterogeneity, and the more muted signals at Ongoke Lake are not unexpected because this lake is substantially shallower and thus sediment mixing was probably more extensive than at Arolik Lake. In the following discussion, we focus on Arolik Lake where data of other proxy indicators complementary to BSi are available.

[11] BSi primarily reflects the abundance of diatoms, which are single-celled algae commonly dominating the primary productivity of high-latitude lakes [Antoniades et al., 2005]. Comparison of BSi with diatom $\delta^{18}\text{O}$, compound-specific δD , and pollen assemblages in the Holocene sediments of Arolik Lake indicates that aquatic productivity was positively related to growing-season temperature and effective moisture [Hu et al., 2003]. The BSi variations ca. 16000–10500 cal BP suggest major fluctuations in lake productivity. At Arolik Lake, this interpretation is supported by stratigraphic changes in the OC:ON ratio (Figure 4),

which mirror those of BSi (Figure 4). This relationship confirms that BSi reflects aquatic productivity, as the OC:ON ratio is typically lower for organic matter of aquatic origin than that of terrestrial sources [Meyers and Lallier-Vergés, 1999].

[12] The abrupt decline in BSi (Figure 3) and increase in OC:ON ratio (Figure 4) around 13000 cal BP suggest that lake productivity decreased dramatically. This aquatic change coincides with the onset of the YD. These data are consistent with previous findings that a YD reversal, characterized by climatic cooling and decreased moisture availability, occurred in southwestern Alaska [Briner et al., 2002; Hu et al., 1995, 2002; Hu and Shemesh, 2003] and that this reversal resulted in a major decrease in aquatic productivity [Hu et al., 1995, 2002]. All of the previous records were based on relatively coarse-resolution analyses.

[13] In addition to evidence for the YD onset, our results reveal two short-lived climatic events that are well known in the circum North Atlantic region. The sharp decrease in aquatic productivity around 13200 cal BP, as evidenced by changes in BSi and OC:ON with patterns similar to those of the YD onset (Figures 3 and 4), coincide with the IACP in the GISP2 ice-core $\delta^{18}\text{O}$ record (Figure 3) [Alley et al., 1993; Stuiver et al., 1995] and the Gerzensee/Killarney event in the lake-sediment records of the North Atlantic [e.g., Levesque et al., 1993; Schwander et al., 2000]. A second brief event, centered at 11300 cal BP, coincides with the Pre-Boreal Oscillation (PBO) [Bjorck et al., 2002; Stuiver et al., 1995], although the magnitudes of BSi and OC:ON changes are small relative to those of the YD and IACP (Figure 3).

[14] The relatively coarse temporal resolution of our diatom record (Figure 4) does not allow for detailed comparisons with the profiles of BSi and OC:ON. Nonetheless, the diatom-assemblage trends support our inferences of aquatic productivity from BSi and OC:ON. In particular, variations in diatom composition (PC2) and diversity (Hill's N_2) follow those in BSi, particularly in the lower portion of the core. PC2 is largely driven by the relative abundance of *Stephanodiscus* taxa, a common planktonic indicator of nutrient-rich water (mainly phosphorus) [Hall and Smol, 1999], and by the

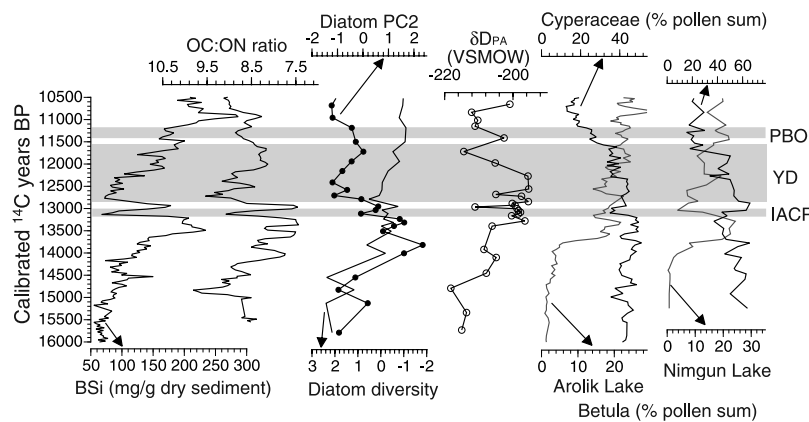


Figure 4. Records of biogenic silica, OC:ON, principal component axis 2 (PC2) and diversity of diatom assemblages, hydrogen-isotopic composition of palmitic acid ($\delta\text{D}_{\text{PA}}$), and select pollen types. All data are from Arolik Lake unless indicated otherwise. The OC:ON ratio is presented as 3-point running averages to highlight stratigraphic trends as the raw data are noisy. Note that the scales of OC:ON and diatom diversity are reversed. Diatom PC2 is driven primarily by *Stephanodiscus parvus*, *Achnanthes pusilla*, *Fragilaria brevistriata*, *Fragilaria construens*, *Fragilaria pseudoconstruens*. See Hu et al. [2002] for detailed description of pollen data from Nimgun Lake.

relative abundance of small *Fragilaria* taxa near the top of the section. PC2 values are relatively low before ca. 14500 cal BP. During this period, planktonic and oligotrophic *Cyclotella* dominate the diatom assemblages, and heavily silicified, tycho planktonic *Aulacoseira* are abundant, indicating a well-mixed water column. PC2 values are generally high between 14500 and 13000 cal BP, reflecting increases in primary productivity and nutrient cycling and suggesting longer ice-free periods, stronger water-column stratification, and higher water temperatures than previously [Douglas and Smol, 1999]. Similar to BSi and OC:ON ratio, PC2 shows marked decreases around 13100 and 12900 cal BP, corresponding to the IACP and the onset of the YD. PC2 values increase in the later stages of the YD to reach a second peak around 11600 cal BP, probably resulting from climatic warming and accelerated nutrient cycling. However, in the post-YD section of the record, PC2 decreases substantially as a result of marked decreases in *Fragilaria* and increases in *Achnanthes* and *Aulacoseira*. In addition, our sample resolution is inadequate to capture any potential diatom-assemblage shifts related to the PBO. Stratigraphic changes in diatom diversity are similar to those of PC2, implying positive relations among algal community complexity, nutrient availability, and primary productivity.

[15] Hydrogen-isotopic analysis of palmitic acid (Figure 4) helps constrain our climatic inferences from the biological proxies [Hu et al., 2003; Huang et al., 2002, 2004; Shuman et al., 2004]. A preliminary calibration study in Alaska shows that surface-sediment δD_{PA} is positively related to lake-water δD [Y. Huang and F.S. Hu, unpublished data], demonstrating the potential of δD_{PA} as a climatic proxy in that region. δD_{PA} increases by 20‰ from ~16000 to ~13300 cal BP, and the trend is broadly similar to those of BSi, inverse of OC:ON, and diatom-PC2. If we assume a temperature-dependent hydrogen isotopic fractionation of 5‰/°C [Estep and Hoerling, 1980] and that no evaporative effect existed, the δD_{PA} increase of ~20‰ is equivalent to a 4°C increase in atmospheric temperature. This estimate is probably conservative. In southwestern Alaska, warmer climate was probably associated with wetter conditions during the LGIT [Hu et al., 2002, 2003], and the magnitude of 2H enrichment from increasing temperature 16000–13000 cal BP may have been dampened by reduced evaporation. After 13300 cal BP, the stratigraphic patterns of δD_{PA} are generally opposite to those of BSi and diatom-PC2, and similar to those of OC:ON. δD_{PA} values are highest of the entire record between 13300 and 12300 cal BP. The most parsimonious explanation is that extremely arid conditions resulted in 2H enrichment ca. 13300–12300 cal BP, after which moisture availability increased markedly. This explanation is consistent with the fact that Arolik Lake is topographically a nearly closed basin at present [Kaufman et al., 2003] and that its isotope hydrology should be highly sensitive to shifting moisture balance. It is also consistent with a pollen-based moisture reconstruction at another site in southwestern Alaska [Hu et al., 2002].

[16] The triple reversals at Arolik Lake are similar to those well documented in the North Atlantic region, as exemplified by the $\delta^{18}O$ record of the Greenland Ice Sheet Project 2 (GISP2) [Stuiver et al., 1995] (Figures 1 and 3). This similarity implies that, as in the North Atlantic region, climatic warming during the last-glacial termination was

punctuated by repeated reversals around the North Pacific rim. To our knowledge, the two short-lived reversals before and after the YD have never been reported from the North Pacific high latitudes. Recent GCM simulations indicate that the YD cooling in the North Atlantic was transmitted through the atmosphere and the oceans to cause climatic deterioration in the North Pacific [e.g., Mikolajewicz et al., 1997]. Similar processes might have induced the IACP and PBO short-lived climatic reversals in the North Pacific.

3.2. Patterns of the Bølling-Allerød Climate: Discrepancies with GISP2

[17] Comparison of the Arolik Lake data with the GISP2 $\delta^{18}O$ record reveals major discrepancies during the Bølling-Allerød interstadial (B-A) between 14600 and 12900 cal BP. In particular, rapid climatic warming occurred at the Bølling onset around the North Atlantic [Stuiver et al., 1995; von Grafenstein et al., 1999] and some other regions [Lea et al., 2003]. In contrast, no evidence for such warming exists in southwestern Alaska based on our records. During the classic Bølling interstadial (~14600–14000 cal BP), BSi and diatom PC2 increase whereas the OC:ON ratio decreases gradually, suggesting a gradual increase in aquatic productivity. A major increase in aquatic productivity, as evidenced by a sharp increase in BSi, occurred around 13600 cal BP at Arolik Lake and 14000 cal BP at Ongoke Lake, in the middle of the Allerød interstadial in GISP2. Similarly, the expansion of *Betula* shrubs occurred around 13600 cal BP at Arolik Lake (Figure 4), likely resulting from increased temperature and moisture. Thus both terrestrial and aquatic ecosystem changes suggest that climatic amelioration in southwestern Alaska lagged behind that of the North Atlantic during the last-glacial termination. Although we cannot completely rule out the possibility of a chronological error, the observed delay is not unique to Arolik and Ongoke lakes. A delay of climatic amelioration by ~1000 years appears to have occurred at Nimgun Lake (Figure 1) in the same region [Hu et al., 2002].

[18] One possible cause for the delayed changes in southwestern Alaska is that a marked climatic warming occurred at the onset of the Bølling but ecosystem response was deterred by moisture deficits. Such conditions would have prevented tundra shrub expansion as well as reduced soil decomposition and nutrient flux to the lake, thereby limiting aquatic productivity. However, our δD_{PA} data (Figure 4) argue against the existence of warm and dry conditions during the Bølling period. Both warm conditions and evaporative enrichment associated with aridity should lead to high δD_{PA} values. On the contrary, the δD_{PA} values of the samples falling within the Bølling are similar to or lower than those of subsequent samples. Thus if the region was dry, these δD_{PA} values would argue that the atmospheric temperatures were low during the Bølling.

[19] In addition to the possible delay of postglacial warming in Alaska, the GISP2 $\delta^{18}O$ profile suggests progressive climatic cooling during the B-A, whereas our data suggest a warming trend. Discrepant B-A climatic trends with that of GISP2 have been observed in other regions as well, including mid-Europe [von Grafenstein et al., 1999], the North Sea and Norwegian Sea [Haflidason et al., 1995; Keigwin and Jones, 1995], and the tropical Atlantic [Lea et al., 2003]. These records imply that, in contrast to the abrupt

climatic reversals, the B-A cooling trend documented in the GISP2 $\delta^{18}\text{O}$ record is likely not a hemispheric or global pattern. Discrepancies also exist among proxy indicators of different processes during the B-A. For example, in contrast to the GISP2 $\delta^{18}\text{O}$ record, CH_4 concentrations in the GISP2 and Taylor Dome ice cores increased or remained high during the B-A (Figure 3).

[20] A recent study [Schaefer *et al.*, 2006] suggests that climatic warming at the last glacial termination in the North Atlantic was delayed by an extensive sea ice cover because of hypercold winters [Denton *et al.*, 2005]. Thus climatic changes during the B-A in that region were not necessarily synchronous with global events. Our lake-sediment data from southwestern Alaska also show evidence of sustained cold conditions during the LGIT, which may be related to other regional-scale drivers, such as the timing of the Bering Strait flooding. Elucidating the causes of these spatial and temporal discrepancies remains a challenge in the investigations of abrupt climatic changes during the LGIT.

[21] **Acknowledgments.** This research was supported by NSF (ATM 9996064 and 0318404 to FSH, and EAR 9808593 to D.K.) and NSERC (to J.P.S.) grants. We thank Y. Axford, J. Briner, I. Walker, and A. Werner for field assistance, S. Yoneji, J. Bright, and S. McMillan for laboratory assistance, and an anonymous reviewer for constructive comments.

References

- Alley, R. B., *et al.* (1993), Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event, *Nature*, *362*, 527–529.
- Antoniades, D., M. S. V. Douglas, and J. P. Smol (2005), Quantitative estimates of recent environmental changes in the Canadian High Arctic inferred from diatoms in lake and pond sediments, *J. Paleolimnol.*, *33*, 349–360.
- Battarbee, R. W., V. J. Jones, R. J. Flower, N. G. Cameron, H. Bennion, L. Carvalho, and S. Juggins (2001), Diatoms, in *Tracking Environmental Change Using Lake Sediments*, vol. 3, *Terrestrial, Algal, and Siliceous Indicators*, edited by J. P. Smol, H. J. B. Birks, and W. M. Last, pp. 155–202, Springer, New York.
- Björck, S., O. Bennike, P. Rosén, C. S. Andresen, S. Bohncke, E. Kaas, and D. Conley (2002), Anomalously mild Younger Dryas summer conditions in southern Greenland, *Geology*, *30*, 427–430.
- Briner, J. P., D. S. Kaufman, A. Werner, M. Caffee, L. Levy, W. F. Manley, M. R. Kaplan, and R. C. Finkel (2002), Glacier readvance during the late glacial (Younger Dryas?) in the Ahklun Mountains, southwestern Alaska, *Geology*, *30*, 679–682.
- Cleveland, W. S., and S. J. Devlin (1988), Locally weighted regression: An approach to regression analysis by local fitting, *J. Am. Stat. Assoc.*, *83*, 596–610.
- Denton, G. H., R. B. Alley, G. C. Comer, and W. S. Broecker (2005), The role of seasonality in abrupt climate change, *Quat. Sci. Rev.*, *24*, 1159–1182.
- Douglas, M. S. V., and J. P. Smol (1999), Freshwater diatoms as indicators of environmental change in the High Arctic, in *The Diatoms: Applications for the Environmental and Earth Sciences*, edited by E. F. Stoermer and J. P. Smol, pp. 227–244, Cambridge Univ. Press, New York.
- Engstrom, D. R., B. C. S. Hansen, and H. E. Wright (1990), A possible Younger Dryas record in southeastern Alaska, *Science*, *270*, 455–458.
- Estep, M. F., and T. C. Hoerling (1980), Biogeochemistry of the stable hydrogen isotopes, *Geochim. Cosmochim. Acta*, *44*, 1197–1206.
- Hafliadason, H., H. P. Sejrup, D. K. Kristensen, and S. Johnsen (1995), Coupled response of the late-glacial climatic shifts of northwest Europe reflected in Greenland ice cores; Evidence from the northern North Sea, *Geology*, *23*, 1059–1062.
- Hall, R. L., and J. P. Smol (1999), Diatoms as indicators of lake eutrophication, in *The Diatoms: Applications for the Environmental and Earth Sciences*, edited by E. F. Stoermer and J. P. Smol, pp. 128–168, Cambridge Univ. Press, New York.
- Hilkert, A. W., C. B. Douthitt, H. J. Schluter, and W. A. Brand (1999), Isotope ratio monitoring gas chromatography/mass spectrometry of D/H by high temperature conversion isotope ratio mass spectrometry, *Rapid Commun. Mass Spectrom.*, *13*, 1226–1230.
- Hill, M. O. (1973), Diversity and evenness: A unifying notation and its consequences, *Ecology*, *54*, 427–432.
- Hu, F. S., and A. Shemesh (2003), A biogenic-silica $\delta^{18}\text{O}$ record of climatic change during the last glacial-interglacial transition in southwestern Alaska, *Quat. Res.*, *59*, 379–385.
- Hu, F. S., L. B. Brubaker, and P. M. Anderson (1995), Postglacial vegetation and climate-change in the northern Bristol Bay region, southwestern Alaska, *Quat. Res.*, *43*, 382–392.
- Hu, F. S., B. Y. Lee, D. S. Kaufman, S. Yoneji, D. M. Nelson, and P. D. Henne (2002), Response of tundra ecosystem in southwestern Alaska to Younger-Dryas climatic oscillation, *Global Change Biol.*, *8*, 1156–1163.
- Hu, F. S., D. Kaufman, S. Yoneji, D. Nelson, A. Shemesh, Y. Huang, J. Tian, G. Bond, B. Clegg, and T. Brown (2003), Cyclic variation and solar forcing of Holocene climate in the Alaskan subarctic, *Science*, *301*, 1890–1893.
- Huang, Y. S., B. Shuman, Y. Wang, and T. Webb (2002), Hydrogen isotope ratios of palmitic acid in lacustrine sediments record late Quaternary climate variations, *Geology*, *30*, 1103–1106.
- Huang, Y. S., B. Shuman, Y. Wang, and T. Webb (2004), Hydrogen isotope ratios of individual lipids in lake sediments as novel tracers of climatic and environmental change: A surface sediment test, *J. Paleolimnol.*, *31*, 363–375.
- Kaufman, D. S., F. S. Hu, J. P. Briner, A. Werner, B. P. Finney, and I. Gregory-Eaves (2003), A similar to 33,000 year record of environmental change from Arolik Lake, Ahklun Mountains, Alaska, USA, *J. Paleolimnol.*, *30*, 343–362.
- Keigwin, L. D., and G. A. Jones (1995), The marine record of deglaciation from the continental margin off Nova Scotia, *Paleoceanography*, *10*, 973–985.
- Lea, D. W., D. K. Pak, L. C. Peterson, and K. A. Hughen (2003), Synchronicity of tropical and high-latitude Atlantic temperatures over the last glacial termination, *Science*, *301*, 1361–1364.
- Levesque, A. J., F. E. Mayle, I. R. Walker, and L. C. Cwynar (1993), A previously unrecognized late-glacial cold event in eastern North America, *Nature*, *361*, 623–626.
- Manley, W. F., D. S. Kaufman, and J. P. Briner (2001), Pleistocene glacial history of the southern Ahklun Mountains, southwestern Alaska: Soil-development, morphometric, and radiocarbon constraints, *Quat. Sci. Rev.*, *20*, 353–370.
- Meyers, P. A., and E. Lallier-Vergés (1999), Lacustrine sedimentary organic matter records of Late Quaternary paleoclimates, *J. Paleolimnol.*, *21*, 345–372.
- Mikolajewicz, U., T. J. Crowley, A. Schiller, and R. Voss (1997), Modelling teleconnections between the North Atlantic and North Pacific during the Younger Dryas, *Nature*, *387*, 384–387.
- Mortlock, R. A., and P. N. Froelich (1989), A simple method for the rapid determination of biogenic opal in pelagic marine sediments, *Deep Sea Res., Part A*, *36*, 1415–1426.
- Peteet, D. M., and D. H. Mann (1994), Late-glacial vegetational, tephra, and climatic history of southwestern Kodiak Island, Alaska, *Ecoscience*, *1*, 255–267.
- Schaefer, J. M., G. H. Denton, D. J. A. Barrell, S. Ivy-Ochs, P. W. Kubik, B. G. Andersen, F. M. Phillips, T. V. Lowell, and C. Schlüchter (2006), Near-synchronous interhemispheric termination of the Last Glacial Maximum in mid-latitudes, *Science*, *312*, 1510–1513.
- Schwander, J., U. Eicher, and B. Ammann (2000), Oxygen isotopes of lake marl at Gerzensee and Leysin (Switzerland), covering the Younger Dryas and two minor oscillations, and their correlation to the GRIP ice core, *Palaeogeogr. Palaeoclim. Palaeoecol.*, *159*, 203–214.
- Shuman, B., P. Newby, Y. Huang, and T. Webb, III (2004), Evidence for the close climatic control of New England vegetation history, *Ecology*, *85*, 1297–1310.
- Stuiver, M., P. M. Grootes, and T. F. Braziunas (1995), The GISP2 $\delta^{18}\text{O}$ climate record of the past 16,500 years and the role of the Sun, ocean, and volcanoes, *Quat. Res.*, *44*, 341–354.
- von Grafenstein, U., H. Erlenkeuser, A. Brauer, J. Jouzel, and S. J. Johnsen (1999), A mid-European decadal isotope-climate record from 15,500 to 5000 years B.P., *Science*, *284*, 1654–1657.
- G. H. Clarke and F. S. Hu, Departments of Plant Biology and Geology, University of Illinois at Urbana-Champaign, Urbana, IL 61801, USA. (fshu@life.uiuc.edu)
- D. S. Kaufman, Departments of Geology and Environmental Sciences, Northern Arizona University, Flagstaff, AZ 86011, USA.
- D. M. Nelson, Program in Ecology and Evolutionary Biology, University of Illinois at Urbana-Champaign, Urbana, IL 61801, USA.
- K. M. Rühland and J. P. Smol, Paleocological Environmental Assessment and Research Laboratory, Department of Biology, Queen's University, Kingston, ON, Canada K7L 3N6.
- Y. Huang, Department of Geological Sciences, Brown University, Providence, RI 02912, USA.