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Little Ice Age Evidence from a South-Central North American Ice Core, U.S.A.

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Abstract

In the past, ice-core records from mid-latitude glaciers in alpine areas of the continental United States were considered to be poor candidates for paleoclimate records because of the influence of meltwater on isotopic stratigraphy. To evaluate the existence of reliable paleoclimatic records, a 160-m ice core, containing about 250 yr of record was obtained from Upper Fremont Glacier, at an altitude of 4000 m in the Wind River Range of south-central North America. The $\delta^{18}O$ (SMOW) profile from the core shows a −0.95‰ shift to lighter values in the interval from 101.8 to 150 m below the surface, corresponding to the latter part of the Little Ice Age (LIA). Numerous high-amplitude oscillations in the section of the core from 101.8 to 150 m cannot be explained by site-specific lateral variability and probably reflect increased seasonality or better preservation of annual signals as a result of prolonged cooler temperatures that existed in this alpine setting. An abrupt decrease in these large amplitude oscillations at the 101.8-m depth suggests a sudden termination of this period of lower temperatures which generally coincides with the termination of the LIA. Three common features in the $\delta^{18}O$ profiles between Upper Fremont Glacier and the better dated Quelccaya Ice Cap cores indicate a global paleoclimate linkage, further supporting the first documented occurrence of the LIA in an ice-core record from a temperate glacier in south-central North America.

Introduction

To date, ice-core records of climate from most alpine glaciers in mid-latitude locations such as the continental United States were considered unsuitable because of the influence of meltwater on chemical and isotopic stratigraphy (Wagenbach, 1989). Since 1988, a glaciological research program has been conducted on the glaciers in the Wind River Range of northwestern Wyoming (Naftz et al., 1991, 1992, 1994; Naftz, 1993; Naftz and Smith, 1993) to determine the existence of an ice-core record of climate that could be linked to other low- and mid-latitude ice-core records (Thompson et al., 1984, 1986, 1988a, 1988b, 1989; Thompson, 1992). In 1991, a continuous 160-m ice core was recovered from the Upper Fremont Glacier, altitude 4000 m, in the Wind River Range, Wyoming, U.S.A. (Fig. 1). This report describes the first successful reconstruction of an isotopic record of paleoclimate from a south-central North American ice core. The record provides evidence for abrupt climatic change during the mid-1800s (the termination of the Little Ice Age [LIA]) and establishes a global linkage with the $\delta^{18}O$ Standard Mean Ocean Water (SMOW) series from two ice-core records from the Quelccaya Ice Cap in South America.

Methodology

SAMPLE COLLECTION

Because the site was in a designated wilderness area, conventional, gasoline-powered drilling systems were not allowed. Instead, a solar-powered drill was used to obtain the ice-core samples (Naftz and Miller, 1992). The drill system consisted of 48 solar panels with a peak power output of 2900 W and produced maximum, short-term drilling rates of 0.13 m min^{-1}. A thermal drill was used during the entire 160-m length of the hole. This was the first use of solar power for remote-site ice coring in the continental United States. The drilling system was constructed and operated by the Polar Ice-Coring Office, Fairbanks, Alaska.

The length of the core barrel was 2 m and the diameter was approximately 9 cm. Ice cores were removed from the core barrel and placed in plastic core bags. The core bags were sealed, placed in plastic core tubes, and stored in snow vaults until removal from the site to a freezer truck via a 10-min helicopter flight.

SAMPLE PROCESSING

The ice-core samples were melted according to strict protocols to minimize sample contamination. Ice cores were cut into

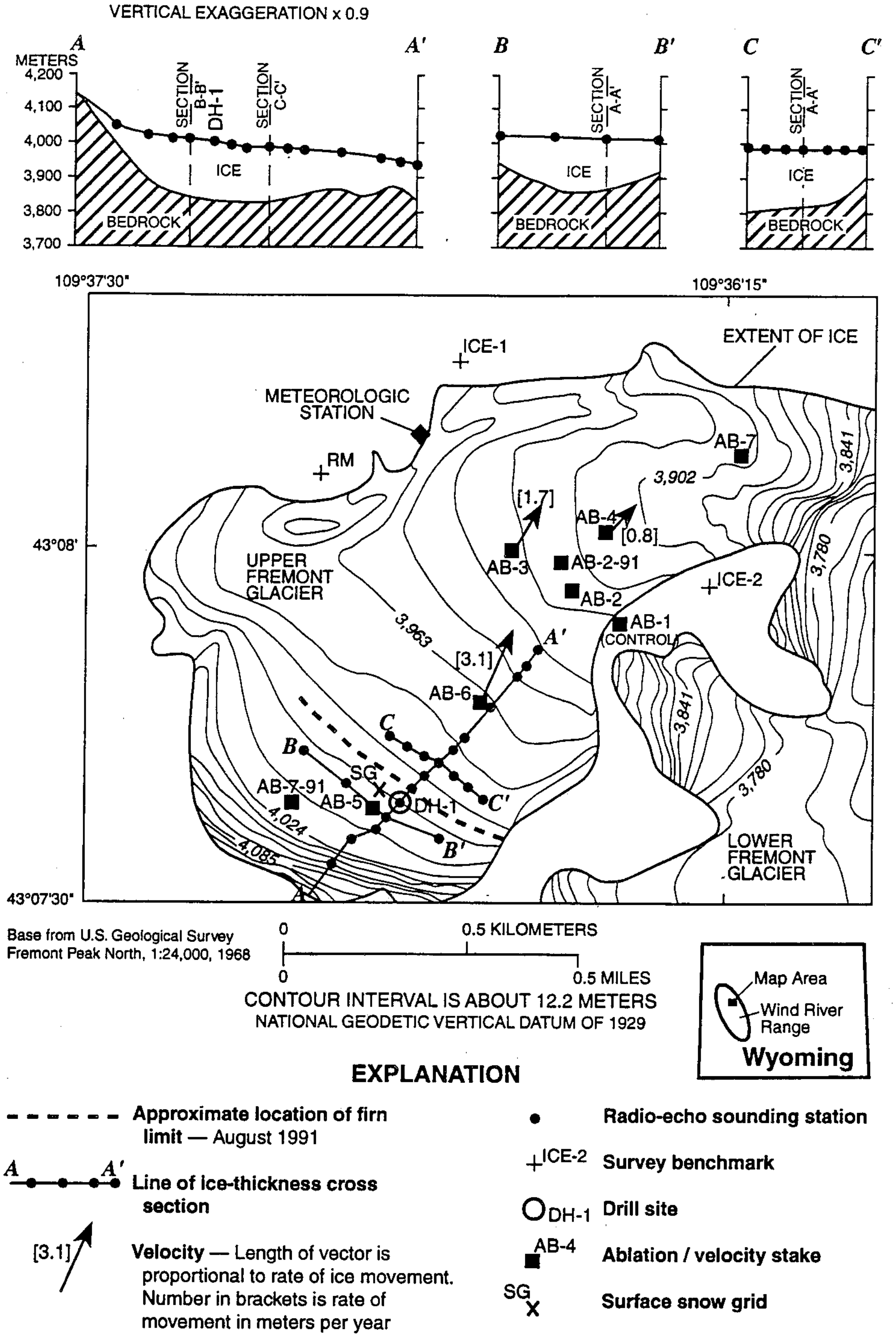

FIGURE 1. Location of firn limit, ice-thickness cross sections, radio-echo sounding stations, survey benchmarks, drill site, ablation/velocity stakes, surface snow grid, and meteorologic station, Upper Fremont Glacier, Wyoming.

20-cm sections using a bandsaw operated in a walk-in freezer (air temperature $<-10°C$). The 20-cm sections were split lengthwise with one split being archived for future research. The surface ice on the remaining split was scraped away using a stainless steel microtome. Each ice sample was thoroughly rinsed with ultrapure (17.8 megaohm) deionized water and placed in a prerinsed and covered plastic container. Each sample was allowed to melt at room temperature for 1 h. After this period, the sample was rinsed in the accumulated meltwater and the melt was discarded to eliminate any remaining isotopic signature from the rinse water. The remaining sample was allowed to melt at room temperature. After complete melting, the meltwater was filtered (0.45 μm), bottled, and preserved. The sample split analyzed for nitrate was placed in a polyethylene bottle and preserved by refrigeration. The sample split to be analyzed for $\delta^{18}O$ was placed in a glass vial with a polyseal cap and sealed with paraffin until analysis. The sample split used for tritium determination was placed in a glass vial until analysis.

SAMPLE ANALYSIS

Nitrate concentration of snow and ice samples was determined by ion exchange chromatography (Fishman and Friedman, 1989) using a Dionex Model QIC[1] ion chromatograph at the U.S. Geological Survey laboratory in Boulder, Colorado. The

[1]Use of brand names in this article is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey.

$\delta^{18}O$ value of each sample was determined using the method developed by Epstein and Mayeda (1953) at the U.S. Geological Survey Stable Isotope Laboratory in Menlo Park, California and reported relative to SMOW in the permil notation. Tritium concentration was determined by electrolytic enrichment/liquid scintillation counting (Thatcher et al., 1977) at the U.S. Geological Survey Tritium Laboratory in Reston, Virginia. Carbon-14 concentration in grasshopper leg parts entrapped in the ice was determined by accelerator mass spectrometry at a contract laboratory in New Zealand.

Results and Discussion

BACKGROUND GLACIOLOGICAL DATA

Glaciologic data in support of the ice-coring project were collected from 1990–91 (Naftz and Smith, 1993). Radio-echo sounding (Watts and England, 1977; Trombley, 1986) was used to determine ice thickness and bedrock topography (Fig. 1). Ice thickness ranged from 60 to 172 m in the upper half of the glacier. The 10-m borehole temperatures (4-d equilibration period) indicated that the ice is at the pressure melting point (0 ± 0.4°C). Annual ablation (including snow, firn, and ice) measured at locations on the glacier (Fig. 1) during 1990–91 averaged 0.93 m yr^{-1}. Densification processes proceed rapidly at the site, with densities exceeding 8.5×10^2 kg m^{-3} at depths 14 m below the surface. Ice velocity decreases in a downslope direction (Fig. 1), ranging from 0.8 to 3.1 m yr^{-1} during the monitoring period. Mean air temperature monitored on the glacier (Fig. 1) from 11 July 1990 to 10 July 1991 was -6.9°C. Snow-pit sampling at the site indicated preservation of the annual $\delta^{18}O$ signature during the initial summer melt season followed by dampening of the signal in subsequent melt seasons (Naftz, 1993; Naftz et al., 1993).

ICE-CORE CHRONOLOGY

Dust layers in the ice core were occasionally too faint to give reliable annual stratigraphic markers (Naftz, 1993); therefore, chemical age-dating techniques in combination with estimated accumulation and ablation rates were used to establish some approximate time horizons in the core for generalized climatic interpretations and comparisons. Tritium concentrations exceeding 300 tritium units at a depth 29 m below the surface corresponded to snowfall deposited in 1963 (Fig. 2) during the peak of above-ground nuclear bomb testing (Carter and Moghissi, 1977). The sharp tritium peaks at 29 and 41 m below the surface coupled with prebomb tritium concentrations in samples below a depth of 47.5 m, indicate that meltwater has not influenced the tritium profile.

Carbon-14 age dating of a grasshopper leg entrapped 152 m below the surface of the glacier (Fig. 3) indicates that ice near the bottom of the core formed from snow deposited between A.D. 1716 and 1820. The grasshopper leg yielded an age of 221 ± 95 Libby ^{14}C yr BP. Accounting for the changes of ^{14}C content in the atmosphere (Stuiver and Pearson, 1993) the following possible time windows and their associated probabilities were obtained using the program CALIB 3.0.1 (Stuiver and Reimer, 1993): A: (A.D. 1633–1704 [0.28]), B: (A.D. 1716–1820 [0.42]), and C: (A.D. 1916–1955 [0.16]). Time window C was eliminated because the size and morphology of the leg is entirely consistent with specimens of *Melanoplus spretus* (Lockwood, written comm., 1994), which became extinct in A.D. 1903 (Lockwood et al., 1994). Time window A appears unlikely because this age would not agree

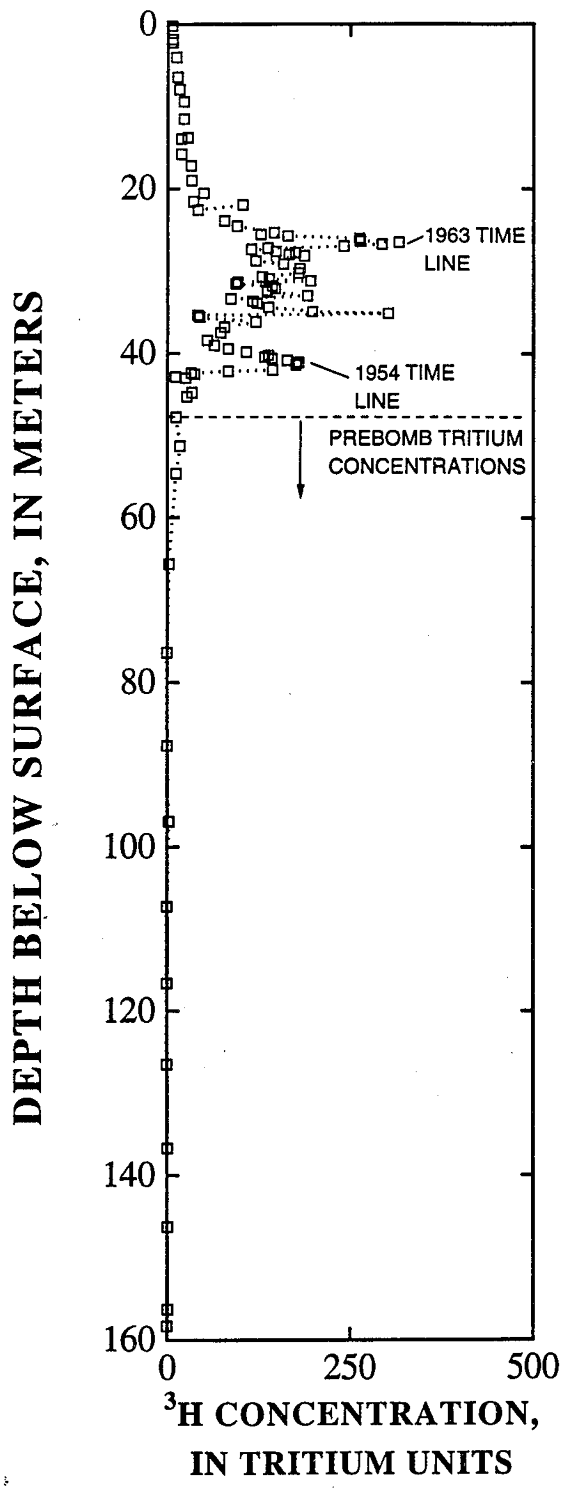

FIGURE 2. Tritium concentration profile and estimated time lines in the ice core collected from Upper Fremont Glacier, Wyoming.

with present-day ice accumulation rates measured in the Wind River Range (Naftz, 1993).

VARIATION OF $\delta^{18}O$ VALUES IN SNOW AND ICE-CORE SAMPLES

One hundred, equally spaced samples of surface snow (top 5 cm) from an 8100-m^2 area adjacent to the drill site (Fig. 4) were analyzed for $\delta^{18}O$ to quantify lateral variability at the drill

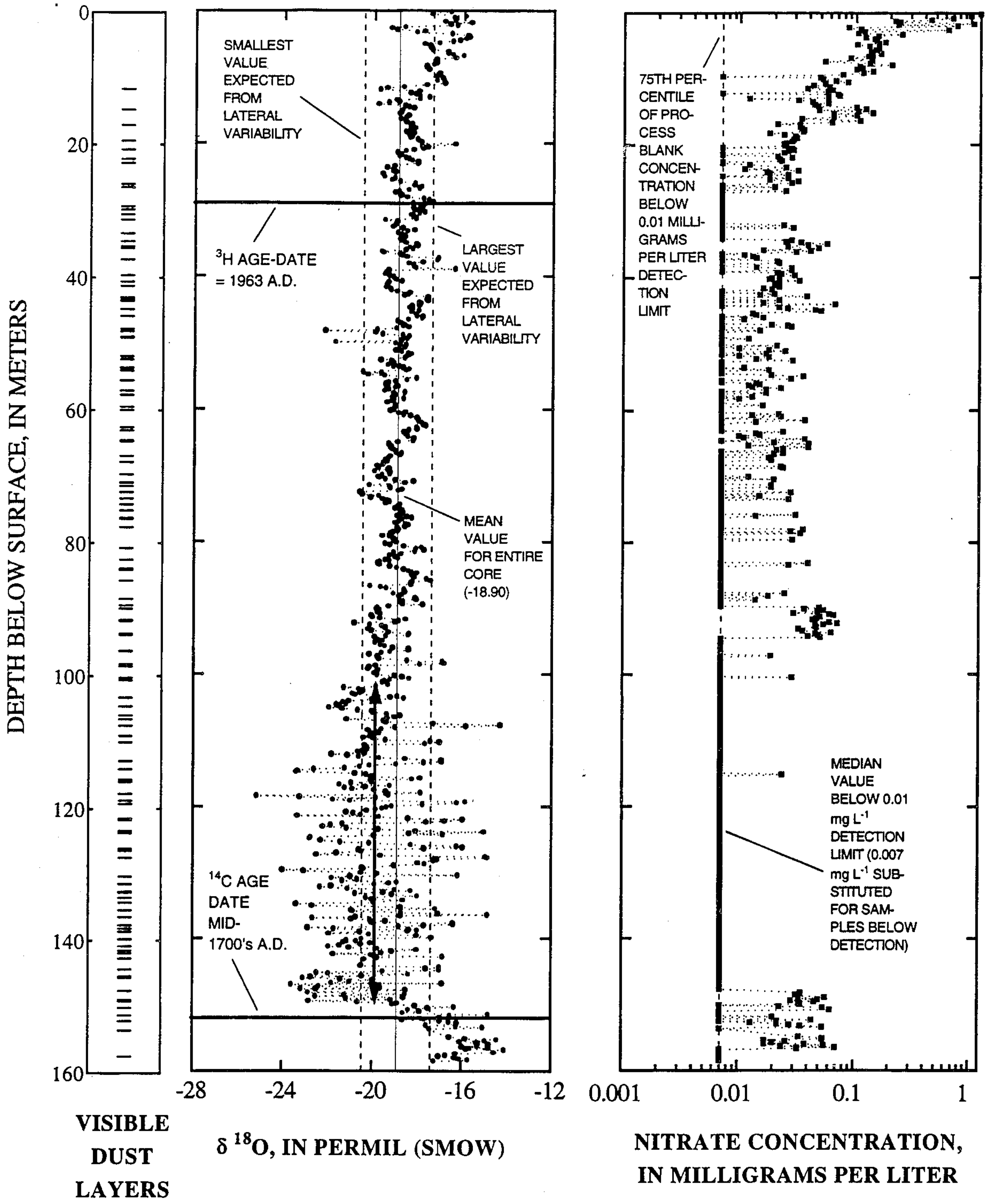

FIGURE 3. Visible dust layers, estimated age horizons, d^{18}O-value, and nitrate concentration profiles in the ice core collected from Upper Fremont Glacier, Wyoming. The large nitrate concentrations above 20 m probably reflect the limited amount of time the ice/firn has been subjected to meltwater elution processes. (Heavy black arrow, mean value for core interval 101.8 to 150 m [-19.85].)

site (Naftz et al., 1994). The snow-grid data were block kriged using 25 square blocks and a maximum search distance of 50 m. The estimate for each block was then contoured (Fig. 4A). The 1 σ uncertainty in the estimate was contoured as Figure 4B. The average 2 σ uncertainty from the block kriging results (mean $= -14.90 \pm 1.53$‰, $N = 25$) was then used to establish the upper and lower limits of significant variability with respect to the δ^{18}O profile in the ice-core record (Fig. 3).

The δ^{18}O profile (Fig. 3) was determined from 760 equally spaced samples along the length of the core. The δ^{18}O values from the upper 101.8 m of the core generally were within the $\pm$ 2 σ range of lateral variability calculated from the surficial snow grid data. From 101.8 to 150 m, the mean δ^{18}O value shifted abruptly to -19.85‰, 0.95‰ lighter than the mean core value of -18.90 (Fig. 3). Using age-to-depth relation developed from the ^{3}H and ^{14}C data, this isotopically lighter section of the core corresponds to the approximate time interval from the A.D. mid-1700s to mid-1800s.

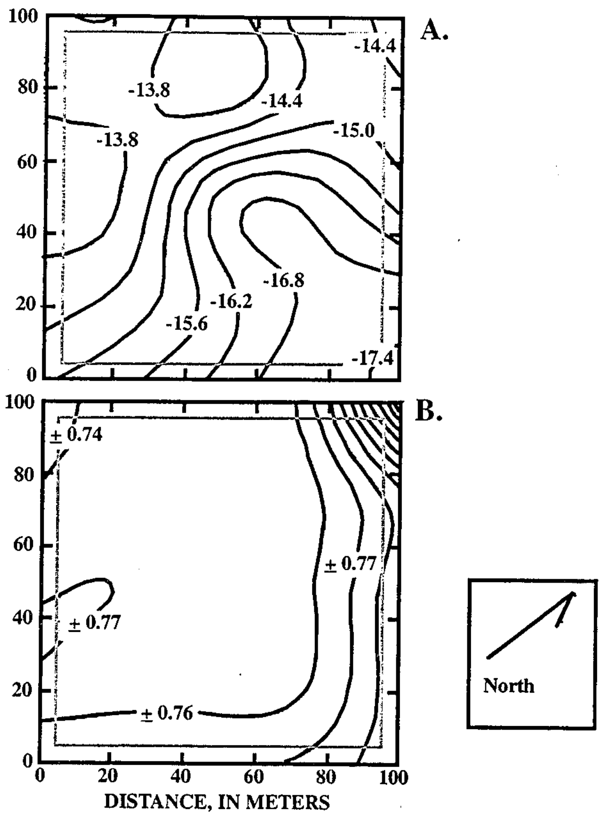

FIGURE 4. Contour maps of snow grid surface showing kriged estimate of d^{18}O values (SMOW) in ‰ (A) and the one sigma uncertainty associated with the estimate in permil (B). Contour interval is 0.6‰ for map A and ±0.01‰ for map B. The dashed area within each map represents the boundary of the original 8100-m^2 snow grid. The data were block kriged by using 25 square blocks of equal area. A maximum of 30 samples was used to estimate each block. One hundred samples were collected within the grid area in August 1991.

In Europe, the time period from the A.D. mid-1700s to mid-1800s was characterized by cool temperatures (Jones and Bradley, 1992) and is generally referred to as the end of the LIA (Thompson, 1992). Written historical records of climate in western North America during this same time period do not exist (Jones and Bradley, 1992); however, temperature trends have been reconstructed with tree ring information in nonalpine settings from the western United States (Fritts and Shao, 1992). These proxy records of temperature from nonalpine settings indicate cool periods from the 1750s to 1770s and from the 1830s to 1840s in the western United States. The pronounced shift to lighter isotopic values in the section of the core corresponding to the approximate time interval of the LIA indicates that a low resolution, stable isotope record of climate has been preserved. Furthermore, the isotopic record from the ice core indicates that more continuous cooling periods may have existed in the alpine regions of the western United States relative to the lower elevation settings where tree ring data were collected for paleotemperature reconstruction in the western United States.

Numerous high-amplitude oscillations exceeding the expected upper and lower limits of $\delta^{18}O$ lateral variability were detected in the 101.8- to 150-m core interval (Fig. 3). Without seasonal dust layers to guide sample selection, the 20-cm composite samples in this section of the core would not consistently be composed of either 100% winter or 100% summer precipitation. Most samples represent a mixture of both summer and winter precipitation with a resulting $\delta^{18}O$ value close to the core mean. The samples with $\delta^{18}O$ values exceeding the limit of lateral variability probably represent a dominant summer season (exceeds upper limit of lateral variability) or winter season (exceeds lower limit of lateral variability) sample composite. These large oscillations identified by the random sampling scheme may reflect increased seasonality or better preservation of the annual signal as a result of the cooler summer temperatures during the LIA. For example, the isotopically enriched horizons in this interval could have resulted from decreased rates of melting of isotopically enriched summer snowfall expected from cooler summer temperatures during the LIA. In contrast, during the 1990 and 1991 field seasons, summer snowfall at the site melted within 4 d of deposition. The abrupt decrease in the large amplitude oscillations at the 101.8-m depth (Fig. 3) indicates a sudden termination of the LIA at this site.

Site-specific tree-ring chronologies and ice-core nitrate concentrations also support the $\delta^{18}O$ climatic interpretations of the Upper Fremont Glacier ice core. Selected tree cores collected 2 km from the drill site consistently showed a sustained period of reduced radial growth beginning about A.D. 1790 and continuing until about A.D. 1840 (Naftz, 1993). This reduction in radial growth probably reflects a prolonged period of cooler summer temperatures that reduced the growing season at these high-altitude sites. With one exception, the nitrate concentration of samples from the 101.8- to 150-m depth interval did not exceed the analytical detection limit of 0.01 mg L^{-1} (Fig. 3). In contrast, a significant number of samples above and below this depth interval contained detectable nitrate concentrations. As in the Upper Fremont Glacier ice core, decreased nitrate concentrations have been detected during ice-age maxima in ice-core records from Greenland (Herron and Langway, 1985). The specific mechanism(s) causing the lower nitrate concentrations in the Upper Fremont Glacier core is unknown. For example, cooler temperatures during the LIA may have decreased the activity of nitrogen-fixing bacteria that may have been present on the snow surface.

The core section from 150 m to bedrock contains a significant isotopic shift to heavier values coupled with an abrupt increase in nitrate concentrations. An explanation for this shift is not obvious and may represent postdepositional melting/freezing processes at the ice-bedrock interface or a rapid shift to warmer climatic conditions in the A.D. mid-1700s. Reconstructed temperature trends from tree ring information in nonalpine settings from the western United States indicate a warming period from A.D. 1650 to 1740 (Fritz and Shao, 1992). Additional data are required from this section of the core before a plausible explanation of this isotopic and nitrate shift can be determined.

PALEOCLIMATE LINKAGE DURING THE TERMINATION OF THE LIA

A linkage between the isotopic responses to climate change appears to exist between ice cores from the Upper Fremont Glacier and Quelccaya Ice Cap (14°S, 71°W, Peruvian Andes, South America) (Fig. 5). Although the resolution and dating of the Upper Fremont Glacier ice-core record are less than those from the South American site, three distinct climate-related features seem to be preserved in both sets of records. First, both records show a shift to more negative $\delta^{18}O$ values within core segments that were deposited during the time period corresponding to historical records of the LIA. Relative to whole-core $\delta^{18}O$ averages, the Quelccaya Ice Cap Summit Core record shows about a

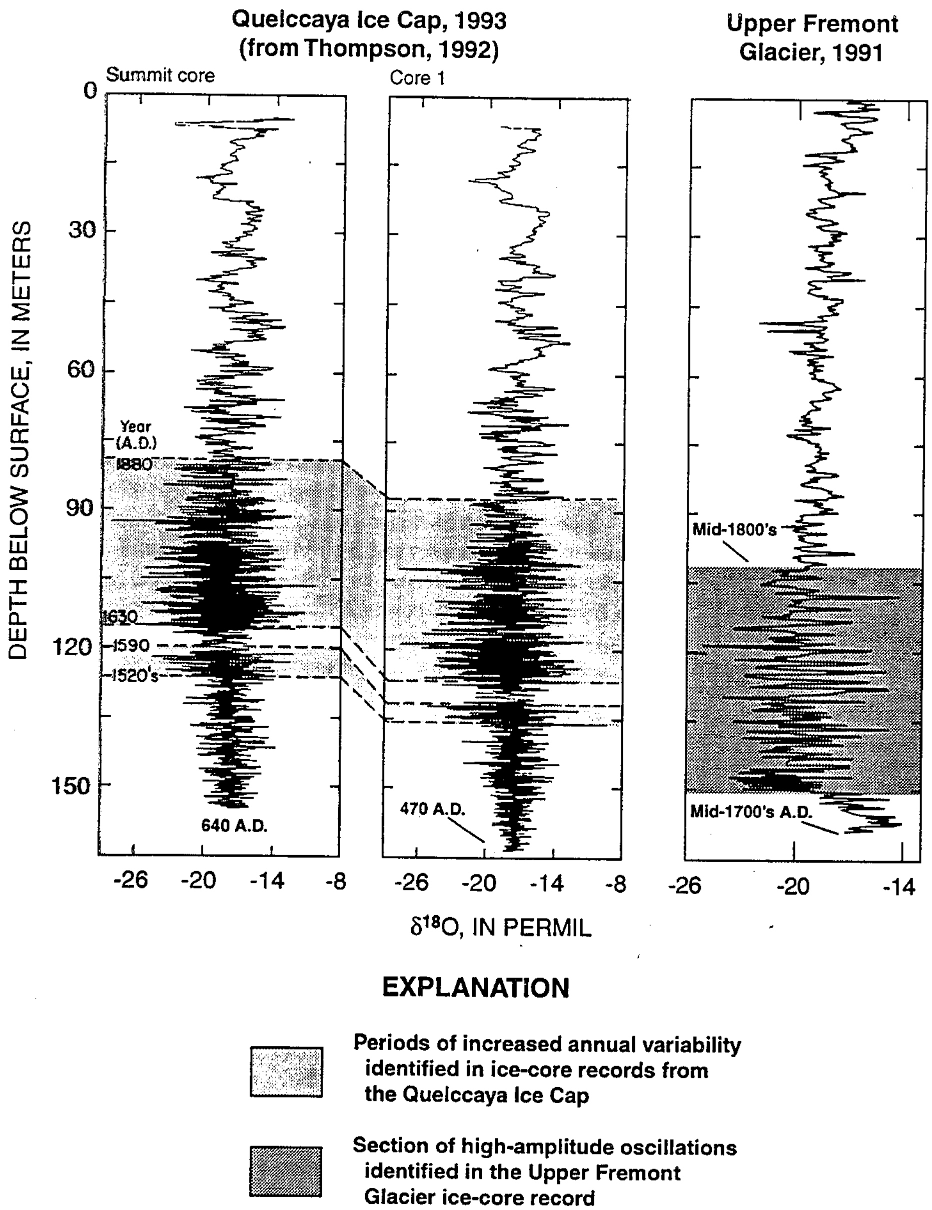

FIGURE 5. Relation of the $d^{18}O$-value profiles from the Quelccaya Ice Cap, Peru, and Upper Fremont Glacier, United States. Shaded areas identify areas of high-amplitude oscillations. Quelccaya Ice Cap plots reprinted from Thompson (1992) and published with permission from Chapman and Hall.

-0.7‰ shift during A.D. 1600 to 1800 (Thompson, 1992), compared to a -0.95‰ shift in Upper Fremont Glacier ice representing snow deposited during the end of the LIA (Fig. 3). Second, both records show an increase in the small-scale variation in the $\delta^{18}O$ signal during the LIA (Fig. 5). In the two Quelccaya ice-core records the average annual range of $\delta^{18}O$ during the LIA (A.D. 1520 to 1880) was twice the average annual range observed after the termination of the LIA (A.D. 1880 to 1980) (Thompson, 1992). Numerous high-amplitude oscillations in $\delta^{18}O$ during the LIA exceed both the upper and lower limits of $\delta^{18}O$ lateral variability in the Upper Fremont Glacier ice-core record (Fig. 3). Third, both cores show an abrupt shift from the high-amplitude isotopic variations during the LIA to much lower amplitude isotopic variations characteristic of post-LIA ice. This transition occurs abruptly (probably in about 2–3 yr) in all three cores. Thus, the linkages of the Upper Fremont Glacier isotopic record to the established paleoclimate record in the Quelccaya ice cores appear to support further the first documentation of the LIA in an ice-core record from a temperate glacier in south-central North America.

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