Variable responses of western U.S. glaciers during the last deglaciation

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ABSTRACT

Cosmogenic ^{10}Be exposure ages from moraines in the Wallowa Mountains, Oregon, identify two maximal late Pleistocene glaciations at 21.1 ± 0.4 ka and 17.0 ± 0.3 ka and a minor glacial event at 10.2 ± 0.6 ka. Our new high-resolution chronology, integrated with other well-dated glacial records from the western United States, demonstrates substantial differences in the synoptic responses of western U.S. glaciers to climate forcing associated with the global Last Glacial Maximum and subsequent millennial-scale events originating in the North Atlantic region. These variable synoptic glacier responses identify large changes in the relative contributions of regional to global controls on the climate of the western United States that accompanied the deglaciation.

Keywords: cosmogenic elements, exposure age, paleoclimatology, Pinedale glaciation, Wallowa Mountains.

INTRODUCTION

A number of proxy records in the western U.S. register responses to mechanisms of late Pleistocene climate change operating at orbital (10^4–10^5 yr) and millennial (10^3 yr) time scales. Orbital-scale mechanisms, such as variations in insolation and ice sheets, and their attendant effects on the regional climate of the western United States since the global Last Glacial Maximum (LGM, 21 ± 2 ka) are now well understood (Bartlein et al., 1998). In contrast, mechanisms responsible for millennial-scale changes in lake (Benson et al., 1998), vegetation (Whitlock and Grigg, 1999), and glacier (Gosse et al., 1995a, 1995b; Phillips et al., 1996) systems remain poorly known. Several studies suggest some relationship between these millennial-scale changes and abrupt climate changes in the North Atlantic region (e.g., Clark and Bartlein, 1995; Phillips et al., 1996), but without a broad spatial distribution of well-dated climate records, assessment of potential teleconnections (Mikolajewicz et al., 1997; Hostetler and Bartlein, 1999) is difficult.

The high sensitivity of alpine glaciers to climate change (Oerlemans et al., 1998) makes records of their fluctuations among the best potential proxies for identifying spatial and temporal terrestrial climate variability. Previously, development of such records has been hampered by limited opportunities for dating, but new surface-exposure dating methods using cosmogenic nuclides (Gosse and Phillips, 2001) are substantially improving our understanding of glacier fluctuations and their relationship to climate change. Here we report a high-resolution cosmogenic ^{10}Be chronology for late Pleistocene glacial events in the Wallowa Mountains, northeastern Oregon (Fig. 1). Our new ^{10}Be record fills an important gap between well-dated glacier records to the west (Sierra Nevada, Olympic Mountains, Puget Lowland) and east (Rocky Mountains) (Fig. 1), thereby providing adequate coverage for identifying synoptic responses to the large climate changes during the last deglaciation.

SAMPLE SITES IN THE WALLOWA MOUNTAINS

Several large valley glaciers drained a late Pleistocene ice cap that covered the central area of the Wallowa Mountains; the most prominent outlet glacier flowed down the Wallowa River valley on the north flank of the Wallowa Mountains. We sampled large (≥1 m) granodiorite boulders for ^{10}Be measurements from several closely spaced moraine crests that comprise the youngest two drift units (T and W) deposited by this outlet glacier. These moraines are characterized by grass-covered, relatively sharp crested ridges with minimal weathering and abundant surface boulders. Moraine segments within the two younger units were further subdivided (TTO and TTY, WTO and WTY) on the basis of field mapping, although these subunits are indistinguishable by relative dating criteria (Crandell, 1967). We adopted these informal map units and sampled boulders from segments of moraine crests assigned to units TTO, TTY, and WTO (Fig. 1). Units TTO and TTY occur as closely paired lateral moraines that flank Wallowa Lake, with a crest-to-crest

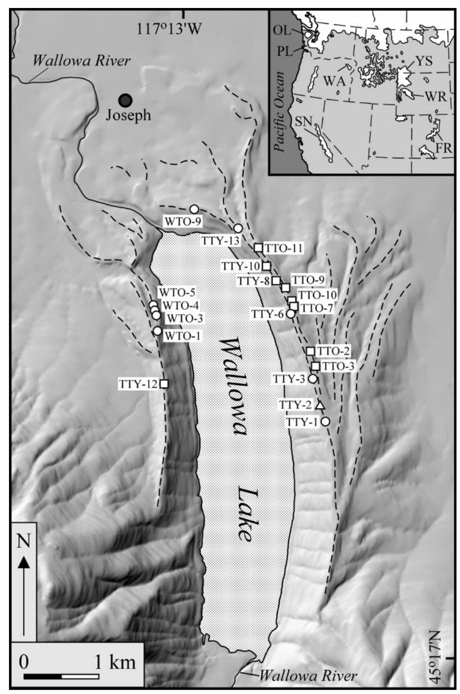

Figure 1. Glacial deposits at Wallowa Lake, as mapped by Crandell (1967). Dashed lines mark prominent moraine crests. Symbols identify locations of surface boulders sampled for ^{10}Be exposure dating: squares—boulders associated with older (TTO age) glacial advance; circles—boulders associated with younger (WTO age) advance; triangle—young outlier from TTY unit (see text). Glacial deposits in Glacier Lake cirque are located 25 km up valley (south) from north end of lake (shown in Fig. DR1, see footnote 1 in text). Inset map shows western United States during last glaciation with locations of glacial records discussed in text. WA—Wallowa Mountains; OL—Olympic Mountains; PL—Puget lobe; SN—Sierra Nevada; YS—Yellowstone; WR—Wind River Mountains; FR—Colorado Front Range.

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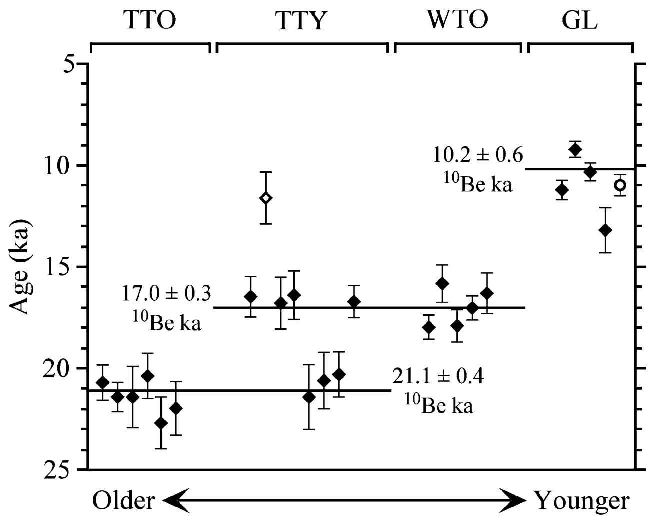

Figure 2. ^{10}Be ages of Wallowa Lake and Glacier Lake moraines, plotted according to drift unit. Older and younger age clusters within TTY unit are associated with events recorded by TTO and WTO moraines, respectively (see Fig. 1). Open diamond—young outlier from TTY unit; open circle—bedrock age from Glacier Lake moraine; neither age is included in weighted means of associated landform. Error bars on each age represent 1σ analytical uncertainty only, and do not include errors due to production rate, scaling, and other uncertainties (see footnote 1 in text). Horizontal lines and stated ages indicate weighted means of each landform.

distance of <50 m separating them over much of their length. Unit WTO corresponds to the well-defined innermost lateral and end moraines at the north end of Wallowa Lake.

We also dated a recessional moraine at Glacier Lake, 25 km up-valley from Wallowa Lake and <1 km from the cirque headwall of the former outlet glacier (Fig. DR1[1]). We sampled large granodiorite boulders on the moraine crest and a polished bedrock surface immediately distal to the moraine.

RESULTS

The cosmogenic ^{10}Be data from the TTO and WTO units yield weighted mean ages of 21.3 ± 0.4 ka (n = 6) and 17.2 ± 0.4 ka (n = 5), respectively (Table DR1, see footnote 1; Fig. 2). Excluding one boulder age as a young outlier (duplicate samples TTY-2A, 2B; weighted mean ^{10}Be age 11.6 ± 1.3 ka), ^{10}Be ages obtained from differing segments of the TTY unit exhibit a strong bimodal distribution with two coherent clusters centered on 20.6 ± 0.8 ka (n = 3) and 16.6 ± 0.5 ka (n = 4) (Table DR1, see footnote 1; Fig. 2). We associate the older and younger age clusters from the TTY morainal deposits with events recorded by the TTO and WTO moraines, respectively. In this regard, we note that boulders TTY-1, TTY-3, TTY-6, TTY-8, TTY-10, and TTY-13 all come from a continuous moraine crest segment on the east side of Wallowa Lake (Fig. 1). We interpret the bimodal clustering of their exposure ages to indicate that this moraine segment is a composite feature that formed during two successive glaciations of approximately equal extent. Following this interpretation, boulders TTY-8 and TTY-10 were deposited during the older (TTO age) glacial advance, whereas boulders TTY-1, TTY-3, TTY-6, and TTY-13 were deposited during the subsequent (WTO age) reoccupation of the moraine. The older age of boulder TTY-12, from the west side of Wallowa Lake, indicates that this moraine segment probably correlates to the TTO moraine on the east side of the lake (Fig. 1). Including all ^{10}Be boulder ages from each age population yields weighted means of 21.1 ± 0.4 ka (n = 9) for the older event and 17.0 ± 0.3 ka (n = 9) for the younger event (Fig. 2). We consider these the best estimates for the ages of two distinct late Pleistocene glaciations recorded by the moraines at Wallowa Lake.

Unlike lower altitude moraines at Wallowa Lake, snow cover and topographic shielding corrections must be applied to the Glacier Lake data (see footnote 1). Excluding one boulder age as a young outlier (sample GL-2; 2.1 ± 0.2 ka, ^{10}Be), the remaining boulder ages yield a weighted mean ^{10}Be age of 10.2 ± 0.6 ka (n = 4), and the bedrock surface (duplicate samples GL-7B, 7C) has a ^{10}Be age of 11.0 ± 0.5 ka (Table DR1, see footnote 1; Fig. 2).

REGIONAL PATTERNS DURING THE GLOBAL LGM

We next evaluate our results in the context of other glacial records and their significance to regional climate change in the western United States. Several records suggest that western U.S. glaciers began to advance several thousand years before the LGM (Sturchio et al., 1994; Hicock and Lian, 1995; Benson et al., 1998; Thackray, 2001). However, glaciers in the northern region subsequently retreated by 23–22 ka and remained retracted during the LGM. The large outlet glacier that drained the northern Yellowstone ice cap was restricted to <50% of its maximum extent between 22.5 and 19.5 ka, and subsequently readvanced to its maximum position (Sturchio et al., 1994; Licciardi et al., 2001). The Puget lobe of the Cordilleran ice sheet was retracted between 22.2 and 18.9 ka, and subsequently readvanced to its maximum extent (Hicock and Lian, 1995; Porter and Swanson, 1998). The glacier in the Olympic Mountains retreated from its LGM position by 22.8 ka and subsequently underwent only minor fluctuations (Thackray, 2001). A contrasting response is identified for glaciers in more southern positions, where cosmogenic ages from moraines in the Wallowa Mountains (21.1 ± 0.4 ka, ^{10}Be), the Wind River Mountains (20.1 ± 1.0 ka, ^{10}Be) (Gosse et al., 1995b), the Sierra Nevada (21 ± 3 ka, ^{36}Cl; Phillips et al., 1996; see footnote 1), and the San Bernardino Mountains (20–18 ka ^{10}Be) (Owen et al., 2003) indicate that these glaciers remained at their maximum extents during the LGM (Fig. 3D).

LGM climate of the western United States was influenced by a large Laurentide ice sheet, low atmospheric greenhouse gas concentrations, low summer insolation, and cold sea-surface temperatures (SSTs) in the northeastern Pacific and North Atlantic (Fig. 3). Simulations with atmospheric general circulation models indicate that the thermal and orographic effects of the Laurentide ice sheet influenced the climate and mass balance of glaciers in the region immediately south of the ice-sheet margin (Hostetler and Clark, 1997; Bartlein et al., 1998; Hostetler and Bartlein, 1999). We suggest that glaciers in the northern region underwent subdued advances or retreated during the LGM in response to their proximity to the Laurentide ice sheet, whereas glaciers farther south were beyond the immediate influence of the ice sheet and thus remained at their maximum extent.

Insofar as cosmogenic exposure ages on glacial moraines identify the final time of moraine occupation, these data also indicate that deglaciation in the southern region of the western United States occurred at 21–20 ka (Fig. 3D). The only substantial changes in potential forcings known to have occurred at this time are increases in North Atlantic thermohaline circulation and attendant air and sea-surface temperatures (Figs. 3A and 3B), suggesting that these contributed to deglaciation of the western United States.

WIDESPREAD DEGLACIATION COMMENCING ca. 17 ka

Our Wallowa chronology and all other dated records from the western United States

[1]GSA Data Repository item 2004013, methods, measurements, and uncertainties, Table DR1 (beryllium data for Wallowa moraines), and Figure DR1 (glacial deposits in Glacier Lake cirque area), is available online at www.geosociety.org/pubs/ft2004.htm, or an request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

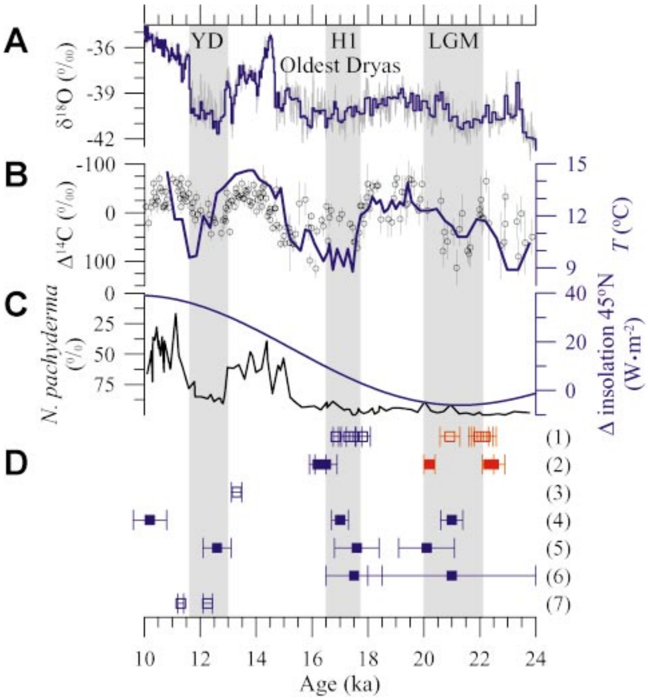

Figure 3. Comparison of potential climate forcings during last deglaciation with western U.S. glacial records. YD—Younger Dryas; H1—Heinrich event 1; LGM—Last Glacial Maximum. A: Greenland Ice Sheet Project 2 oxygen isotope proxy record of air temperature over Greenland. Gray line—high-resolution data from Stuiver and Grootes (2000); blue line—lower resolution data from Grootes et al. (1993). B: Black symbols—detrended record of $\Delta^{14}C_{atm}$ from Lake Suigetsu, Japan (Kitagawa and van der Plicht, 2000) as proxy for strength of Atlantic thermohaline circulation; blue line—record of North Atlantic sea-surface temperatures (SSTs) derived from alkenones (Bard et al., 2000). C: Black line—record of % *N. pachyderma* (s.) as proxy of SSTs of eastern Pacific (Mix et al., 1999); blue line—anomaly of July insolation at 45°N relative to today (Berger and Loutre, 1991). D: Age data from western U.S. glacial records. Filled symbols—cosmogenic ages; open symbols—calibrated radiocarbon ages; red symbols—ages constraining glacier retraction; blue symbols—ages constraining glacier advance. Locations and references: (1) Puget lobe (Hicock and Lian, 1995; Porter and Swanson, 1998); (2) northern Yellowstone outlet glacier (Sturchio et al., 1994; Licciardi et al., 2001); (3) Canadian Rockies (Reasoner et al., 1994); (4) Wallowa Mountains; (5) Wind River Mountains (Gosse et al., 1995a, 1995b); (6) Sierra Nevada (Phillips et al., 1996; see footnote 1 in text); (7) Colorado Front Range (Menounos and Reasoner, 1997).

demonstrate that glaciers readvanced following the LGM, possibly in response to the Oldest Dryas cold period that began ca. 19 ka (Fig. 3A), and reached their maximum late Pleistocene extent ca. 17 ka (Fig. 3D). The more continuous records from the Puget lobe (Porter and Swanson, 1998) and Sierra Nevada (Benson et al., 1998) show that rapid deglaciation began shortly thereafter, which is supported by cosmogenic ages elsewhere that indicate moraine abandonment between 17.5 and 16.5 ka (Fig. 3D). Initiation of widespread deglaciation across the western United States ca. 17 ka therefore appears to coincide with Heinrich event 1 (H1) (Bond et al., 1999; Bard et al., 2000). Owing to uncertainties in dating resolution of events and in the radiocarbon calibration during this time (Kitagawa and van der Plicht, 2000), we cannot determine the precise phasing of H1 and glacier retreat, but our results clearly indicate that deglaciation began during the Oldest Dryas, ~2 k.y. prior to the onset of the Bølling warming (Fig. 3A).

Our new ^{10}Be data strengthen the argument that some glacial events in the western United States correlate with Heinrich events (Clark and Bartlein, 1995; Phillips et al., 1996; Benson et al., 1998). Previously, these correlations were used to argue that the climate of the western United States responded principally to changes in the Laurentide ice sheet and North Atlantic region climate. Climate model simulations that evaluated this hypothesis (Hostetler and Bartlein, 1999) indicate that western U.S. glaciers may have remained in positive mass balance in response to orographic collapse of the Laurentide ice sheet and continued North Atlantic cold conditions, such as those that characterized H1 and the Oldest Dryas. Negative mass balance was only achieved in response to North Atlantic warming, such as that associated with the Bølling-Allerød warm period. The pattern of deglaciation across the western United States during the Oldest Dryas that we document here thus requires additional controls on the mass balance of western U.S. glaciers at this time. Moreover, contemporaneous retreat of alpine and ice-sheet margins elsewhere in North America, Europe, New Zealand, and South America (Denton et al., 1999) suggests that deglaciation of the western United States may have been part of a global phenomenon.

Denton et al. (1999) attributed widespread glacier recession coeval with H1 to an increase in atmospheric water vapor and attendant global warming. We note, however, that climate models (Mikolajewicz et al., 1997; Hostetler et al., 1999) and proxy data (Figs. 3A and 3B) (Sánchez Goñi et al., 2002; Nebout et al., 2002) show strong cooling over circum–North Atlantic regions during Heinrich events, suggesting that the colder SSTs and increased sea-ice cover of the North Atlantic during H1 (Mikolajewicz et al., 1997; Bard et al., 2000) offset global warming in these regions. Proxy data also indicate extreme dryness in Europe during Heinrich events (Sánchez Goñi et al., 2002), suggesting that deglaciation in this region (Denton et al., 1999) may have occurred by moisture starvation in association with the colder North Atlantic.

GLACIER RESPONSES DURING THE YOUNGER DRYAS

Evidence for glaciation in the western United States during the Younger Dryas cold interval is marked by an apparent regional contrast in glacier responses. Dated records indicate Younger Dryas readvances of glaciers in the Colorado Front Range (13.1–11.3 ka, calibrated) (Menounos and Reasoner, 1997), the Wind River Mountains (12.6 $\pm$ 0.5 ka, ^{10}Be) (Gosse et al., 1995a), and the Canadian Rocky Mountains (13.2–11.6 ka, calibrated) (Reasoner et al., 1994), indicating glaciation throughout the Rocky Mountains at that time. In the Pacific Northwest, records indicate Younger Dryas oscillations of valley glaciers in the southern Coast Mountains of British Columbia (12.8–11.9 ka, calibrated) (Friele and Clague, 2002) and in the North Cascades (ca. 12.6 ka, calibrated) (Kovanen and Easterbrook, 2001). The Puget lobe, the margin of which had retreated into British Columbia, may also have fluctuated during the Younger Dryas (Kovanen and Easterbrook, 2002) (Fig. 3D). Farther south, new evidence suggests that glaciers in the San Bernardino Mountains of southern California readvanced $\leq$1 km during the Younger Dryas (13–12 ka, ^{10}Be) (Owen et al., 2003). Our new data from the Wallowa Mountains suggest that the Glacier Lake moraine (10.2 $\pm$ 0.6 ka, ^{10}Be) formed after the Younger Dryas, although the precise timing is difficult to resolve, given the limited number of boulder ages from this moraine. Dated lacustrine sediments document the absence of Younger Dryas glaciation near Mount Rainier, Washington (Heine, 1998), and in the Sierra Nevada (Clark and Gillespie, 1997).

Relative to the LGM, the orographic and thermal effects of the Laurentide ice sheet on the climate of the western United States were substantially reduced by the time of the Younger Dryas (Bartlein et al., 1998), whereas insolation (Fig. 3C) and atmospheric CO_2 were nearly at interglacial levels. We thus suggest that Younger Dryas glaciation in the western United States occurred in response to hemispheric cooling caused by a reduction in the Atlantic thermohaline circulation (Mikolajewicz et al., 1997). Because glaciers respond to precipitation as well as temperature, the absence of Younger Dryas glaciation in the Sierra Nevada and near Mount Rainier may reflect negative regional precipitation anomalies along with strong sensitivity of these glacier systems to changes in precipitation.

CONCLUSIONS

Our new cosmogenic ^{10}Be chronology of the Wallowa ice cap, integrated with the existing distribution of well-dated western U.S. glacial records, provides evidence for variable spatial responses of alpine glaciers during the last deglaciation. Nonuniform regional responses during the LGM may reflect the influence of a fully developed Laurentide ice sheet on the climate of the western United States. Widespread glacier retreat that began ca. 17 ka appears to be part of a global pattern of deglaciation associated with Heinrich event 1. The regional contrast in glacier responses across western North America during the Younger Dryas may reflect localized precipitation anomalies in association with a broader pattern of hemispheric cooling originating in the North Atlantic region. These variable synoptic glacier responses identify large changes in the relative contributions of regional to global controls on the climate of the western United States that accompanied the deglaciation.

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