Latest Pleistocene advance of alpine glaciers in the southwestern Uinta Mountains, Utah, USA: Evidence for the influence of local moisture sources

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ABSTRACT

Cosmogenic surface-exposure [10]Be dating of Last Glacial Maximum (LGM) moraines indicates that glaciers in the southwestern Uinta Mountains remained at their maximum positions until ca. 16.8 ± 0.7 ka, ~2 k.y. after glaciers in the neighboring Wind River Range and Colorado Rockies began to retreat. The timing of the local LGM in the southwestern Uintas overlaps with both the hydrologic maximum of Lake Bonneville and preliminary estimates of the local LGM in the western Wasatch Mountains. This broad synchroneity indicates that Lake Bonneville and glaciers in northern Utah were responding to similar climate forcing. Furthermore, equilibrium line altitudes (ELAs) for reconstructed LGM alpine glaciers increase with distance from the Lake Bonneville shoreline, rising from ~2600 m to ~3200 m over the 120 km length of the glaciated Uintas. This pronounced ELA gradient suggests that the magnitude of the latest Pleistocene glacial advance in the western Uintas was due, at least in part, to enhanced precipitation derived from Lake Bonneville; thus, the lake acted as a local amplifier of regional climate forcing. This relationship underscores the sensitivity of alpine glaciers to moisture availability during the latest Pleistocene, and further demonstrates the importance of local moisture sources on glacier mass balance.

Keywords: Last Glacial Maximum, alpine glaciers, climate, Rocky Mountains, late Pleistocene.

INTRODUCTION

Glacier mass balance is the primary control on the terminus position of nonsurging glaciers and is ultimately controlled by ablation season temperature and winter snow accumulation (e.g., Paterson, 1994). The advance and retreat of valley glaciers in response to changes in mass balance produce a suite of geomorphic features that can be interpreted to yield a record of past glacier fluctuations. Thus, the geomorphic record can provide important information about climate conditions during past glaciations.

A nearly uniform 1 km lowering of snowlines along the north-south extent of the American Cordillera was attributed by Broecker and Denton (1989) to uniform temperature depression, suggesting that glacier advance during the global Last Glacial Maximum (LGM, ca. 21–18 ka) was primarily a function of reduced ablation. Other research-ers have also noted the greater efficacy of ablation in controlling the position of glacier termini over shorter time scales (e.g., Lowell et al., 1999), and modeling studies have concluded that LGM alpine glaciers in the western U.S. were generally more sensitive to changes in temperature than in precipitation (Hostetler and Clark, 1997).

The importance of precipitation changes, however, cannot be overlooked, and recent research has illuminated the importance of precipitation as a control on valley glacier behavior in the western U.S. In particular, glacier advances late in the Pleistocene, several millennia after the global LGM and onset of the retreat of the Laurentide Ice Sheet, have been reported by studies in the central and northern Rocky Mountains (Benson et al., 2005; Licciardi et al., 2001; Thackray et al., 2004; see locations in Fig. 1). These advances were locally more extensive than those during the global LGM, requiring a climatic mechanism that could drive glacier advance during an interval of global deglaciation. In some cases, precipitation increases associated with reor-ganization of atmospheric circulation during the latest Pleistocene appear to have provided that mechanism. Specifically, moist westerly airflow increased in at least the northern Rocky Mountains after the global LGM as the glacial anticyclone weakened in response to orographic collapse of the Laurentide Ice Sheet (Licciardi et al., 2004; Meyer et al., 2004; Thackray et al., 2004). Thus, while temperature depression was responsible for the advance and retreat of glaciers over orbital time scales, changes in precipitation reflecting fluctuations in large-scale atmospheric circulation may have been regionally dominant over millennial scales.

The identification of latest Pleistocene glacial advances in response to increased moisture availability raises several questions that are important to furthering the understanding of the glacial history of the Rocky Mountains and the paleoclimate of this region: (1) What was the spatial distribution of latest Pleistocene glaciers? (2) Were advances synchronous across the Cordillera? (3) What was the role of meso-scale or local moisture sources? To address these questions, we reconstructed the extent of LGM mountain glaciers in the Uinta Mountains of northeastern Utah and determined the timing of their maximum advance and onset of their retreat. Our selection of this study area was based on several criteria. First, the Uintas contained the largest area of ice between the Colorado Rockies and the Sierra Nevada during the LGM (Atwood, 1909; Laabs and Carson, 2005; Munroe, 2005). Second, the location of the range provides an important point of reference from which to consider latest Pleistocene glacier fluctuations in other parts of the Rocky Mountains. Third, the unique east-west orientation of the Uintas allows the opportunity to evaluate the relationship between LGM glacier extent and west-east climate gradients (Munroe and Mickelson, 2002). Fourth, the glacial Uintas were located downwind of Lake Bonneville,
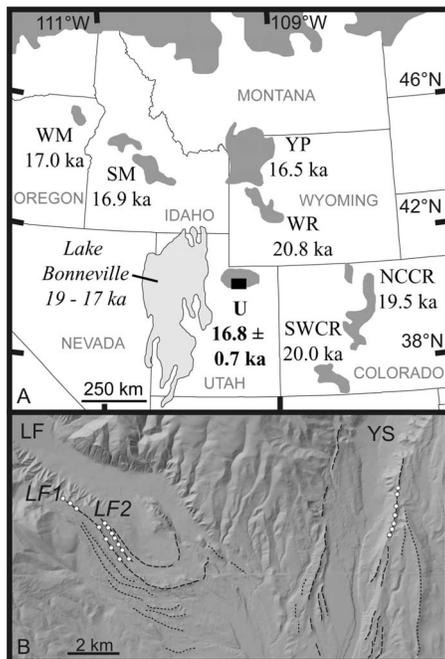
Figure 1. A: Map of glacial localities in central and northern Rocky Mountains south of the Cordilleran and Laurentide Ice Sheets (margins shown schematically along northern part of map) recently dated by cosmogenic surface-exposure or radiocarbon methods. Dark gray polygons are locations of Last Glacial Maximum glacier systems in Wallowa Mountains (WM; Licciardi et al., 2004), Sawtooth Mountains (SM; Thackray et al., 2004), Yellowstone Plateau (YP; Licciardi et al., 2001), Wind River Mountains (WR; Benson et al., 2004), north-central Colorado Rockies (NCCR; Benson et al., 2005), southwestern Colorado Rockies (SWCR; Benson et al., 2005), and Uinta Mountains (U; this study). Mean ^{10}Be boulder-exposure ages from Benson et al. (2005) are recalculated using production rates employed in this study and in Licciardi et al. (2001, 2004). Box indicates location of Lake Fork and Yellowstone River Canyons. B: Shaded-relief map of Lake Fork (LF) Canyon and Yellowstone (YS) Canyon terminal moraines. Long-dashed lines indicate Pinedale-equivalent moraines and short-dashed lines indicate pre-Pinedale moraines. Circles mark locations of sampled boulders.

the largest pluvial lake in the Great Basin (Gilbert, 1890; Oviatt, 1997), allowing the possible significance of a local moisture source to be evaluated.

Here we report the first cosmogenic surface-exposure ages that delimit the timing of the local LGM in northeastern Utah. We also present reconstructed equilibrium line altitudes (ELAs) for Uinta Mountain glaciers during the LGM, and then discuss the climatic and temporal relationship between the mass balance of these glaciers and the hydrologic balance of Lake Bonneville. Our results add to the current understanding of the latest Pleistocene behavior of alpine glaciers in the west-

ern U.S. and highlight the importance of local moisture sources on glacier mass balance near the end of the last glaciation.

METHODS
Cosmogenic Surface-Exposure Dating

To set limits on the timing of the LGM in the southwestern Uinta Mountains, we targeted the most prominent lateral and terminal moraines from the last glaciation, located at the mouths of the Lake Fork and Yellowstone Canyons (Fig. 1). We sampled 7 boulders on the outermost lateral moraine from the last glaciation in Yellowstone Canyon and 14 boulders on a terminal moraine in Lake Fork Canyon. The latter is a compound feature; seven samples are from the distal ridge (labeled LF1 in Fig. 1) and seven samples are from the proximal ridge (labeled LF2 in Fig. 1). The Lake Fork terminal moraine complex was targeted to determine the time interval between the formation of the two ridges. Suitable boulders of metaquartzite or weakly metamorphosed sandstone were sampled with a sledgehammer and chisel. Laboratory procedures used to isolate beryllium in boulder samples were adopted from Bierman et al. (2003) and were completed at the University of Wisconsin Cosmogenic Nuclide Preparation Lab (laboratory methods are summarized in the GSA Data Repository[1]). Concentrations of ^{10}Be were measured at the Purdue Rare Isotope Measurement Lab by accelerator mass spectrometry (AMS).

Boulder-exposure ages were calculated from measured ^{10}Be concentrations using the age equation from Lal (1991) and a production rate of 5.1 ± 0.3 (2σ) atoms g SiO_2^{-1} yr^{-1} scaled for site elevation and geographic latitude following Stone (2000). We consider the effects of topographic shielding, boulder erosion, and snow cover to be minimal at all sample locations (age calculations and moraine boulder properties are described in Tables DR1 and DR2). Although previous studies suggest that uncertainties of scaling production rates are as much as 20% (e.g., Bierman et al., 1999), we report only the analytical uncertainty of the AMS measurements in our age estimates. The mean of acceptable boulder-exposure ages ($\pm 2\sigma$) from each moraine, weighted according to the inverse-variance of measurement precision, was computed for individual moraines in the Yellowstone and Lake Fork canyons. This method of age interpretation is based on the

notion that scatter among boulder-exposure ages is caused by geologic phenomena, namely partial shielding of boulder surfaces during a period of moraine stabilization, along with analytical errors. The uncertainties associated with the weighted-mean boulder-exposure age estimates incorporate analytical uncertainties of each individual age.

LGM Glacier ELA Reconstructions

ELAs were estimated for all reconstructed LGM valley glaciers in the Uintas during the LGM (n = 44) using a weighted average of individual estimates obtained from three separate ELA estimation techniques: accumulation area ratio (AAR = 0.65), toe-headwall altitude ratio (THAR = 0.40), and the uppermost elevation of continuous lateral moraines (LM) after Meierding (1982). Weighting factors were AAR = 3, THAR = 2, and LM = 1, following Locke (1990) and Munroe and Mickelson (2002). The LGM extent of valley glaciers was reconstructed for the northern Uintas by Munroe (2001, 2005) and for the southern Uintas by Shakun (2003). Paleoglacier extents in the western Uintas were taken from Oviatt (1994).

RESULTS
Cosmogenic Surface-Exposure Dating

Seven boulders on the Yellowstone Canyon moraine (Table DR1; see footnote 1) and 16 boulders on the Lake Fork moraines (Table DR2) yield in situ cosmogenic ^{10}Be exposure ages that range from 11.5 ± 1.2 to 19.9 ± 2.0 ka (Fig. 2). The exposure ages from the Yellowstone Canyon moraine span this relatively broad range, but all of the ages overlap at 2σ (Fig. 2; Table DR1). A long interval of moraine-crest stabilization and/or subsequent boulder exhumation are possible explanations for the younger ages on this moraine, although our observations of moraine morphology failed to identify evidence of variable moraine-crest erosion rates among the sample sites. Nonetheless, the scatter of boulder-exposure ages on the Yellowstone moraine precludes a precise determination of the time of moraine abandonment. Thus, we consider the weighted-mean age of the seven boulders, 15.2 ± 2.6 ka, to be a working estimate for the age of the Yellowstone moraine until more data are available, and base our following interpretations of the deglacial chronology for the southwestern Uintas on the cosmogenic surface-exposure ages yielded by boulders on the Lake Fork moraines (Tables DR1 and DR2).

In the Lake Fork Canyon, a single sample that returned an exposure age of 11.5 ± 1.2 ka on ridge LF1 is a statistical outlier at 2σ compared to six other samples (Fig. 2; Table

Figure 2. Cosmogenic [10]Be exposure ages of moraine boulders from Yellowstone Canyon (diamonds), Lake Fork 1 (squares), and Lake Fork 2 (triangles) moraines (see location maps in Fig. 1). Error bars are 2σ analytical errors and open symbols are considered outliers. Gray areas show weighted-mean age of each moraine with its 2σ uncertainty.
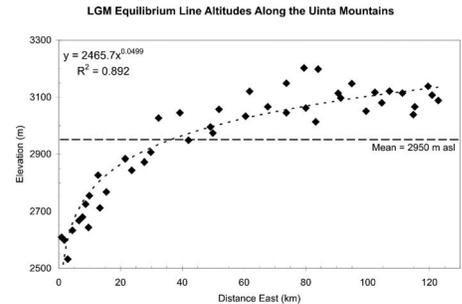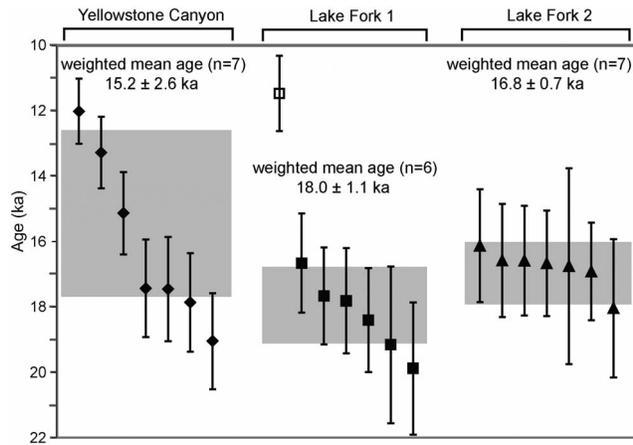

Figure 3. Reconstructed equilibrium line altitudes for Last Glacial Maximum (LGM) glaciers in Uinta Mountains from Munroe (2001), Shakun (2003), and Oviatt (1994) (asl—above sea level).

DR2 [see footnote 1]), and is therefore omitted from further calculations. The six oldest boulder-exposure ages from LF1 are indistinguishable from one another and have a weighted mean of 18.0 ± 1.1 ka. The morphology of the Lake Fork terminal moraine system requires that the glacier retreated after ca. 18 ka to form ridge LF2. The seven boulder-exposure ages from this inner ridge are also indistinguishable from one another with a weighted mean of 16.8 ± 0.7 ka. Because LF2 is the innermost ridge in the Lake Fork terminal moraine system (Fig. 1) and yields a robust set of moraine-boulder exposure ages (Fig. 2; Table DR2), we consider this age to best represent the onset of deglaciation in the southwestern Uinta Mountains.

ELA Reconstructions

The weighted means of the reconstructed ELAs rise ~600 m in elevation from west to east along the 120 km length of the Uinta Mountains; most of that rise occurs in the western 80 km of the range (Fig. 3). Assuming a uniform temperature depression at 40°N latitude in northern Utah, this ELA pattern indicates the presence of a profound east-west precipitation gradient during the LGM. Furthermore, the lowest ELAs in the western Uintas and Wasatch Mountains are as much as ~400 m lower than those predicted by the reconstructed regional LGM ELA pattern in Porter et al. (1983), suggesting that a local source may have enhanced moisture delivery to these glaciers.

DISCUSSION AND CONCLUSIONS

Our interpretation that deglaciation began in the southwestern Uintas ca. 16.8 ± 0.7 ka suggests that glaciers there reached, or persisted at, their maximum extents considerably later than glaciers in neighboring ranges of the Rocky Mountains. For example, Benson et al. (2005) documented mean ages for the onset of deglaciation of 18.4 ka in north-central Colorado, 18.9 ka in southwestern Colorado, and

18.7–20.8 ka for the Wind River range (assuming no snow cover, and with [10]Be ages for the Wind River range adjusted to the production rate used here). However, the exposure ages determined for the Uinta moraines are consistent with those from the western Wasatch Mountains (Lips et al., 2005) that indicate a glacial maximum as late as 17–15 ka. The Wasatch dates are preliminary, but their overlap with the Uinta results reported here suggests that the LGM and subsequent retreat in these two ranges roughly coincided with the hydrologic peak of Lake Bonneville and its subsequent climate-driven fall below the Provo shoreline after ca. 14.0 ^{14}C ka (16.2–17.1 ka) (Oviatt, 1997). Previous studies suggest that northward migration of the polar jet stream during the latest Pleistocene in response to the orographic collapse of the Laurentide Ice Sheet altered the climate of the northern Great Basin in a way that allowed transgression of Lake Bonneville several millennia after the global LGM (e.g., Benson and Thompson, 1987; Oviatt, 1997). This phenomenon may also have supported the growth of alpine glaciers in the Wasatch and southwestern Uinta Mountains during the latest Pleistocene through increased precipitation and/or reduced summer temperatures. Thus, the broad synchroneity of the local LGM in northeastern Utah and the lake highstand likely reflects a regional climate forcing.

Evidence for climate forcing at a subregional scale is provided by the westward sloping ELA gradient, which indicates that the magnitude of the glacial advance in the western Uintas was influenced by a local moisture source. Given the near constancy of modern summer adiabatic lapse rates (6.1°C/km) across the Uintas (Laabs et al., 2006), it is unlikely that there was appreciable spatial variability in altitude-temperature relationships in the range during the LGM. Thus, mean ablation season temperatures at the western ELAs, which are ~600 m lower that those in the east, would have been ~3.7 °C warmer than those

at the highest ELAs, and western Uinta glaciers must have received more precipitation than those farther east in order to reach their maximum positions. Given the sensitivity of ELAs to precipitation and temperature (Seltzer, 1994), and assuming the modern summer lapse rate, the ELA depression requires that the western Uinta glaciers received at least 1000 mm more winter accumulation than glaciers at the eastern end of the range, where ELAs were higher and ablation season temperatures were presumably cooler (Munroe and Mickelson, 2002). This difference exceeds the modern orographic effect in the range, which is on the order of a few hundred millimeters of water-equivalent precipitation. Thus, the lower ELAs in the western Uintas reflect substantial enhancement of the modern orographic precipitation pattern.

When considered together, the ELA pattern and the synchroneity between the hydrologic highstand of Lake Bonneville and the peak of glaciation in the southwestern Uintas suggest that Lake Bonneville was responsible for the additional precipitation (Munroe and Mickelson, 2002). Modeling studies have concluded that much of the lake remained ice free in January (Hostetler et al., 1994), and this large area of relatively warm water only ~100 km upwind from the western Uintas would have enhanced winter snowfall considerably beyond modern levels. This assertion is supported by glacier modeling by Laabs et al. (2006), which concluded that the Little Cottonwood glacier in the western Wasatch Mountains immediately downwind of Lake Bonneville was sustained at its LGM extent under a temperature depression of ~7 °C and a precipitation increase of ~3.5 times modern. Similar results were obtained by earlier efforts directed at modeling Lake Bonneville and the Little Cottonwood glacier (McCoy and Williams, 1985). Additional modeling is needed to determine how far downwind the effects of Bonneville moisture would have propagated, but the ELA pattern is strong evidence of en-

hanced precipitation at the upwind end of the Uintas. We propose that the regional climate forcing responsible for synchroneity of the Lake Bonneville highstand and the local LGM was amplified by the lake, affecting the magnitude of the local LGM at the western, upwind, end of the Uintas. Moisture from Lake Bonneville may therefore have been responsible for maintaining southwestern Uinta glaciers at their maximum positions long after the onset of glacier retreat in other parts of the central Rocky Mountains.

An important test of this hypothesis will be to determine the timing of the local LGM at the eastern end of the Uintas, where the ELA pattern suggests the influence of Lake Bonneville was absent. Further work could be directed at identifying an analog for ridge LF2 in the Yellowstone Canyon, although our field reconnaissance indicates that boulders suitable for cosmogenic surface-exposure dating are rare on inner moraines in this valley. More precise estimates of [10]Be production rates at middle latitudes and high elevations will also improve our ability to accurately correlate moraine-boulder exposure ages with calibrated radiocarbon ages of Lake Bonneville sediments. This may become increasingly important if ongoing geomorphic investigations and the acquisition of additional radiocarbon ages from the Provo shoreline (e.g., Godsey et al., 2005; Oviatt and Thompson, 2005) support refinements to the chronology of Lake Bonneville's regressive phase.

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