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Chronology of the Last Glacial Maximum in the Upper Bear River Basin, Utah

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Abstract

The headwaters of the Bear River drainage were occupied during the Last Glacial Maximum (LGM) by outlet glaciers of the Western Uinta Ice Field, an extensive ice mass ($\sim 685 \text{ km}^2$) that covered the western slope of the Uinta Mountains. A well-preserved sequence of latero-frontal moraines in the drainage indicates that outlet glaciers advanced beyond the mountain front and coalesced on the piedmont. Glacial deposits in the Bear River drainage provide a unique setting where both ^{10}Be cosmogenic surface-exposure dating of moraine boulders and ^{14}C dating of sediment in Bear Lake downstream of the glaciated area set age limits on the timing of glaciation. Limiting ^{14}C ages of glacial flour in Bear Lake (corrected to calendar years using CALIB 5.0) indicate that ice advance began at 32 ka and culminated at about 24 ka. Based on a Bayesian statistical analysis of cosmogenic surface-exposure ages from two areas on the terminal moraine complex, the Bear River glacier began its final retreat at about 18.7 to 18.1 ka, approximately coincident with the start of deglaciation elsewhere in the central Rocky Mountains and many other alpine glacial localities worldwide. Unlike valleys of the southwestern Uinta Mountains, deglaciation of the Bear River drainage began prior to the hydrologic fall of Lake Bonneville from the Provo shoreline at about 16 ka.

Introduction

The western Uinta Mountains of Utah (Fig. 1) contain a rich record of alpine glaciation during the Quaternary Period (e.g., Atwood, 1909; Bryant, 1992; Munroe, 2001; Laabs, 2004). Mapping of well-preserved latero-frontal moraines, erosional trimlines, till sheets, and glacial-erosional forms has enabled precise reconstructions of ice extent in this area (e.g., Munroe, 2005; Laabs and Carson, 2005; Refsnider, 2006) during the late Pleistocene Smiths Fork Glaciation (Bradley, 1936). This event represents the Last Glacial Maximum (LGM) in the Uinta Mountains (Munroe et al., 2006) and is approximately correlative to the Pinedale Glaciation elsewhere in the Rocky Mountains (e.g., Gosse et al., 1995; Benson et al., 2005).

Glacier reconstructions can provide a framework for inferring regional paleoclimate over millennial timescales, but only when placed into a precise chronological context. Moraines in the upper Bear River valley are well-preserved and contain boulders suitable for cosmogenic surface-exposure dating. Bear Lake, located approximately 150 km downstream of the Smiths Fork–age

terminal moraine (Fig. 1), contains a sedimentary record that spans the last glacial cycle (e.g., Colman et al., 2005). In this paper, we report the first cosmogenic ^{10}Be surface-exposure ages (hereafter referred to as “cosmogenic-exposure ages”) for terminal moraines in the Bear River valley and describe a unique situation in which the timing of glacial activity is limited by both ^{14}C dating of lake sediments and *in situ* cosmogenic ^{10}Be surface-exposure dating of moraine boulders.

GLACIAL CHRONOLOGIES OF THE ROCKY MOUNTAINS AND INTERPRETATIVE STRATEGIES FOR COSMOGENIC-EXPOSURE DATING

Recent research in the Uinta Range and elsewhere in the central and northern Rocky Mountains has provided increasingly precise limits on the timing of the Pinedale Glaciation, due in part to developments in cosmogenic surface-exposure dating methods (e.g., Gosse and Phillips, 2001). On the basis of cosmogenic-exposure ages of moraine boulders in the Wind River Mountains, Gosse et al. (1995) suggested that the LGM in the central Rocky

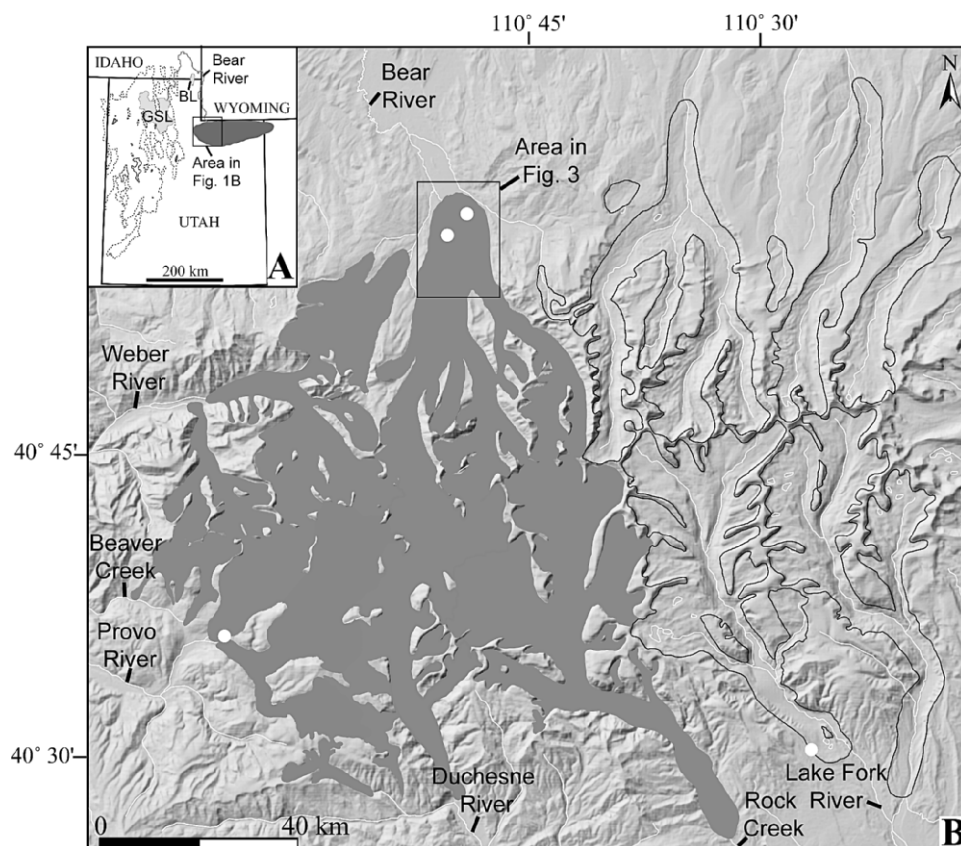


FIGURE 1. (A) Location of the Uinta Mountains and Bear River valley in Utah, Idaho, and Wyoming, U.S.A. BL = Bear Lake, GSL = Great Salt Lake. Dark gray area indicates the extent of the Uinta Mountains. Dotted line indicates the extent of pluvial Lake Bonneville. (B) Shaded relief map of the western Uinta Mountains. Gray area indicates the maximum extent of the Western Uinta Ice Field during the Smiths Fork Glaciation. Dark lines indicate the maximum extent of glaciers elsewhere in the Uinta Range during the Smiths Fork Glaciation. White dots show locations on moraines where boulders have been sampled for cosmogenic ¹⁰Be exposure dating in the Bear River (this study), Provo valley (Refsnider, 2006), and Lake Fork valleys (Munroe et al., 2006). White lines show the extent of major streams and their tributaries.

Mountains lasted from 23 to 16 ka. More recently, Benson et al. (2005) refined these age estimates and showed that widespread glacial retreat in the Rocky Mountains began somewhat earlier, at about 18 to 17 ka. These latter ages are consistent with cosmogenic-exposure ages from elsewhere in the central and southern Rocky Mountains (e.g., Brugger, 2006). They are also consistent with limiting ages on glaciation in some areas in the northern Rocky Mountains, such as the Yellowstone Plateau (Licciardi et al., 2004) and Sawtooth Mountains (Thackray et al., 2004), where the onset of ice retreat may have occurred shortly after 17 ka. In the Uinta Mountains, Munroe et al. (2006) estimated that deglaciation of the southwestern part of the range (Fig. 1) began at about 16.8 ± 0.7 ka, based on cosmogenic-exposure dating of moraine boulders.

Although the currently available chronology of the last glaciation in the Rocky Mountains suggests widespread contemporaneous retreat, careful consideration of cosmogenic-exposure ages is important because of differences in how these data sets can be interpreted. For example, Benson et al. (2005) suggested the youngest cosmogenic-exposure age from a moraine represents the time of moraine abandonment (i.e., deglaciation), as long as none of the moraine boulders have been exhumed. However, moraines may undergo a period of surface instability following the onset of ice retreat, during which erosion of the moraine crest exhumed boulders that would yield exposure ages that postdate moraine formation (Putkonen and Swanson, 2003; Briner et al., 2005). Given this possibility, Putkonen and Swanson (2003) suggested the *oldest* cosmogenic-exposure age in a sample set provides the best estimate of the moraine age. Other workers have reported the error-weighted mean of a sample of moraine-boulder cosmogenic-exposure ages as an estimate of moraine age (e.g., Licciardi et al., 2004; Douglass et al., 2005), which assumes that scatter within a set

of ages reflects random analytical errors (brought about by sample preparation and analysis) and random or systematic geologic errors (caused by, for example, moraine-crest erosion or inherited cosmogenic nuclides in moraine boulders). As a result of these different approaches, and because the scatter among cosmogenic-exposure ages from a single moraine is typically about 30% (Putkonen and Swanson, 2003), reported estimates of moraine ages depend significantly on how the set of ages is interpreted. In this paper, we explore the impact of these different interpretive frameworks by comparing radiocarbon and cosmogenic-exposure age limits on the last glaciation in the northwestern Uinta Mountains.

REGIONAL CLIMATE RECORDS FOR THE LAST GLACIATION

Glacial reconstructions in the Uinta Mountains based on field mapping of surficial deposits and landforms by Atwood (1909), Munroe (2005), Laabs and Carson (2005), and Refsnider (2006) provide the framework for inferring climatic conditions (namely summer temperature and winter precipitation) for the local LGM. Maps of Smiths Fork ice extents indicate that the central and eastern valleys of the Uinta Mountains were occupied by discrete valley glaciers ranging from about 4 to 43 km in length (Munroe, 2001; Laabs and Carson, 2005), whereas the westernmost valleys were occupied by outlet glaciers of the Western Uinta Ice Field, draining into the Duchesne River, Provo River, Beaver Creek, Weber River, and Bear River valleys (Fig. 1) (Munroe, 2001; Refsnider, 2006).

Increases in effective precipitation during the LGM in the western United States have been attributed to a southward

displacement of the polar jet stream and associated storm tracks by North American ice sheets (e.g., Benson and Thompson, 1987; Hostetler and Clark, 1997). Hostetler and Benson (1990) and Hostetler et al. (1994) suggested that the Lake Bonneville highstand was sustained by greater-than-modern amounts of precipitation; the latter study inferred summer temperature depressions of 7–9°C, and suggested that the lake was a moisture source for glaciers in the downwind Wasatch Mountains. Munroe and Mickelson (2002) and Munroe et al. (2006) interpreted a dramatic westward lowering of reconstructed glacier equilibrium-line altitudes across northeastern Utah to indicate that glaciers in the western Uinta Mountains may also have received lake-effect moisture derived from Lake Bonneville. Numerical glacier modeling suggests that a summer temperature depression of 7–9°C, a change estimated by previous studies (e.g., Brugger and Goldstein, 1999), would require a precipitation increase of as much as 2× modern to generate the glaciers known to have formed in the Wasatch Mountains and the westernmost Uinta Mountains (Laabs et al., 2006; Refsnider, 2006).

Although a southward displacement of the jet stream and associated storm tracks is observed in general atmospheric circulation modeling experiments (e.g., Bartlein et al., 1998), several studies of paleoclimate proxies infer a much colder and possibly drier-than-modern climate during the LGM in this region (e.g., Galloway, 1970; Brakenridge, 1978). For instance, Kaufman (2003) estimated paleotemperature in the northern Great Basin from amino-acid paleothermometry of fossil shells and concluded that climate during the period 24 to 12 ka was characterized by temperature depressions as much as 13°C colder than modern. Lemons et al. (1996) used volumes of delta sediment deposited in Lake Bonneville to infer precipitation during its highstand, and suggested that lake expansion was caused by a relative precipitation increase of less than 33%. Such a minor precipitation increase would have required a relatively large temperature depression or significantly increased cloud cover to reduce evapotranspiration in the Great Basin (and evaporation from the lake surface) sufficiently to form Lake Bonneville. Glacial reconstructions in the northern Rockies by Murray and Locke (1989) are broadly consistent with these results, suggesting that glaciers advanced under cold and dry continental climates most similar to the modern climate in the Alaskan interior ranges (e.g., the Brooks Range). Similarly, Porter et al. (1983) suggested that ice advance during the LGM in the Rocky Mountains under a colder and drier-than-modern climate would have required a temperature depression of 10°C or more. Overall, these discrepancies among regional climatic estimates for the last glaciation highlight the need for further research on paleoclimate proxies in the interior western United States. By refining the chronology of maximum ice extent in the northern Uinta Mountains, the data reported here contribute to this effort.

Setting

THE BEAR RIVER VALLEY

The Bear River drainage basin is located near the boundary of the central Rocky Mountains and the Basin and Range physiographic provinces. The river drains an area of approximately 20,000 km² (Fig. 2), and flows northward into Wyoming before re-entering Utah and debouching into Bear Lake valley. The river currently bypasses Bear Lake (except for artificial diversions) and continues flowing northward toward Soda Springs, Idaho (Fig. 2), where it turns to follow a southerly course toward Great Salt Lake. Runoff from the Uinta and

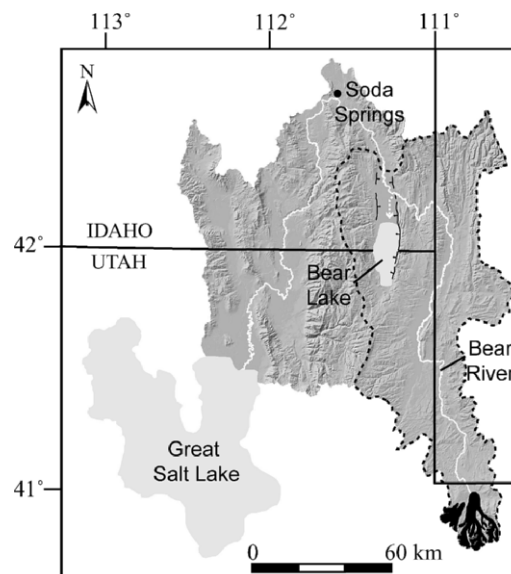


FIGURE 2. Shaded relief map of the Bear River drainage basin. Black area indicates the maximum extent of ice during the Smiths Fork Glaciation in the upper part of the basin. Black lines with ball-and-bar symbols indicate the approximate locations of normal-fault scarps on the east and west sides of Bear Lake valley (from Reheis et al., 2005), in which Bear Lake is located at the southern end. Dashed white arrow indicates the diverted flow direction of Bear River when it entered Bear Lake. Dashed black line outlines the approximate location of the Bear Lake drainage divide prior to about 17 ka. Black dot indicates the location of Soda Springs, Idaho, where the Bear River begins flowing southward en route to Great Salt Lake.

northeastern Wasatch Mountains contributes the largest volume of flow to the river; tributaries that drain minor watersheds elsewhere in Utah, Wyoming, and Idaho contribute smaller volumes.

The bedrock geology of the upper Bear River basin consists of folded Precambrian quartzite, sandstone, and shale of the Uinta Mountain Group at the core of the Uinta Range, and north-dipping Paleozoic sedimentary rocks on the north flank of the range (Bryant, 1992; Munroe, 2001; Reheis et al., 2005). Paleozoic to Cenozoic sedimentary rocks underlie most of the middle and lower parts of the basin. Surficial valley-bottom deposits include till, outwash, and alluvium in the upper part of the basin above Bear Lake valley and extensive, locally faulted lacustrine and marsh deposits within Bear Lake valley. Farther downstream, the river flows over alluvium in most areas, and locally over volcanic rocks and tufa near its northernmost point in Idaho (Reheis et al., 2005).

GLACIAL GEOLOGY OF THE UPPER BEAR RIVER DRAINAGE

At the local LGM, the Mill Creek tributary valley was occupied by a valley glacier while the East, Hayden, and West Forks of Bear River were occupied by north-flowing outlet glaciers of the Western Uinta Ice Field (Munroe, 2001) (Figs. 1 and 3). The most prominent moraines of these glaciers are found near the junction of the East Fork and Hayden Fork (Fig. 3). Inner lateral ridges merge near stream junctions, whereas the outermost lateral ridges of the Hayden and East Forks of Bear River grade northward into a broad upland of hummocky topography, suggesting that valley glaciers coalesced to form a piedmont lobe beyond the mouths of the tributary canyons (Figs. 1 and 3). The terminal moraine complex grades into

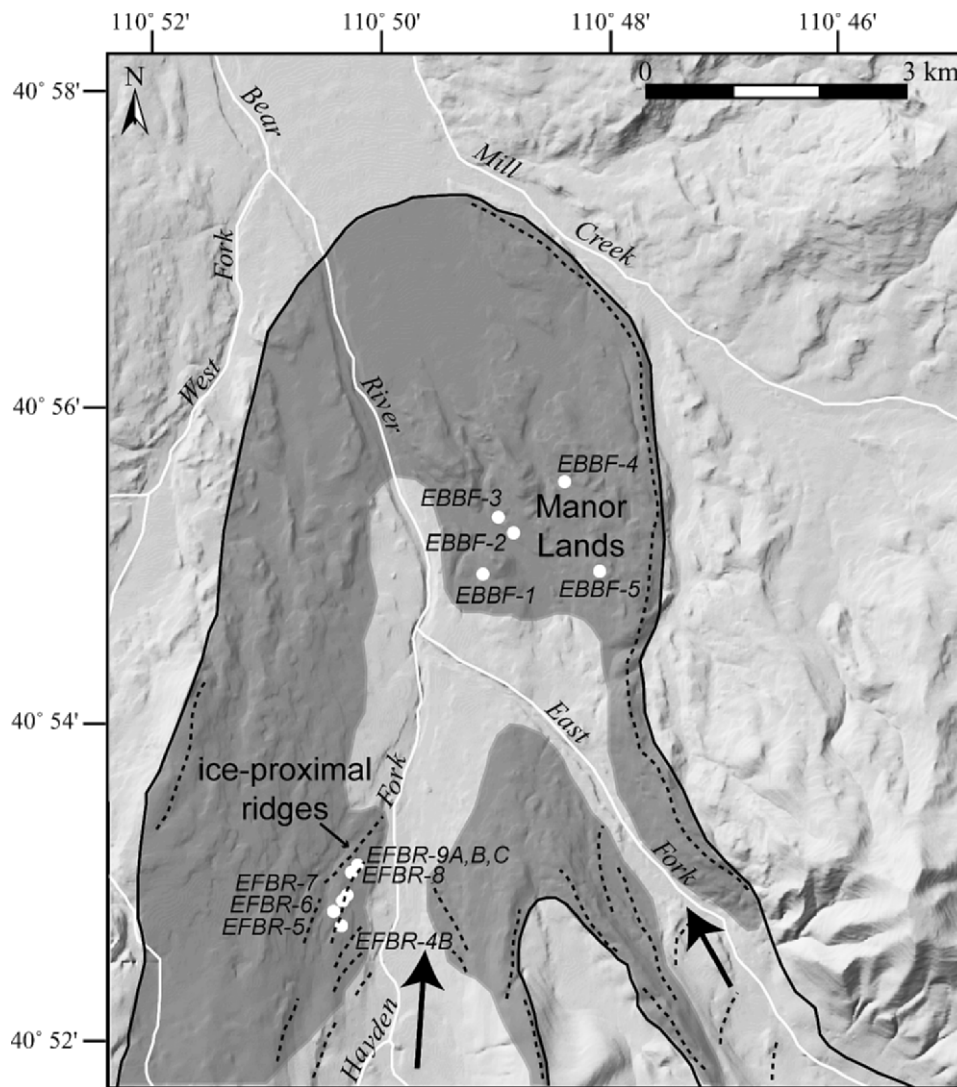


FIGURE 3. Shaded-relief map of the mouth of glaciated tributaries in the Bear River basin. Arrows indicate the general direction of ice flow in the Hayden Fork and East Fork valleys. Gray area shows the extent of the Smiths-Fork equivalent terminal moraine; dashed lines indicate moraine crests and the solid black line indicates the reconstructed LGM ice extent (modified from Munroe, 2001). White dots indicate locations of moraine boulders sampled for cosmogenic-exposure dating in the Manor Lands (the distal part of the moraine) and on an ice-proximal ridge (ridges indicated by the small arrow) in the Hayden Fork valley.

outwash that forms laterally extensive terraces along Bear River for most of its course between the Uinta Mountains and Bear Lake valley (Reheis et al., 2005). Moraines are composed of sandy till rich in quartzite clasts derived from the Precambrian Uinta Mountain Group, which provides a suitable lithology for cosmogenic-exposure dating.

Farther upvalley, recessional moraines are preserved in the main glacial troughs (Fig. 3), although till sheets and colluvium cover bedrock over much of the valley bottoms. Surface deposits become thinner at the valley heads, and cirques in these areas contain extensive bedrock ledges displaying glacial striae and polish. Cirque-floor moraines deposited during the final deglaciation of the valley are found within a few kilometers of the headwall in several tributaries.

SEDIMENTARY RECORDS FROM BEAR LAKE

Bear Lake valley is a half-graben located downstream from the formerly glaciated headwaters of the Bear River basin (Fig. 2). Motion on valley-bounding normal faults has led to subsidence in the southern end of the valley throughout the Quaternary Period (Laabs and Kaufman, 2003), resulting in the formation of a deep tectonic basin, the southern end of which is occupied by Bear Lake (Fig. 2). The lake contains a sedimentary record of at least the last ca. 250 ka (Colman et al., 2006). Sediment cores spanning the last

32 ka were collected from the lake in 1996 by workers from the University of Minnesota and the U.S. Geological Survey, and a much longer core was acquired in 2000 during testing of the GLAD800 coring platform (Dean et al., 2002). ^{14}C -based age models for the 1996 cores were developed by Colman et al. (2005) and provide chronological limits on records of paleoenvironmental change identified in the cores. These records include evidence of variations in glacial flour content, which were interpreted as indicating glacier advance and retreat in the upper Bear River basin, and of changes in the input of Bear River to Bear Lake (Dean et al., 2006). In summarizing these records herein, we report calendar-year-corrected ages from a ^{14}C -based age model of the 1996 lake sediment cores (Colman et al., 2006) (Colman et al. [2006] reported using CALIB 5.0 of Stuiver et al. [1998] to convert ^{14}C ages of Bear Lake sediment to calendar years.) for comparison with cosmogenic-exposure ages of moraine boulders. Individual calendar-year-corrected ^{14}C age estimates are in Colman et al. (2005).

Although the Bear River currently bypasses Bear Lake en route to Great Salt Lake, the river entered the lake at various times in the past including most of the last glacial cycle (Dean et al., 2006). Lacustrine sediments deposited when the river entered the lake are characterized by a reddish color and high siliciclastic content, attributed to the presence of red, brown, and purple Precambrian quartzite and sandstone in the headwaters of Bear

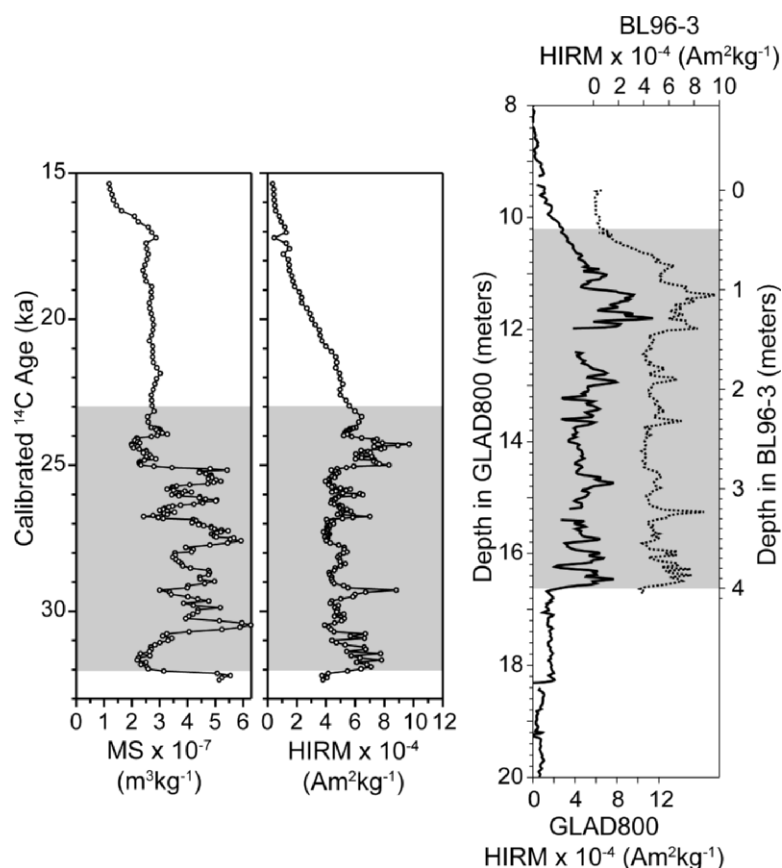


FIGURE 4. Physical properties of sediment cores retrieved from Bear Lake. Left: magnetic susceptibility (MS) and hard isothermal remanent magnetism (HIRM) of sediment from core BL96-3, with age model from Colman et al. (2005). Right: depth plots of HIRM for BL96-3 alongside HIRM for a portion of the GLAD800 core (C. Heil, unpublished data). The relatively low HIRM values for sediment at depth greater than about 16.7 m suggests that core BL96-3 captures the entire interval of glacial flour, and that the inferred onset of ice advance began at 32 ka. Gray areas indicate the interval 32 to 17 ka as estimated in core BL96-3 from a calendar year-corrected ^{14}C -age model (Dean et al., 2006).

River. This interpretation is supported by oxygen and strontium-isotope data from endogenic carbonate sediments (Dean et al., 2006). Magnetic properties of the lake sediment also reflect input of sediment derived from the upper Bear River; hard isothermal remanent magnetization (HIRM) is largely a measure of hematite content, and relatively high values of HIRM indicate sediment derived from quartzite in the Uinta Mountains. High HIRM values were interpreted by Rosenbaum (2005) to represent glacially derived sediment (i.e., glacial flour). According to the ^{14}C -based age-depth model in core BL96-3 (Rosenbaum, 2005; Colman et al., 2005), high HIRM values occur at 32 ka, just above the bottom of core BL96-3 (Fig. 4). By matching HIRM records from core BL96-3 with those in a GLAD800 sediment core retrieved from Bear Lake (Fig. 4; C. Heil, unpublished data), it is evident that HIRM values were consistently low prior to 32 ka. After this time, HIRM values remained high until about 24 ka, when maximum HIRM values occurred between 25 and 24 ka (Dean et al., 2006) and declined rapidly thereafter (Fig. 4). The decline was interrupted by a minor increase at about 23 ka, then continued until 17 ka, at which time Bear River likely abandoned Bear Lake (Dean et al., 2006). Despite the uncertainty of inferring extent of ice advance from estimates of sediment discharge, Rosenbaum and Reynolds (2004) suggest that glacial flour influx is a reliable indicator of ice extent when averaged over intervals greater than tens of years. If so, the HIRM values of Bear Lake sediments indicate that (1) glacial advance in the upper Bear River valley began at 32 ka; (2) ice reached its LGM extent at 25 ka and persisted for about 1000 years; (3) ice retreat began at 24 ka with one or more minor readvances at about 23 ka extent; and (4) ice retreat continued until at least 17 ka, at which point the river ceased to enter the lake.

Methods

COSMOGENIC SURFACE-EXPOSURE DATING

Two areas within the terminal moraine complex were considered suitable for cosmogenic surface-exposure dating on the basis of moraine morphology and abundance of large boulders on the ridge crest. The first area is characterized by low relief (<10 m), continuous, lateral and frontal ridges with gently rounded crests on the ice-proximal side of the moraine complex about 2 km south of the junction of the East and Hayden Forks of Bear River (Fig. 3). The second area is a broad upland of hummocky topography rising as much as 70 m above the outwash surface at the distal part of the moraine complex (this area is locally known as and referred to herein as the “Manor Lands”; Fig. 3). Both of these areas are vegetated by conifers, aspen, shrubs, and grasses. Based on geomorphic relationships with multiple outwash fans in the valley, the ice-proximal ridge was originally interpreted to mark the termini of valley glaciers in the Hayden Fork valley during the Smiths Fork Glaciation and the Manor Lands area was considered the terminus of a piedmont lobe during a pre-Smiths Fork glaciation (Munroe, 2001). However, our interpretation now (due in part to the findings of this study) is that these two areas are part of a compound terminal moraine deposited by the Bear River glacier during the Smiths Fork Glaciation.

Each moraine was surveyed to identify quartzite boulders suitable for cosmogenic-exposure dating. Ideally, such boulders have been exposed continuously since deposition, have not been eroded at their surfaces, have not been exhumed or overturned as a result of moraine-crest lowering, and have not been shielded by thick snow cover. Boulders that stand highest above the moraine

TABLE 1
¹⁰Be Data and Cosmogenic-Exposure Ages for the Bear River Terminal Moraine.

Sample locale*	Sample ID [†]	Elevation (m a.s.l.)	Latitude (°N)	Longitude (°W)	Boulder height (m)	¹⁰ Be/ ⁹ Be	2σ	[¹⁰ Be] (atoms · g SiO ₂ ⁻¹)	2σ	Sample thickness (cm)	Scalin g factor [‡]	Production rate [§]	Age (ka)	2σ [#]
Manor Lands	EBBF-1	2646	40.9178	110.8210	0.95	6.84E-13	7.51E-14	5.48E+05	6.01E+04	4.5	6.79	33.83	16.3	1.8
Manor Lands	EBBF-2	2593	40.9224	110.8165	0.95	7.05E-13	6.91E-14	5.83E+05	5.72E+04	6.5	6.46	32.19	18.2	1.8
Manor Lands	EBBF-3	2592	40.9241	110.8187	0.95	6.86E-13	3.81E-14	5.34E+05	2.97E+04	6.5	6.46	32.17	16.7	0.9
Manor Lands	EBBF-4	2578	40.9280	110.8090	1.12	7.79E-13	8.49E-14	5.70E+05	6.22E+04	3.5	6.56	32.68	17.5	1.9
Manor Lands	EBBF-5	2645	40.9181	110.8039	1.51	7.93E-13	9.48E-14	6.24E+05	7.46E+04	4.5	6.79	33.81	18.5	2.2
										weighted mean (n = 5; 2σ; MSWD = 1.3)				17.1 ± 0.7
Proximal ridge	EFBR-4A	2640	40.8788	110.8417	0.55	9.82E-13	3.60E-13**	7.29E+05	2.67E+05	4.5	6.76	33.68	21.8††	8.0**
Proximal ridge	EFBR-4B	2640	40.8788	110.8417	0.86	8.94E-13	6.85E-14	6.46E+05	4.95E+04	4.0	6.79	33.82	19.2	1.5
Proximal ridge	EFBR-5	2651	40.8805	110.8429	1.31	8.25E-13	7.31E-14	6.44E+05	5.71E+04	5.0	6.78	33.78	19.2	1.7
Proximal ridge	EFBR-7	2715	40.8822	110.8409	0.62	8.60E-13	1.60E-13	6.41E+05	1.19E+05	3.0	6.87	34.23	18.8	3.5
Proximal ridge	EFBR-8	2640	40.8847	110.8403	0.87	8.42E-13	5.93E-14	6.45E+05	4.55E+04	4.5	6.76	33.68	19.2	1.4
Proximal ridge	EFBR-9A	2654	40.8855	110.8394	0.87	7.87E-13	6.91E-14	5.92E+05	5.20E+04	5.0	6.80	33.84	17.6	1.6
Proximal ridge	EFBR-9B	2654	40.8855	110.8394	1.47	9.36E-13	1.07E-13	7.13E+05	8.13E+04	5.0	6.82	33.98	21.1††	2.4
Proximal ridge	EFBR-9C	2654	40.8855	110.8394	0.86	7.90E-13	7.18E-14	5.76E+05	5.24E+04	3.0	6.80	33.84	17.1	1.6
										weighted mean (n = 6; 2σ; MSWD = 1.5)				18.5 ± 0.7

* See locations of each part of the moraine in Figure 3.

† All samples are composed of quartzite or weakly-metamorphosed sandstone.

‡ Computed using CRONUS online exposure-age calculator (Balco and Stone, 2006; http://hess.ess.washington.edu/math/index_archived.html).

Analytical uncertainty only.

** The relatively high analytical and age uncertainty are due to loss of approximately 50% of sample mass during preparation of AMS targets.

†† Age considered an outlier based on MSWD statistics (see text).

crest (>0.5 m in height) with a wide base (>1.5 m wide) were targeted because they are most likely to have had the thinnest snow cover and are least likely to have been exhumed or overturned. The presence of glacial polish and/or striae was also important because these features indicate that a boulder has experienced virtually no surface erosion since it was deposited. Five boulders were sampled in the Manor Lands, and eight boulders were sampled on an ice-proximal ridge south of the mouth of the Hayden Fork of Bear River (Fig. 3, Table 1); all boulders were sampled with a sledge hammer and chisel.

Boulder samples were crushed, sieved and etched to separate 30–50 g of purified quartz. Laboratory procedures used to isolate beryllium in boulder samples were completed at the University of Wisconsin Cosmogenic Nuclide Preparation Lab following methods of Bierman et al. (2002) and Munroe et al. (2006). Ratios of $^{10}\text{Be}/^9\text{Be}$ in each sample were measured by accelerator mass-spectrometry (AMS) at the Purdue University PRIME Lab.

To calculate cosmogenic-exposure ages from measured beryllium ratios, we used a sea-level, high-latitude ^{10}Be production rate of 4.98 ± 0.34 atoms $\text{g}^{-1} \text{SiO}_2 \text{yr}^{-1}$ scaled for location and sample thickness following Balco and Stone (2006; a description of the scaling scheme is available online at http://hess.ess.washington.edu/math/index_archived.html). Because all sampled boulders displayed glacial polish, the effects of boulder-surface erosion were considered negligible, and low angles to the horizon (less than 10° in all directions) at each sample site precluded any need for production rate adjustments for topographic shielding. The heights of most sampled boulders do not exceed the maximum winter snow depth (as measured at a nearby SNOTEL station; <http://www.wcc.nrcs.usda.gov/snow/>), suggesting that snow cover may affect the production of cosmogenic nuclides in boulders. However, the actual effects of seasonal snow cover on production rates are difficult to quantify because past winter snow depths are unknown. Because the sampled ridges form local high points oriented perpendicular to the predominant wind direction, making snow accumulation unlikely and minimizing the potential for snow shielding, we assume that the ridges are mostly windswept of snow. Other effects on production rates of cosmogenic nuclides, including variations in the intensity of Earth's geomagnetic field and the position of the dipole axis, have been shown to be minimal at middle latitudes for samples that integrate the effects of a changing field over the last ca. 20 ka (Licciardi et al., 1999).

Individual cosmogenic-exposure age calculations are based on the estimated production rate of *in situ* cosmogenic ^{10}Be (Balco and Stone, 2006), the measured $^{10}\text{Be}/^9\text{Be}$ in the boulder sample, and the half-life of ^{10}Be . Ratios are corrected (<3%) for meteoric ^{10}Be measured in sample blanks. Although the systematic uncertainty of scaling production rates of *in situ* cosmogenic nuclides may be as much as 20% (e.g., Desilets and Zreda, 2001), we consider only analytical uncertainty of the AMS measurements, at least for comparing exposure ages from the Bear River moraines with others from within the interior western United States region. Accordingly, the 2σ analytical uncertainty of each individual cosmogenic-exposure age is reported here.

Results

Five cosmogenic-exposure ages from boulders atop the distal Manor Lands area of the moraine complex range from 16.3 ± 1.8 ka to 18.5 ± 2.2 ka ($\pm 2\sigma$ analytical uncertainty; Table 1, Fig. 5). This relatively narrow range of ages suggests that all sampled boulders were deposited at approximately the same time or that they have experienced similar exposure histories. The

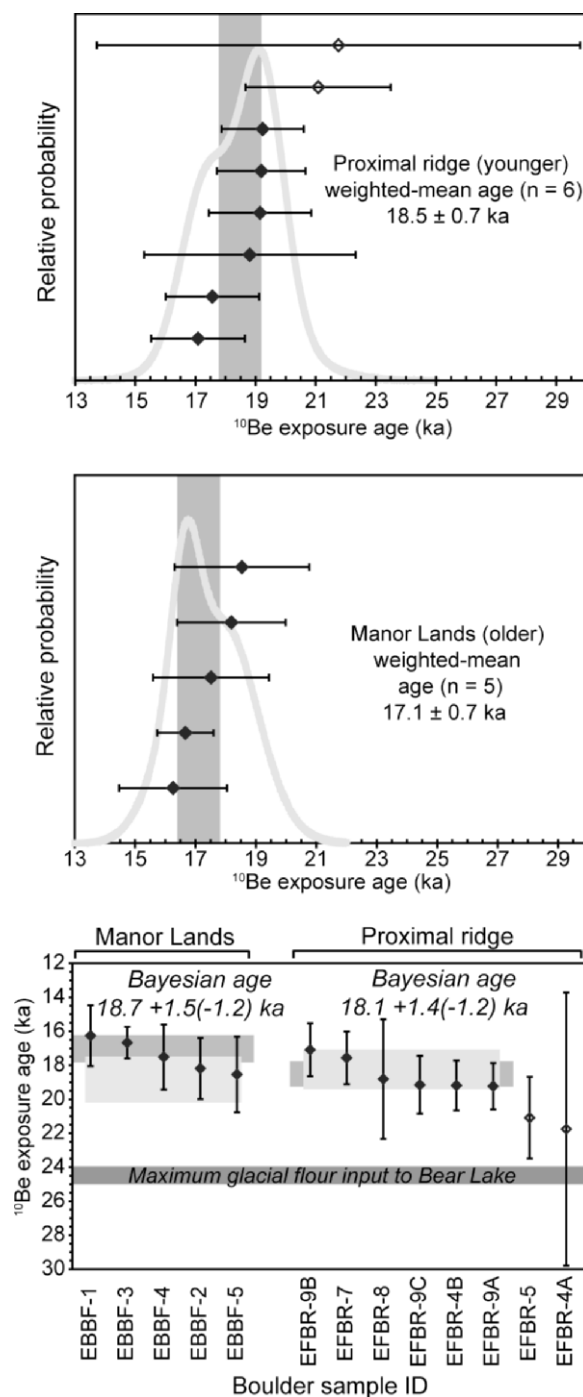


FIGURE 5. ^{10}Be cosmogenic-exposure ages of boulders on the Bear River terminal moraine. The Manor Lands part of the moraine is located approximately 5 km down-valley of the ice-proximal ridge (see Fig. 3). Top: relative probability plot of cosmogenic-exposure ages from the ice-proximal ridge; middle: relative probability plot of cosmogenic-exposure ages from the Manor Lands area; bottom: scatter plot of ages from both moraines. Error bars are 2σ analytical uncertainty of AMS analysis. Light gray bars in the bottom plot show age estimates based on a Bayesian statistical analysis (see text for explanation). Darker gray bars show the range of the error-weighted mean estimate with uncertainty at 2σ . The darkest gray bar indicates the interval of maximum HIRM values in Bear Lake sediment (Dean et al., 2006). Open symbols indicate statistical outliers (see text for explanation).

error-weighted mean of the five ages is 17.1 ± 0.7 ka (2σ). Eight cosmogenic-exposure ages from an ice-proximal ridge of the moraine near the mouth of the Hayden Fork valley range from 17.1 ± 1.6 ka to 21.8 ± 8.0 ka (Table 1, Fig. 5); the relatively broad range of these ages suggests that boulders were deposited at different times or that they have experienced different exposure histories. To identify potential outliers in this age distribution, we used the mean square of weighted deviates (MSWD) statistic of the error-weighted mean age. A MSWD value of 1 indicates expected analytical uncertainties within a set of ages, whereas values greater than 1 indicate that analytical uncertainties may be underestimated or that, in the case of cosmogenic-exposure dating, geological uncertainties are significant (Douglass et al., 2006). A minimum MSWD value of 1.5 was attained by excluding the two oldest ages (samples EFBR-9B and 4A, Table 1) from the error-weighted mean age calculation. The error-weighted mean of the remaining six cosmogenic-exposure ages from the ice-proximal ridge is 18.5 ± 0.7 ka (2σ).

Nearly all of the acceptable cosmogenic-exposure ages from the two sampled areas of the terminal moraine complex overlap at 2σ (Fig. 5), making it difficult to distinguish ages of the two sections on the moraine based on the data sets alone. Morphostratigraphic relations indicate that the Manor Lands section of the moraine must have been deposited before the ice-proximal ridge (Fig. 3), yet the oldest acceptable cosmogenic-exposure age comes from the ice-proximal ridge. If none of the moraine boulders contain inherited nuclides, this result suggests that cosmogenic-exposure ages from the Manor Lands area are younger than the actual age of moraine deposition. Thus, the weighted-mean ages do not provide stratigraphically consistent age estimates for the two sections of the moraine. In the following discussion, we explore several possible explanations for the distribution of cosmogenic-exposure ages of the terminal moraine complex and an alternative approach for determining the best age estimates for each section of it.

Discussion

INTERPRETATION OF COSMOGENIC-EXPOSURE AGES

The discrepancy in cosmogenic-exposure ages from the two sections of the terminal moraine complex could be explained by the hummocky topography of the Manor Lands area (internal relief >20 m), which suggests that melting of buried, stagnant ice followed the onset of ice retreat (e.g., Everest and Bradwell, 2003) and may have caused exhumation or overturning of boulders at the surface during a period of instability. This explanation is supported by Zech et al. (2005a, 2005b), who inferred that prolonged moraine stabilization explains broad scatter among cosmogenic ^{10}Be exposure ages of boulders from hummocky late Pleistocene moraines. However, the five cosmogenic-exposure ages from the Manor Lands do not display broad scatter; indeed, all ages overlap at 2σ (Table 1, Fig. 5). An alternative explanation of the age distribution is that boulders on the ice-proximal ridge contain inherited nuclides, thereby yielding ages older than the actual age of moraine deposition. In this case, we would also expect to find broad scatter among cosmogenic-exposure ages rather than the observed cluster of cosmogenic-exposure ages at ca. 19 ka; thus, we conclude that the sampled boulders on the Manor Lands moraines do not contain significant concentrations of inherited nuclides.

We suggest that scatter among ages from both moraines reflects analytical uncertainty only because: (1) nearly all acceptable cosmogenic-exposure ages from the Bear River terminal

moraine complex overlap at 2σ , (2) the MSWD value of the error-weighted mean of cosmogenic-exposure ages from each moraine is close to 1, and (3) no single geologic solution can explain the tight clustering of ages on the Manor Lands moraine and the distribution of ages from the ice-proximal ridge. Therefore, to resolve stratigraphically consistent age estimates for the Manor Lands and the ice-proximal ridge, we applied a Bayesian statistical analysis, which utilizes the overlap between cosmogenic-exposure ages from separate sections of the moraine and the morphostratigraphic observation that the ice-proximal ridge must post-date formation of the Manor Lands area. The Bayesian approach has been applied to ^{14}C dating (e.g., Buck et al., 1996) and more recently to optically stimulated luminescence dating (Rhodes et al., 2003) in settings where stratigraphic relationships between dated horizons are clear and can be combined with statistical uncertainties of age estimates to reduce their uncertainty. It is particularly useful in cases where ages overlap significantly. Here, we used a Monte Carlo implementation of Bayesian statistics (from Ludwig, 2003), which computes stratigraphically consistent ages based on one age estimate from each moraine. This analysis requires selection of one individual cosmogenic-exposure age (not the mean age) from each ridge to be used in the analysis. We selected the cosmogenic-exposure age of sample EFBR-9C (19.2 ± 1.7 ka) from the ice-proximal ridge and sample EBBF-4 (17.5 ± 1.9 ka) from the Manor Lands area, both of which are close to the error-weighted mean age of their respective ridge (and therefore closely represent the “mean” age of the ridge) and have analytical uncertainties similar to other samples from their respective ridge (Table 1). Applying Bayesian statistical analysis to ages at the younger or older end of the age ranges on each section of the terminal moraine would reduce the overlap between age estimates, thereby reducing the reliability of the Bayesian method. We report the modal ages and 95% confidence uncertainties as determined from the Monte Carlo iterations (both labeled “Bayesian age” in Fig. 5), which are $18.7 +1.5/-1.2$ ka and $18.1 +1.4/-1.2$ ka for the Manor Lands and ice-proximal ridges, respectively. The reduced age uncertainty (although asymmetric about the mode) relative to the individual exposure ages along with the stratigraphically consistent age estimates reflect the utility of the Bayesian method.

INTERPRETATION OF APPARENT DISAGREEMENT BETWEEN ^{14}C AND COSMOGENIC ^{10}Be AGE LIMITS

If the two Bayesian ages, $18.7 +1.5/-1.2$ ka for the Manor Lands and $18.1 +1.4/-1.2$ ka for the ice-proximal ridge, respectively, represent the time of terminal-moraine abandonment (i.e., the onset of ice retreat) in the Bear River valley, they are in obvious disagreement with the ^{14}C ages of glacial flour in Bear Lake, which suggest that the Bear River glacier reached its maximum extent at 25 to 24 ka, and began retreating shortly thereafter (Dean et al., 2006). The disagreement between the two age limits requires an explanation; here we evaluate the three that appear most likely.

- (1) Some studies have noted that cosmogenic-exposure age limits for glacial deposits in the western United States tend to be systematically younger than corresponding ^{14}C age limits; e.g., Pierce (2004) and Putkonen and Swanson (2003) suggested that the oldest cosmogenic-exposure age provides the best limit on the time of moraine abandonment. In the Bear River drainage, the slight overlap between the oldest cosmogenic-exposure age on the terminal moraine (21.4 ± 2.4 ka; Table 1) and the lower ^{14}C age limits on glacial flour

(23 ka) in Bear Lake is consistent with this theory, and suggests that even the oldest exposure age within a sample set provides only a minimum age of moraine abandonment. However, this explanation implies that shielding by snow and/or sediment yields cosmogenic-exposure ages in the Manor Lands area as much as 4000 years younger than the actual age of the moraine surface. Although the potential for snow shielding is difficult to quantify, conservative accounts for this effect on tall boulders (with heights that exceed the historical snow depth) are generally less than ~500 years (Benson et al., 2004, 2005). In addition, we observed that the ice-proximal moraine ridge has the low-relief, rounded form that Putkonen and Swanson (2003) considered least likely to experience significant lowering over a period of 20 kyr. It is, therefore, reasonable to assume that the sampled boulders have been stable since deposition, and we maintain that the Bayesian age estimates are useful limits of the actual moraine age.

Another possible explanation for disagreement between the two data sets is that pollen-based ^{14}C age limits on glacial flour in Bear Lake are older than the sediments in which the pollen was deposited. This possibility is suggested by Colman et al. (2005), who noted that some pollen samples extracted from Bear Lake sediment for ^{14}C dating were very small and that individual pollen grains appeared to be corroded (Colman et al., 2005), suggesting that some of the grains may have been temporarily stored in the drainage basin before deposition in the lake. The potential for such systematic error of the ^{14}C ages is difficult to evaluate because no other chronological data for Bear Lake sediments are available. However, all ^{14}C ages between 32 and 17 ka in the Bear Lake sediment cores are stratigraphically consistent, suggesting that large, variable systematic errors among the pollen-based ^{14}C ages, which would be expected if grains were temporarily stored in the drainage basin before being deposited in the lake, are unlikely. Other settings in which ^{14}C age limits on glacially derived sediment in a downstream lake can be compared with cosmogenic-exposure ages of moraines are few. The only locality we know of where cosmogenic-exposure ages of moraine boulders can be compared to ^{14}C age limits of macrofossils in glacially derived sediment from a downstream lake is at the center of the Yellowstone Plateau (Porter et al., 1983; sample W-2285). There, ^{14}C age limits on peat-rich mud overlying till are about 15.8 ka and the youngest cosmogenic-exposure ages from the nearby Deckard Flats moraine are 14.0 ± 0.4 ka (as reported in Pierce, 2004). This 1.8 kyr discrepancy between age estimates is significantly less than the ~5 kyr difference observed in the Bear River/Bear Lake records, suggesting that, although cosmogenic-exposure ages on moraines are generally observed to be somewhat younger than corresponding ^{14}C age limits in downstream lakes, significant offset between the two data sets (of several thousand years) should not be an expected result.

An additional explanation, which we find most convincing, assumes that both the ^{14}C and cosmogenic ^{10}Be exposure ages are accurate, but that the glacial signal in Bear Lake sediment may be somewhat complex. The HIRM record from Bear Lake provides unequivocal evidence of increased input of sediment from the headwaters of Bear River during the interval 32 to 24 ka (Dean et al., 2006), but it is not clear whether declining inputs of sediment with high HIRM values after 24 ka necessarily indicate the final ice retreat from the terminal moraine. For example, the decline may

instead represent a decrease in basal meltwater and/or sediment production by the glacier, or an increase in sediment input from the local Bear River catchment. Rosenbaum (2005) interpreted increasing MS values to represent a rising influx of sediment from the local Bear Lake catchment beginning at about 20.3 ka (Fig. 4) and accelerating rapidly at 18.5 ka (see Fig. 1 in Rosenbaum, 2005). This relative increase may represent a declining input of glacial flour, or just a declining input of Bear River sediment. Viewed this way, the sedimentary records in Bear Lake indicate that the Bear River glacier reached its maximum extent by about 24 ka, but indicate relatively little about deglaciation in the river headwaters. The cosmogenic-exposure ages of the Bear River terminal moraine therefore provide more useful limits on the time of deglaciation, and suggest that ice occupied or fluctuated near its terminal moraine until about 18.7 to 18.1 ka.

We favor this explanation of the chronological data primarily because (1) we find no compelling reason to disregard either the ^{14}C or the cosmogenic-exposure age limits on glacial sediments in Bear River valley, and (2) the explanation is based on interpretations of multiple properties of Bear Lake sediment, not just the HIRM data. The suggestion that the Bear River glacier was at or near its terminal moraine between about 24 and 18 ka is somewhat speculative, but consistent with interpretations of glacial records in other ranges of the central Rocky Mountains; for example, Gosse et al. (1995) described chronological evidence that the Fremont glacier in the nearby Wind River Range fluctuated near its terminal position during the period 26 to 19 ka (^{10}Be cosmogenic-exposure ages increased by 8% based on the production rate used here).

SIGNIFICANCE OF AGE LIMITS AND IMPLICATIONS FOR PALEOCLIMATE

The ^{14}C ages from Bear Lake indicating growth of glaciers in the Uinta Mountains beginning at about 32 ka and a local LGM at about 24 ka are consistent with limits on the global LGM (e.g., Peltier and Fairbanks, 2006) and on the LGM elsewhere in the Rocky Mountains. For instance, the chronology of late Pleistocene glaciation in the Sierra Nevada Range (Benson et al., 1996; Bischoff and Cummins, 2001) suggests the onset of the Tioga advance occurred at about 32 ka. In the Colorado Front Range, ^{14}C -based glacial chronologies indicate that ice advance began before 30 ka (Nelson et al., 1979) and culminated between 27 and 24 ka (Madole, 1980; Rosenbaum and Larson, 1983). Other cosmogenic-exposure ages consistent with the Bear Lake results are yielded from terminal moraines in the Wind River Range, where Benson et al. (2005) report an oldest cosmogenic ^{10}Be surface-exposure age of 24.8 ka (ages increased by 8% for the production rate used here). Benson et al. (2005) also documented cosmogenic ^{36}Cl exposure ages from the Colorado Rockies, with oldest ages near 21 ka in several northern ranges and 21.5 ka in the San Juan Mountains (Benson et al., 2005). Brugger (2006, 2007) reported cosmogenic-exposure ages as old as about 22 ka from moraines in the Sawatch Mountains and Taylor River Range of central Colorado, based on measured concentrations of ^{10}Be and ^{36}Cl nuclides. On the Colorado Plateau, Marchetti et al. (2005) documented cosmogenic ^3He exposure ages of moraine boulders that range from 23.1 ka to 20.0 ka. Assuming that none of the above cosmogenic-exposure ages are from boulders with inherited nuclides, all of these ages suggest that outlet glaciers of the Western Uinta Ice Field in the Bear River

valley advanced approximately in phase with glaciers elsewhere in the region.

The suggestion that the Bear River glacier fluctuated near its maximum position between 23 and 18 ka is also consistent with reports of millennial-scale climatic forcing in the region. Most notably, Oviatt (1997) described a temporal correspondence between oscillations of Lake Bonneville and Heinrich events in the North Atlantic. If Heinrich events affected climate in western North America, as suggested by Clark and Bartlein (1995), then the Bear River glacier may have responded on similar time scales. As noted by Oviatt (1997), the abrupt warming following Heinrich events (e.g., Bond and Lotti, 1995) may have impacted the hydrologic balance of Lake Bonneville and the mass balance of glaciers in neighboring mountain ranges. Alternatively, iceberg discharges into the North Atlantic may have disrupted northern hemisphere atmospheric circulation patterns in such a way that westerly storm tracks associated with the polar jet stream shifted course, causing an overall decrease in the water budget for the Lake Bonneville region. If the presence of Lake Bonneville augmented snowfall over glaciers in the western Uinta Mountains (Munroe and Mickelson, 2002; Munroe et al., 2006), changes in the surface area of the lake would have directly impacted glacier mass balance.

Cosmogenic-exposure ages limiting the start of deglaciation in the Bear River valley to about 18.7 to 18.1 ka are generally consistent with ages for the termination of the local LGM from elsewhere in the Uinta Mountains and in other Rocky Mountain ranges. For example, Munroe et al. (2006) and K. Refsnider (written communication, 2007) reported weighted-mean cosmogenic ^{10}Be exposure ages of moraines limiting the onset of deglaciation in the southwestern Uinta Mountains of 16.6 ± 0.7 ka in the Lake Fork valley and 17.0 ± 1.1 ka in the North Fork Provo valley (Fig. 1; both ages adjusted for the production rate and exposure-age calculation scheme used here). In the northern Rocky Mountains, Licciardi et al. (2004) reported a weighted-mean age for an ice-proximal ridge of a terminal moraine in the Wallowa Mountains (Oregon) of 17.0 ± 0.3 ka, and Licciardi et al. (2001) reported cosmogenic-exposure ages for a terminal moraine on the Yellowstone Plateau with weighted means of 16.2 ± 0.3 (^{10}Be) and 16.5 ± 0.4 (^3He). In the central and southern Rocky Mountains, the mean of cosmogenic ^{36}Cl exposure ages from terminal moraines in north-central Colorado is 18.4 ka and 18.9 ka for the San Juan Mountains, and that for ^{10}Be surface-exposure ages from terminal moraines in the Wind River Range (near Pinedale, Wyoming) is 19.1 ka (Benson et al., 2005; ages increased by 8% for the production rate used here).

Deglaciation of the southwestern Uinta Mountains at 16.6 ± 0.7 ka (Munroe et al., 2006) was approximately coincident with the climate-driven regression of Lake Bonneville from the Provo shoreline at ca. 16 ka (age calibrated from Oviatt, 1997, using CALIB 5.0; Stuiver et al., 2006), yet final ice retreat in the Bear River valley began at about 18.7 to 18.1, suggesting that the northern part of the Western Uinta Ice Field destabilized while Lake Bonneville was at its hydrologic maximum. While this sequencing appears inconsistent with the hypothesis that Lake Bonneville augmented precipitation in the western Uinta Mountains during the lake highstand (Munroe and Mickelson, 2002; Munroe et al., 2006), it might reflect variations in local atmospheric circulation patterns that increased moisture delivery to the southwestern Uinta Mountains relative to the Bear River drainage. For example, if the dominant airflow followed a southwest to northeast course across the Great Basin, moisture derived from the surface of Lake Bonneville may have nourished

glaciers in the southwestern Uinta Mountains whereas the Bear River glacier existed in an orographic precipitation shadow (Fig. 1). Such moisture may have sustained glaciers in the southwestern Uinta Mountains at their maximum positions for several millennia during the Lake Bonneville highstand (19 to 16 ka), while glaciers elsewhere in the range (including the Bear River headwaters) fluctuated in response to temperature changes and/or changes in the surface area of the lake driven by effective moisture variations. In any event, retreat of the Bear River glacier from its terminal moraine at about 18.7 to 18.1 ka was coincident with retreat in other mountain ranges worldwide (cf. Schaefer et al., 2006).

Summary and Conclusions

The combination of cosmogenic ^{10}Be surface-exposure ages of moraine boulders and a ^{14}C chronology of glacially derived sediment in Bear Lake provides useful limits on the timing of the LGM in the Bear River drainage basin. Although we acknowledge potential systematic errors of both data sets, we propose that disagreement between them is not cause to reject results of either dating method. Instead, we suggest the ^{14}C ages of glacial flour retrieved from Bear Lake best limit the onset of ice advance in the headwaters of Bear River and the culmination of glaciation in the drainage, whereas cosmogenic-exposure ages of the Bear River terminal moraine provide age limits for the start of deglaciation.

Viewed in this interpretative framework, the two data sets allow the history of the Bear River outlet of the Western Uinta Ice Field to be reconstructed. The initial expansion of the ice field began at about 32 ka in response to global climatic cooling, and glacial flour generation was abundant while outlet glaciers advanced and coalesced on the piedmont, reaching their maximum extent at 25 to 24 ka. The Bear River glacier likely fluctuated near its terminal moraine in response to millennial-scale climate forcing during the interval 23 to 18 ka. Glacial flour production during this interval was apparently diminished. Cosmogenic-exposure ages from the terminal moraine complex indicate that the Bear River glacier retreated at about 18.7 to 18.1 ka. Ice retreat in this valley, therefore, began before Lake Bonneville dropped from the Provo shoreline (ca. 16 ka, Oviatt, 1997; or possibly as late as ca. 13 ka, Godsey et al., 2005). Thus, it appears that moisture derived from Lake Bonneville, which may have nourished glaciers in the Uinta Mountains until the lake began its hydrologic fall, was less successful at reaching glaciers in the northwestern sector of the range.

Because the Bear River began to bypass Bear Lake by about 17 ka (Dean et al., 2006), sediment in the lake does not provide a record of upvalley glacial activity after this time. Therefore, additional exposure dating of recessional moraines in the headwaters of the Bear River valley is necessary to gain further insight to whether spatially restricted advances occurred after 18.1 ± 1.4 – 1.2 ka in response to the lingering presence of Lake Bonneville or the Younger Dryas cooling event.

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