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Glacial Geology and Equilibrium Line Altitude Reconstructions for the Provo River Drainage, Uinta Mountains, Utah, U.S.A.

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Abstract

The Provo River drainage in the western end of the Uinta Mountains was glaciated repeatedly during the Pleistocene, and glacial deposits from the Smiths Fork and Blacks Fork glaciations (Pinedale and Bull Lake equivalents, respectively) are well preserved throughout the area. Reconstruction of the Smiths Fork ice extent based on air photo analysis and field mapping reveals that the broad upland surfaces in the Provo River drainage were covered by an ice field from which distributary glaciers emanated. This ice field also covered parts of the Weber and Bear River drainages to the north and the North Fork Duchesne drainage to the east. Equilibrium line altitudes for glaciers in the Provo River drainage were ~2900 m a.s.l., consistent with previous studies which recognized a dramatic decrease in equilibrium line altitudes toward the western end of the range. End moraine sequences and hypsometric differences between glaciers in the Provo River drainage suggest that ice retreat rates likely differed considerably among the glaciers, reflecting variable dynamic responses to a similar climate forcing during deglaciation.

Introduction

The Uinta Mountains are an east-west–trending range that extends 200 km from the Wasatch Mountains in northeastern Utah into extreme northwestern Colorado (Fig. 1). Whereas no glaciers are found in these mountains today, the western half of the range was repeatedly glaciated during the Pleistocene (Atwood, 1909; Oviatt, 1994; Laabs and Carson, 2005; Munroe, 2005). Two major ice advances identified by Atwood (1909) were subsequently termed the "Blacks Fork" and "Smiths Fork" glaciations and correlated with the classic Wind River Range Bull Lake and Pinedale advances, respectively (Bradley, 1936), with the Pinedale glaciation taking place during the Last Glacial Maximum (LGM). Evidence also exists for an earlier, more extensive glacial advance, designated the Little Dry glaciation in the northern Uintas (Bradley, 1936) and the Altonah glaciation in the southern Uintas (Laabs and Carson, 2005). The Blacks Fork glaciation, though it has not been directly dated, likely occurred during Marine Isotope Stage (MIS) 6 and in places was slightly more extensive than the Smiths Fork glaciation (MIS 2). Munroe et al. (2006) dated the onset of deglaciation following the *local* LGM (we will subsequently use LGM to refer to the local Last Glacial Maximum) on the south slope of the Uinta Mountains at 16.8 ± 0.7 ka (95% confidence) using cosmogenic ^{10}Be surface-exposure ages from moraine boulders.

The western end of the Uinta Mountains can be divided into four major drainages. The Provo and Weber Rivers, as well as the smaller Beaver Creek, flow westward, the Bear River drains to the north, and the North Fork Duchesne River drains to the south (Fig. 1). The landscape in the western Uintas is one of broad upland basins separated by isolated peaks and arêtes (Fig. 2). Repeated Pleistocene glaciations have scoured away most cirque headwalls, leaving behind only low, broad cols. Previous mapping in the western end of the range suggests that during the LGM, an ice field covered the broad upland surfaces and fed a series of confluent distributary glaciers, whereas farther east in the range, glaciers were restricted to discrete cirques and valley systems (Fig. 1; Atwood, 1909; Oviatt, 1994; Laabs and Carson, 2005; Munroe, 2005). Munroe and Mickelson (2002), Munroe et al. (2006), and Oviatt (1994) also documented that reconstructed LGM equilibrium line altitudes (ELAs) decreased dramatically toward the western end of the range. However, no detailed mapping of glacial features or past ice extents in the Provo River drainage has previously been conducted. In this paper, we describe the glacial geology of the Provo River drainage and present reconstructions of the glaciers that occupied the area during the LGM.

Methods

GEOLOGIC MAPPING

Surficial geology was mapped using 1:40,000 scale aerial photographs, 1:24,000 scale topographic maps, and 30-m digital elevation models (DEMs). Preliminary mapping was checked in the field, and indicators of ice flow direction were measured where possible. All deposits, landforms, and features mapped were digitized in a Geographic Information System (GIS).

GLACIER RECONSTRUCTIONS

Former ice extents in the Provo River drainage were mapped using terminal and lateral moraines and trim lines that could be traced upvalley from terminal moraines. Above the equilibrium line, the approximate position of the Smiths Fork glacier margins was determined using the extent of diamicton with Smiths Fork–type characteristics (discussed in more detail below), erratic boulders, and minor changes in slope angle on the flanks of higher peaks that stood as nunataks during the LGM. Glacier hypsometry was calculated in a GIS using elevation intervals of 100 m for the combined North Fork Provo, Main Fork Provo, and Soapstone Basin glacier complex. The locations of the 100-m

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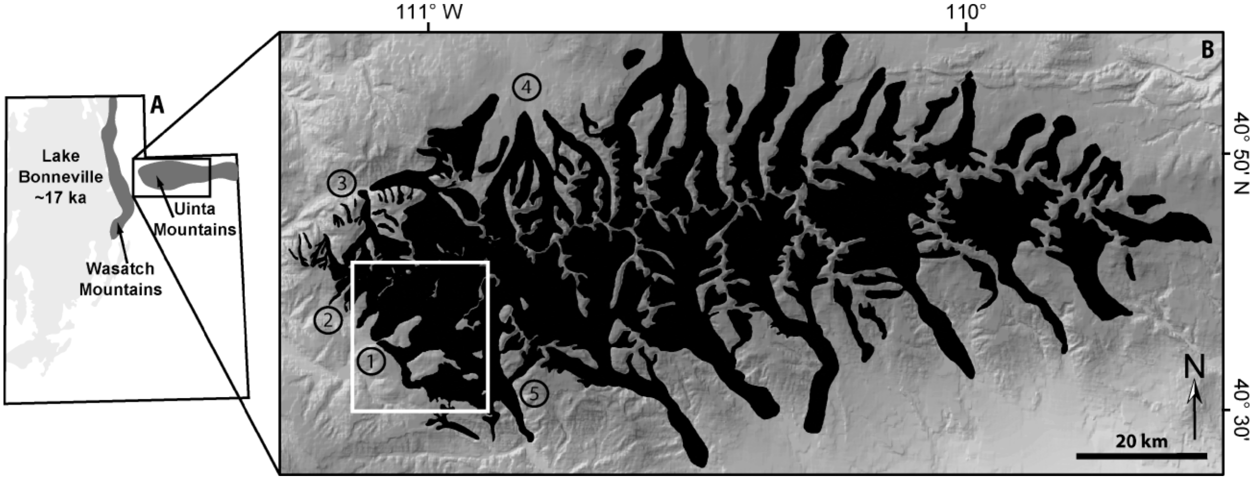

FIGURE 1. (A) Location of the Uinta Mountains, the Wasatch Range, and pluvial Lake Bonneville in Utah. (B) The reconstructed ice extent in the Uinta Mountains during the LGM. Numbers denote major watersheds in the western part of the range: 1—Provo River; 2—Beaver Creek; 3—Weber River; 4—Bear River; 5—North Fork Duchesne River. The white rectangle indicates the mapping area for this study; see Figure 3 for greater detail of this area.

ice surface contours were estimated based on ice margin elevations and ice flow direction, though this method may result in a minor underestimation of the surface elevation below the ELA and an overestimation above the ELA. Because of more variable flow directions along the range crest, ice thickness estimates become increasingly subjective at the highest elevations, so the uncertainty in glacier hypsometry is greatest in these areas.

PALEO-EQUILIBRIUM LINE ALTITUDE DETERMINATION

The altitude of a glacier equilibrium line during the Smiths Fork glaciation was determined from a weighted mean of estimates of ELAs using the accumulation-area ratio (AAR), toe-to-headwall altitude ratio (THAR), and the highest elevation of lateral moraines (LM) using the methods of Meierding (1982) and Munroe et al. (2006). Typical AAR values range from 0.50 to 0.80 for modern glaciers in equilibrium with the current climate (Meier and Post, 1962). Glaciers flowing from ice fields, which have large accumulation areas relative to their ablation areas, generally have a higher AAR (0.7 to 0.8), and piedmont glaciers, with relatively large ablation areas relative to their accumulation areas, generally have a lower AAR (0.5 to 0.6; Meier and Post, 1962; Leonard, 1984). A median AAR value of 0.65 is often used for valley glaciers with a relatively equally distributed hypsometry (e.g., Porter, 1975; Leonard, 1984; Murray and Locke, 1989; Munroe and Mickelson, 2002; Brugger, 2006). The hypsometry of the glacier system in the Provo River drainage led us to use a value 0.75; possible implications of this choice are discussed below. The AAR ELA estimate was calculated in a GIS using a simple method in which the polygon representing the ice extent was split iteratively at an elevation along the ice margin until an AAR of 0.75 was achieved. As long as the ELA was not near a confluence of tributary glaciers, we assume that the ice surface elevation at the ELA is closely approximated ($\pm$5 m) by the elevation at the ice margin.

The THAR method we applied incorporates the elevation at the crest of the terminal moraine and the highest elevation at which a cirque headwall steepens to an angle greater than 60°.

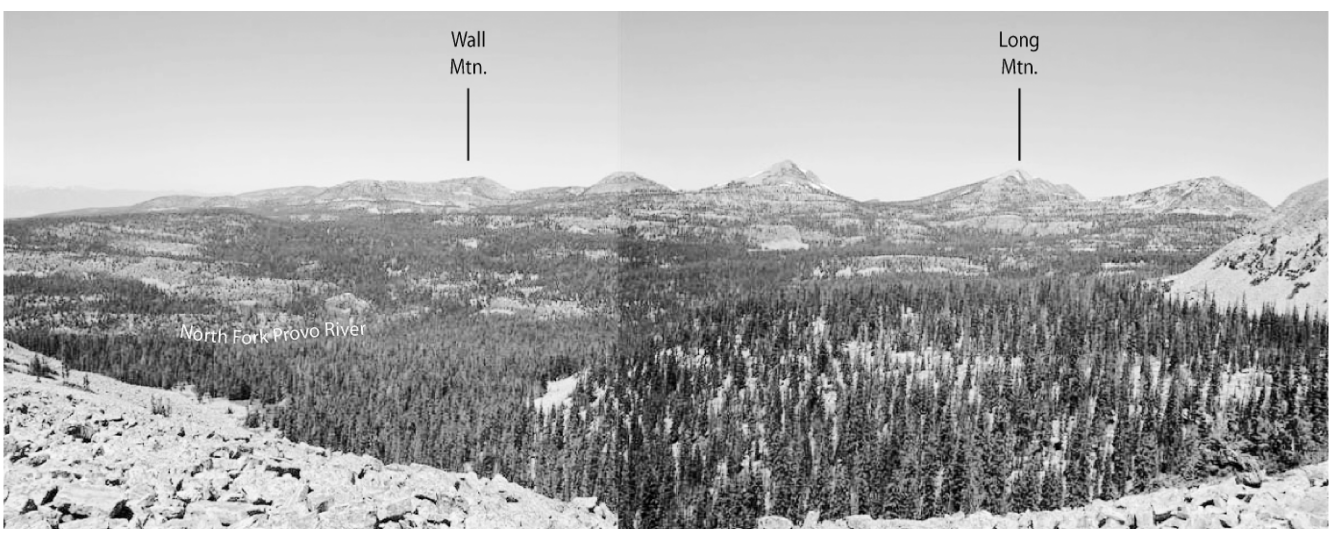

FIGURE 2. Looking northwest across the North Fork Provo drainage (as seen from location 2 in Fig. 3). Wall Mountain was at the western boundary of the ice field, and the peaks to the right on the horizon form the divide between the Provo and Weber River drainages. Note the broad, U-shaped cols between each summit, indicative of intense episodes of glaciation during the Pleistocene.

Cirque and valley glaciers often have a THAR of 0.35 to 0.40 (Meierding, 1982; Murray and Locke, 1989), but for relatively narrow glaciers with broad accumulation areas like those in the Provo River drainage, a higher THAR is often more representative (Péwé and Reger, 1972). Accordingly, we used a THAR of 0.5.

The LM ELA estimate is the least reliable method for determining paleo-ELAs, especially in places where valley walls are steep, as they are in much of the Provo River drainage. On steep slopes, lateral moraines can be rapidly removed by erosion or may never have been deposited (Meierding, 1982). For these reasons, the LM method may underestimate the ELA (Fig. 3) and is weighted the least. A weighted ELA taking into account the reliability of each method was calculated using weighting factors of 3 for AAR, 2 for THAR, and 1 for LM (Munroe et al., 2006).

Results

GLACIAL FEATURES

The high basins in the Provo River drainage slope gently upward to the north and contain many small lakes, often found in former cirque basins and in broad cols where headwalls have been eroded away. Mountain peaks were left standing as nunataks in an ice field that fed glaciers flowing north into the Weber and Bear River drainages and south into the Provo and North Fork Duchesne drainages (Fig. 3). Indicators of flow direction show that ice flow converged toward the broad U-shaped valleys of the North Fork and Main Fork Rivers, which are incised into the uplands (Refsnider, 2006). The ice flow divides at the head of the Provo River drainage were generally only several hundred meters south of the modern topographic divide separating the Provo River and Weber River basins. Ice also flowed into the North Fork Duchesne drainage through the low divide north of Murdock Basin.

We have identified glacial deposits that represent the occurrence of at least two major glaciations in the Provo River drainage. The upper reaches of the Provo River drainage have a nearly continuous cover of Smiths Fork–age diamicton. In the highest basins, this diamicton forms a thin cover over ice-sculpted bedrock, and in the middle elevations, diamicton of varying thickness is widespread. The surface topography in most of these middle elevations mimics the underlying bedrock topography. Older glacial deposits are preserved in the northern end of Pine Valley below the confluence of the Main Fork and North Fork Provo Rivers (Fig. 3). Here a series of subdued, disjunct ridges with comparatively few boulders has previously been interpreted as Blacks Fork–age deposits by Atwood (1909), who based his interpretations on morphology. More recently, Sullivan et al. (1988) used soil profile development indices to estimate the age of these deposits as Bull Lake, and Bryant (1992) reached a similar conclusion, mapping the deposits as pre-Pinedale till. On the west side of the Pine Valley and south of the extent of Blacks Fork ice (Fig. 3), sporadic sub-rounded quartzite erratic boulders are mixed with very angular colluvium derived from upslope. These sub-rounded boulders and cobbles are nearly 100 m above the valley floor, which is likely too high for them to have been deposited by a higher stage of the Provo River, so a glacial origin is the only other plausible explanation. This may be an indication that the Blacks Fork terminal moraine was actually farther downvalley, or perhaps these boulders were deposited in a more extensive pre–Blacks Fork glaciation, as documented elsewhere in the Uinta Mountains by Bradley (1936), Munroe (2005), and Laabs and Carson (2005).

Moraines of Smiths Fork age are generally sharp-crested features and contain many boulders, whereas the older Blacks Fork moraines are more subdued and have relatively fewer boulders (Atwood, 1909; Munroe, 2005; Laabs and Carson, 2005). Moraine ridges from Smiths Fork ice advances are found throughout the mapping area and are generally well preserved. These moraines are composed of unsorted and unstratified sandstone and quartzite material, ranging in size from fine sand to boulders.

The Smiths Fork end moraines deposited in the Main Fork Provo Valley abut the end moraines deposited at the mouth of the North Fork Provo Valley (Fig. 3). Preservation of the moraines in the Main Fork Provo Valley is substantially poorer, as they have been affected by meltwater stream erosion during deglaciation and subsequent erosion by the Main Fork Provo River. Lateral moraines are not well preserved in this valley except on the south slope upvalley of the former confluence of the Main Fork Provo and Soapstone glaciers. Measurements of cosmogenic ^{10}Be concentrations in samples from 12 boulders on the outermost moraine in the North Fork Provo Valley show that the feature was deposited by ~19 ka, confirming a Smiths Fork age for the moraine (Refsnider, 2006).

Moraines deposited in Soapstone Basin were not recognized in previous mapping by Atwood (1909) or Bryant (1992), though the latter did identify a small area of Pinedale-age (Smiths Fork) till in this area. Lateral moraines on the west side of the Soapstone Creek valley are continuous for more than 5 km, though sections of these moraines have been reduced to broad, round-crested forms due to fluvial and hillslope processes. However, the sharp-crested morphology and high degree of preservation of some of these lateral moraines, specifically those deposited in Soapstone Basin on steep slopes immediately south of the Main Fork Provo Valley (Fig. 3), suggests that these are Smiths Fork rather than Blacks Fork in age.

A spectacular series of nested moraines is found in the Main Fork Provo Valley upstream from Iron Mine Basin (Fig. 3). More than 30 sharp-crested, closely spaced ridges were identified, most of which are 2–4 m tall and separated from adjacent moraines by 10–20 m. These moraines may have been deposited by annual readvances or by repeated brief stabilizations of the terminus position during deglaciation.

One of the most striking glacial landforms in the mapping area is a narrow, steep-sided medial moraine cut by sizeable meltwater channels on the divide between Boulder Creek and the North Fork Provo valleys. It is difficult to determine whether this moraine was deposited during the LGM or later during retreat when the ice surface was lower, but meltwater streams subsequently cut channels 10–15 m in width across the northern end of the moraine, and kame terraces up to 10 m in width were deposited below the moraine crest (Fig. 4). The meltwater drainages terminate abruptly on either side of the divide, but they slope downward to the west over their entire length. More rapid deglaciation in the North Fork Provo drainage relative to that in Boulder Creek Canyon, discussed below in the context of a rising ELA, likely supplied the meltwater responsible for cutting these channels across the divide and depositing the large kame terraces along the margin of the more slowly receding ice in Boulder Creek Canyon.

LGM GLACIER RECONSTRUCTION

A reconstruction of past ice extents in the Provo River drainage, based primarily on geomorphic evidence, is presented in

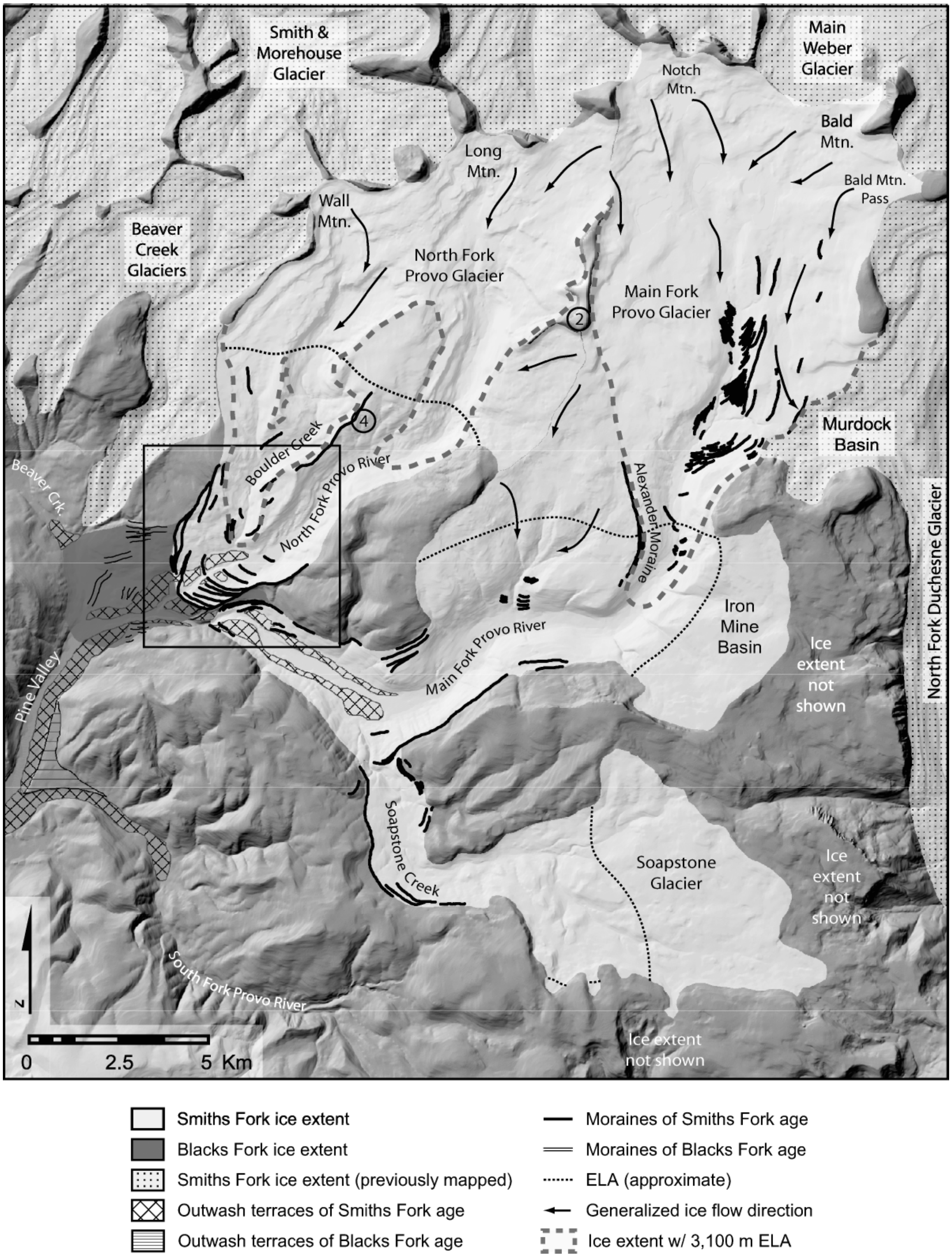

FIGURE 3. Ice extent in the upper Provo River drainage during the local LGM. The ages of deposits are based on moraine morphology, cosmogenic ^{10}Be surface-exposure dating of boulders on the North Fork Provo terminal moraine (Refsnider, 2006), and previous age estimates (Atwood, 1909; Sullivan et al., 1988; Bryant, 1992). The approximate positions of equilibrium lines, based on a weighted average of the AAR, THAR, and LM approaches, are shown with dotted lines. Evidence of small glaciers in the South Fork Provo drainage south of Soapstone Basin has not yet been mapped in detail and is not included in this figure. Also, glacial activity on the broad divides between the Provo River and North Fork Duchesne drainages appears to have been more extensive than determined by Laabs and Carson (2005), so the ice-covered areas on the east side of these divides require additional mapping. The thick dashed lines show the estimated ice extent if ELAs rose ~200 m from their LGM position to 3100 m a.s.l. The circled numbers 2 and 4 refer to the locations from which the photos in Figures 2 and 4 were taken.

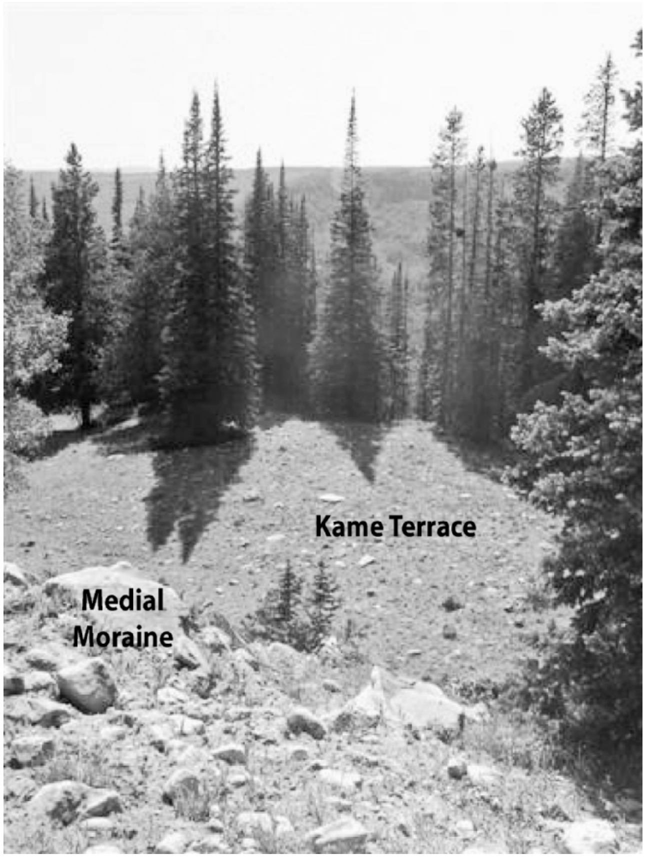

FIGURE 4. A 20-m-wide kame terrace above Boulder Creek (as seen from location 4 in Fig. 3). The ridge in the immediate foreground is a medial moraine of Smiths Fork age deposited by ice flowing down the North Fork Provo and Boulder Creek Canyons.

Figure 3. The total glaciated area during the LGM was approximately 185 km^2. Ice margin positions are clearly marked by lateral moraines and trimlines in most of the ablation zone. Above the equilibrium line, however, the elevation of the ice margin is more uncertain. Moraine ridges from both Smiths Fork and Blacks Fork glaciations and related outwash terraces are preserved throughout the North Fork and Main Fork Provo Valleys. Based on the reconstructed ice extent for the Smiths Fork glaciation, the hypsometry for the combined North Fork Provo, Main Fork Provo, and Soapstone Glacier system is plotted in Figures 5A and 5B. The termini for the North Fork Provo and Main Fork Provo Glaciers were at approximately 2300 m a.s.l., and the upper limit of ice, discussed below, was likely 3360 m a.s.l. in the vicinity of Bald Mountain. The only evidence, though not conclusive, for ice thicknesses in the divides between the Provo and Weber River drainages are subtle breaks in slope on the summit ridges of Notch and Wall mountains, 4 km west and 14 km west-southwest of Bald Mountain, respectively (Fig. 3). At Notch Mountain, the upper ice limit was approximately 3350 m a.s.l., and evidence at Wall Mountain suggests an ice limit of 3320 m a.s.l.

PALEO-EQUILIBRIUM LINE ALTITUDE ESTIMATION

Equilibrium line altitudes reconstructed using the AAR, THAR, and LM methods are listed in Table 1. Using an AAR of 0.75, the ELAs for the North Fork Provo and Main Fork Provo

TABLE 1

Equilibrium line altitudes (ELAs) (m a.s.l.). THAR = toe-to-headwall altitude ratio, AAR = accumulation-area ratio.

Method	North Fork Provo Glacier	Main Fork Provo Glacier
Highest lateral moraine (LM)	2948	2803
Elevation of terminal moraine	2320	2334
Elevation of headwall	3350	3360
THAR = 0.5	2835	2852
AAR = 0.65	2950	2950
AAR = 0.75	2910	2910
Weighted ELA, AAR = 0.65*	2910	2890
Weighted ELA, AAR = 0.75*	2890	2870

* Weighted such that AAR = 3, THAR = 2, LM = 1.

Glaciers were ~2950 m a.s.l. Using the THAR method with a ratio of 0.5, the ELAs for the North Fork Provo and Main Fork Provo glaciers were 2835 and 2852 m a.s.l., respectively, and the maximum elevations of lateral moraines are 2948 and 2803 m a.s.l., respectively. The weighted ELAs for the North Fork Provo and Main Fork Provo Glaciers were 2891 and 2873 m a.s.l., respectively.

Discussion

LGM GLACIER RECONSTRUCTIONS

Munroe and Mickelson (2002) describe the pattern of glaciation during the LGM in the Uinta Mountains as small valley glaciers in eastern and central drainages with more complex glaciers and broader accumulation areas in basins farther west in the range. The glacier reconstructions presented here (Fig. 3) are consistent with such a pattern. Previous studies described the western end of the Uinta Mountains as covered by an ice cap (Atwood, 1909; Barnhardt, 1973) or ice field (Oviatt, 1994) during the LGM. Our results agree with these previous studies, but as Oviatt (1994) argued, it should be referred to as an ice field, because ice flow was controlled primarily by the underlying bedrock topography. The ice field (Fig. 3), which we call the Western Uinta Ice Field, fed distributary glaciers that flowed north into the Bear River drainage (Munroe, 2005), west into the Weber (Oviatt, 1994) and Provo River drainages (Refsnider, 2006), and south into the North Fork of the Duchesne and Rock Creek drainages (Laabs and Carson, 2005).

Atwood (1909) and Barnhardt (1973) both commented on the difficulty associated with reconstructing the vertical limit of the ice field, because nearly all high-elevation slopes are covered by talus, relatively devoid of vegetation, and reveal few surfaces with preserved striations or polish. Striations and similar features found in the resistant Uinta Mountain Group quartzite in most cols between the Provo and Weber drainages provide evidence of a considerable thickness of ice during the LGM at the modern topographic divides. In the Bald Mountain area, Atwood (1909) suggested that the ice margin coincided with the modern upper tree line, which is at approximately 3260 m a.s.l. However, striations in cols at this same elevation provide evidence that ice in these divides was sufficiently thick to support widespread erosion due to basal sliding, making Atwood's (1909) estimate of a 3260 m a.s.l. ice limit unlikely. Barnhardt (1973) suggested alternatively that the ice surface in the cols was at approximately

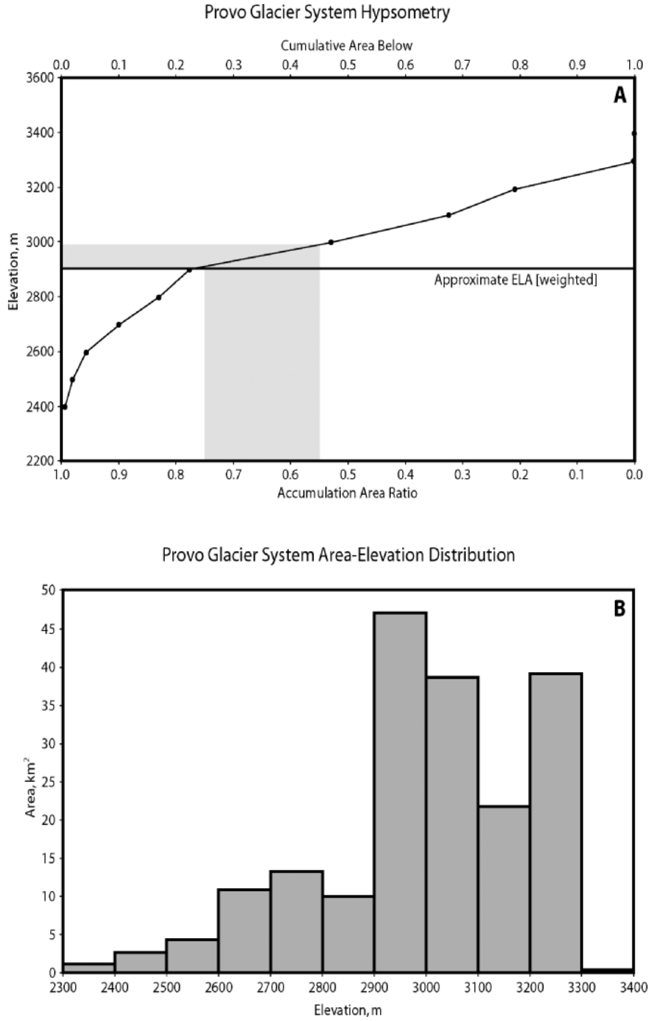

FIGURE 5. The distribution of ice in the upper Provo River drainage during the LGM. (A) The cumulative ice surface area at 100-m elevation intervals (upper *x*-axis) and accumulation area ratios (lower *x*-axis). The approximate weighted ELA for the entire Provo glacier system is indicated by the solid black horizontal line. The shaded gray boxes show the possible ELA range based only on accumulation area ratios spanning the range of 0.55–0.75. Due to the hypsometry, this range of AARs constrains possible ELAs to less than 100 m. (B) The surface area of ice between 100-m-elevation intervals. Note that approximately half of the area of the glacier system falls between 2900 and 3100 m a.s.l.

3360 m a.s.l. based on evidence of ice in the narrow col on the northwest side of Bald Mountain, which is consistent with the observation of minor breaks in slopes on the flanks of peaks to the west noted above.

PALEO-EQUILIBRIUM LINE ALTITUDES AND HYPSOMETRY

The reconstruction of paleo-ELAs is affected by some subjectivity since appropriate AAR and THAR values must be chosen. Though probable ranges of such values have been determined for different mountain glacier settings (Meier and Post, 1962; Porter, 1975; Meierding, 1982), the appropriate local ratios, as discussed above, will vary depending on glacier hypsometries. Attempts have been made to account for hypsometry using more rigorous techniques involving a balance ratio or geometric index (e.g., Furbish and Andrews, 1984; Benn and Lehmkuhl, 2000), but these methods require a very detailed reconstruction of ice surface topography for the entire glacier which was not feasible in this study for reasons previously discussed. However, characteristics of the hypsometries of the former glaciers in the Provo River drainage allow ELAs to be tightly constrained.

Reconstructed weighted ELAs for the North Fork Provo and Main Fork Provo Glaciers are 2890 and 2870 m a.s.l., respectively (Table 1). An AAR value of 0.75, which we used, is frequently applied to glaciers flowing from ice fields (Meier and Post, 1962; Leonard, 1984). An AAR value of 0.65 is generally used for reconstructed alpine valley glaciers (e.g., Porter, 1975; Leonard, 1984; Murray and Locke, 1989; Brugger, 2006) and has been previously applied to paleo-glaciers in the Uinta Mountains (Shakun, 2003; Munroe et al., 2006). When an AAR of 0.65 is used in ELA calculations for the glaciers in the Provo River drainage, ELAs only increase by approximately 20 m (Fig. 5A). Reducing the THAR from 0.5 to 0.4, the value used by Oviatt (1994), Shakun (2003), and Munroe et al. (2006), lowers ELAs by less than 20 m. The insensitivity of the ELAs of glaciers in the Provo River drainage is due to their hypsometries. Twenty-five percent of the area of this glacier system is between 2900 and 3000 m a.s.l., resulting in a gently sloping hypsometric curve over this elevation interval (Figs. 5A, 5B). As a result, changes in the AAR within this interval do not have a strong effect on the ELA, and accordingly, we estimate that the ELAs reconstructed here for the North Fork Provo and Main Fork Provo Glaciers have an uncertainty of ± 30 m.

Previous studies have conclusively shown that reconstructed ELAs in the Uinta Mountains progressively decrease toward the western end of the range, indicating a possible enhancement in precipitation due to pluvial Lake Bonneville, which covered much of western Utah during and following the LGM (Fig. 1; Munroe and Mickelson, 2002; Munroe et al., 2006). At that time, westerly circulation patterns carried weather systems across Lake Bonneville and over the Uinta Mountains as they do today (Mitchell, 1976). Moisture in these systems was likely augmented by evaporation from the large surface area of the lake, likely resulting in enhanced precipitation in the Wasatch Range and western Uinta Mountains and lower ELAs for glaciers more proximal to the lake (McCoy and Williams, 1985). ELAs were generally between 3000 and 3200 m a.s.l. for glaciers in the central and eastern Uinta Mountains, and ELAs for the glaciers in the western end of the range were between 2532 and 2800 m a.s.l. (Munroe et al., 2006). The ELAs for the glaciers in the Provo River drainage correspond quite well with the ELAs of the Smith and Morehouse (2837 m a.s.l.) and Main Weber (2870) glaciers, which flowed north from the Western Uinta Ice Field (Munroe et al., 2006).

As discussed above, the Provo glacier system, including the North and Main Fork Provo and Soapstone Glaciers, had 25% of its total area between 2900 and 3000 m a.s.l. and another 25% between 3000 and 3100. If ELAs rose ~100 m from 2910 m a.s.l. to approximately 3000 m a.s.l. (using the AAR method with a ratio of 0.75), the size of the accumulation area would have decreased by 32% due to the broad upland surfaces in the accumulation area; if ELAs rose 200 m, the accumulation area would have decreased by 52%. The majority of the area between 2900 and 3100 m a.s.l. is in the area that supplies ice to the Main Fork Provo Glacier, so the effect of such a reduction in accumulation area would be far greater for the Main Fork Provo Glacier than for the North Fork Provo Glacier. Only about 10%

of the glacier system was between 3100 and 3200 m a.s.l., so a continued increase in the ELA would have a less pronounced effect on the ice marginal positions.

Valley hypsometries in the Provo River drainage imply that a rise in ELAs would have been accompanied by differential retreat rates of individual outlet glaciers; this is supported by geomorphic evidence of rapid deglaciation in the Main Fork Provo Valley and slower deglaciation in the North Fork Provo Valley. Approximately 12 km upvalley from the Main Fork Provo terminal moraine is a 4- to 8-m-tall moraine on the east side of Alexander Lake, here referred to as the Alexander Moraine (Fig. 3). This feature is exceptionally sharp-crested, more bouldery than the moraines downvalley, and is continuous for 4 km. Such a large moraine was likely deposited by a steady-state glacier during overall deglaciation. In Figure 3, the dashed line shows a hypothetical reconstruction of the glacier system in the Provo River drainage with ELA of 3100 m a.s.l. and an AAR of 0.75. In this model, the Main Fork Provo Glacier would retreat 12 km upvalley, and moraine deposition along the western ice margin would likely occur at the approximate location of the Alexander Moraine. The North Fork Provo Glacier separates into two glaciers below the 3100 m a.s.l. ELA, and moraines in Boulder Creek Canyon help constrain potential ice margin positions. The North Fork Provo Glacier retreats ~4–5 km, and ice in the Boulder Creek Canyon only retreats a few kilometers because much of its accumulation area remains above 3100 m a.s.l. Thus, in response to a 200 m increase in the ELA, the Main Fork Provo Glacier retreats nearly three times farther upvalley than the North Fork Provo Glacier. The succession of large, well-preserved moraines deposited by the North Fork Provo Glacier is also indicative of relatively slow rates of retreat, whereas only 5–6 moraines, all less than 3 m tall, are preserved between the Main Fork Provo Glacier terminal moraine and the Alexander Moraine. However, immediately upvalley from the Alexander Moraine, the number of preserved moraines increases quickly, suggesting the rate of ice retreat may have slowed after the ELA rose above 3100 m a.s.l., though it is also possible this is an artifact of moraine preservation.

Conclusions

The preservation of geomorphic evidence of the most recent glaciation in the southwestern Uinta Mountains allows for detailed reconstruction of the ice extent during the LGM. The upper Provo River drainage was covered by an ice field, here named the Western Uinta Ice Field, during the Smiths Fork glaciation. This ice field fed distributary glaciers which flowed down the North Fork and Main Fork Provo Valleys, and the smaller Soapstone Glacier flowed from Soapstone Basin and joined the Main Fork Glacier. Reconstructed ELAs for the Main Fork and North Fork Provo Glaciers can be tightly constrained due to the hypsometry of the Western Uinta Ice Field; weighted ELAs of 2868 and 2886 m a.s.l., respectively, are in agreement with the westward-decreasing trend in ELAs previously documented in the Uinta Mountains (Munroe and Mickelson, 2002; Munroe et al., 2006). Geomorphic evidence also suggests that the glaciers in the upper Provo River drainage had very different deglaciation histories due to differences among the aerial distribution of the accumulation zones across broad upland surfaces within different drainages. The Main Fork Provo Glacier likely receded far more rapidly than the North Fork Provo Glacier initially after the onset of deglaciation.

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References Cited

Atwood, W. W., 1909: Glaciation of the Uinta and Wasatch Mountains. U.S. Geological Survey, *Professional Paper*, 61: 96.

Barnhardt, M. L., 1973: Late Quaternary geomorphology of the Bald Mountain area, Uinta Mountains, Utah. Master's thesis. University of Utah, Salt Lake City, 109 pp.

Benn, D. I., and Lehmkuhl, F., 2000: Mass balance and equilibrium-line altitudes of glaciers in high-mountain environments. *Quaternary International*, 65/66: 15–29.

Bradley, W. H., 1936: Geomorphology of the north flank of the Uinta Mountains: U.S. Geological Survey. *Professional Paper*, 185-I: 163–199.

Brugger, K. A., 2006: Late Pleistocene climate inferred from the reconstruction of the Taylor River Glacier Complex, southern Sawatch Range, Colorado. *Geomorphology*, 75: 318–329.

Bryant, B., 1992: Geologic and structure maps of the Salt Lake City 1° × 2° quadrangle, Utah and Wyoming. U.S. Geological Survey, *Miscellaneous Investigations Map*, I-1997: 1:125,000 scale.

Furbish, D. J., and Andrews, J. T., 1984: The use of hypsometry to indicate long-term stability and response of valley glaciers to changes in mass transfer. *Journal of Glaciology*, 30: 199–211.

Laabs, B. J. C., and Carson, E. C., 2005: Glacial geology of the southern Uinta Mountains. *In* Dehler, C. M., Pederson, J. L., Sprinkel, D. A., and Kowallis, B. J. (eds.), *Uinta Mountain Geology*. Salt Lake City, Utah: Utah Geological Association, publication, 33: 235–253.

Leonard, E. M., 1984: Late Pleistocene equilibrium-line altitudes and modern snow accumulation patterns, San Juan Mountains, Colorado, U.S.A. *Arctic and Alpine Research*, 16: 65–76.

McCoy, W. D., and Williams, L. D., 1985: Applications of an energy-balance model to the late-Pleistocene Little Cottonwood Canyon glacier with implications regarding the paleohydrology of Lake Bonneville. *In* Kay, P. A., and Diaz, H. F. (eds.), *Problems of and prospects for predicting Great Salt Lake levels*. Salt Lake City, Utah: University of Utah, 40–53.

Meierding, T. C., 1982: Late Pleistocene glacial equilibrium-line altitudes in the Colorado Front Range: a comparison of methods. *Quaternary Research*, 18: 289–310.

Meier, M. F., and Post, A. S., 1962: Recent variations in mass net budgets of glaciers in western North America. *International Association of Hydrological Sciences Publication*, 58: 63–77.

Mitchell, V. L., 1976: The regionalization of climate in the western United States. *Journal of Applied Meteorology*, 15: 920–927.

Munroe, J. S., 2005: Glacial geology of the northern Uinta Mountains. *In* Dehler, C. M. (ed.), *Uinta Mountain Geology*. Salt Lake City, Utah: Utah Geological Association, publication, 33: 215–234.

Munroe, J. S., and Mickelson, D. M., 2002: Last Glacial Maximum equilibrium-line altitudes and paleoclimate, northern Uinta Mountains, Utah, U.S.A. *Journal of Glaciology*, 48: 257–266.

Munroe, J. S., Laabs, B. J. C., Shakun, J. D., Singer, B. S., Mickelson, D. M., Refsnider, K. A., and Caffee, M. W., 2006: Latest Pleistocene advance of alpine glaciers in the southwestern Uinta Mountains, northeastern Utah, USA: evidence for the influence of local moisture sources. *Geology*, 34: 841–844.

Murray, D. R., and Locke, W. W. III, 1989: Dynamics of the late Pleistocene Big Timber Glacier, Crazy Mountains, Montana, U.S.A. *Journal of Glaciology*, 35: 183–190.

Oviatt, C. G., 1994: Quaternary geologic map of the upper Weber River drainage basin, Summit County, Utah. Utah Geological Survey, *Map*, 156.

Péwé, T. L., and Reger, R. D., 1972: Modern and Wisconsinan snowlines in Alaska. *Proceedings of the 24th International Geologic Congress*, Section 12: 187–197.

Porter, S. C., 1975: Equilibrium-line altitudes of late Quaternary glaciers in the Southern Alps, New Zealand. *Quaternary Research*, 5: 27–47.

Refsnider, K. A., 2006: *Late Quaternary glacial and climate history of the Provo River drainage, Uinta Mountains, Utah*. M.S. thesis. University of Wisconsin–Madison, 121 pp.

Shakun, J., 2003: *Last Glacial Maximum equilibrium-line altitudes and paleoclimate, northeastern Utah*. Senior thesis. Middlebury College, Middlebury, Vermont, 55 pp.

Sullivan, J. T., Nelson, A. R., LaForge, R. C., Wood, C. K., and Hansen, R. A., 1988: Central Utah regional seismotectonic study for USBR dams in the Wasatch Mountains. U.S. Department of the Interior Bureau of Land Reclamation, *Seismotectonic Report*, 88-5: 269.

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