Glacial Geology of the Northern Uinta Mountains

Jeffrey S. Munroe*

ABSTRACT

Glacial deposits on the north slope of the Uinta Mountains were investigated through map and air photo interpretation combined with field mapping. Deposits representing the Smiths Fork (local Pinedale equivalent), Blacks Fork (Bull Lake equivalent) and pre-Blacks Fork Glaciations are present in the form of terminal and lateral moraines, ground moraine, and outwash valley trains. Nineteen separate valley glaciers covering ~940 km^2 were present in the northern Uinta Mountains during the peak of the Smiths Fork Glaciation. Ice cover was even more extensive (~1100 km^2) during the penultimate Blacks Fork Glaciation. Distinct patterns of glaciation are evident from east to west along the north slope. Smiths Fork-age deposits in the Ashley Creek/Carter Creek/Sheep Creek region at the eastern end of the north slope indicate extensive ice stagnation. In contrast, Smiths Fork-age glaciers at the western end did not stagnate, and retreated actively after formation of their terminal moraines. In the central part of the north slope, bedrock structure apparently exerted a strong control on glacier dynamics in the Burnt Fork/Beaver Creek/Henrys Fork region. Reconstructed glacier equilibrium line altitudes for the Smiths Fork Glaciation drop dramatically at the western end of the range, indicating that glaciers in the western Uintas received considerably more precipitation than those farther east, probably due to lake effect snow derived from pluvial Lake Bonneville. Many drainages also contain small cirque-floor moraines indicating glacier advances after the Smiths Fork Glaciation, likely during the latest Pleistocene/early Holocene. Two drainages also contain evidence for Neoglaciation in the late Holocene, but before the classic Little Ice Age.

INTRODUCTION

The glacial history of the Uinta Mountains has received relatively little attention compared with other ranges of the western U.S. At least seven Masters theses and doctoral dissertations were completed on aspects of glacial geology and geomorphology in the northern Uintas between 1960-1995 (Schoenfeld, 1969; Barnhardt, 1973; Grogger, 1974; Gilmer, 1986; Schlenker, 1988; 1995; Zimmer, 1996), but only one paper was published (Grogger, 1975). Thus, until recently, two seminal U.S. Geological Survey reports from the early 1900's (Atwood, 1909; Bradley, 1936) comprised the main body of published literature on the range. That situation changed in the mid-1990s when a new group of researchers began a series of systematic investigations of glacial and fluvial geomorphology across the Uintas. The first major component of this effort was a detailed three-year (1998-2001) field study of the glacial geomorphology of the north slope of the Uintas, from the Carter Creek area at the east end, to the Mirror Lake Highway. This paper presents an overview of that work, including the distribution of glacial deposits representing the most recent and penultimate glaciations in the Uintas, interpretations of paleoclimate conditions during the last glacial maximum (ca. 20 ka BP), and evidence for Neoglaciation. The results of studies of the glacial history of the southern Uintas, and of the subalpine fluvial system on the north slope, are presented elsewhere in this volume(see Laabs and Carson, this volume; Carson, this volume).

Geology Department, Middlebury College, Middlebury, VT 05753
jmunroe@middlebury.edu

Munroe, J.S., 2005, Glacial geology of the northern Uinta Mountains, in Dehler, C.M., Pederson, J.L., Sprinkel, D.A., and Kowallis, B.J., editors, Uinta Mountain geology: Utah Geological Association Publication 33, p. 215-234.

History of Glacial Investigations

W.W. Atwood of the U.S. Geological Survey undertook the first study of the glacial record in the Uintas, producing a map and a report detailing the distribution of glacial deposits on the north and south slopes of the range (Atwood, 1909). Atwood divided end moraines, ground moraine and outwash deposits into two groups representing an older and younger glaciation. He also noted evidence for an earlier glaciation in some valleys. Despite difficulties including a primitive topographic base and a lack of aerial photos, Atwood's final map was a substantial contribution to our understanding of the Uinta Mountains and remained the most comprehensive investigation of the glacial record in the Uintas for almost a century.

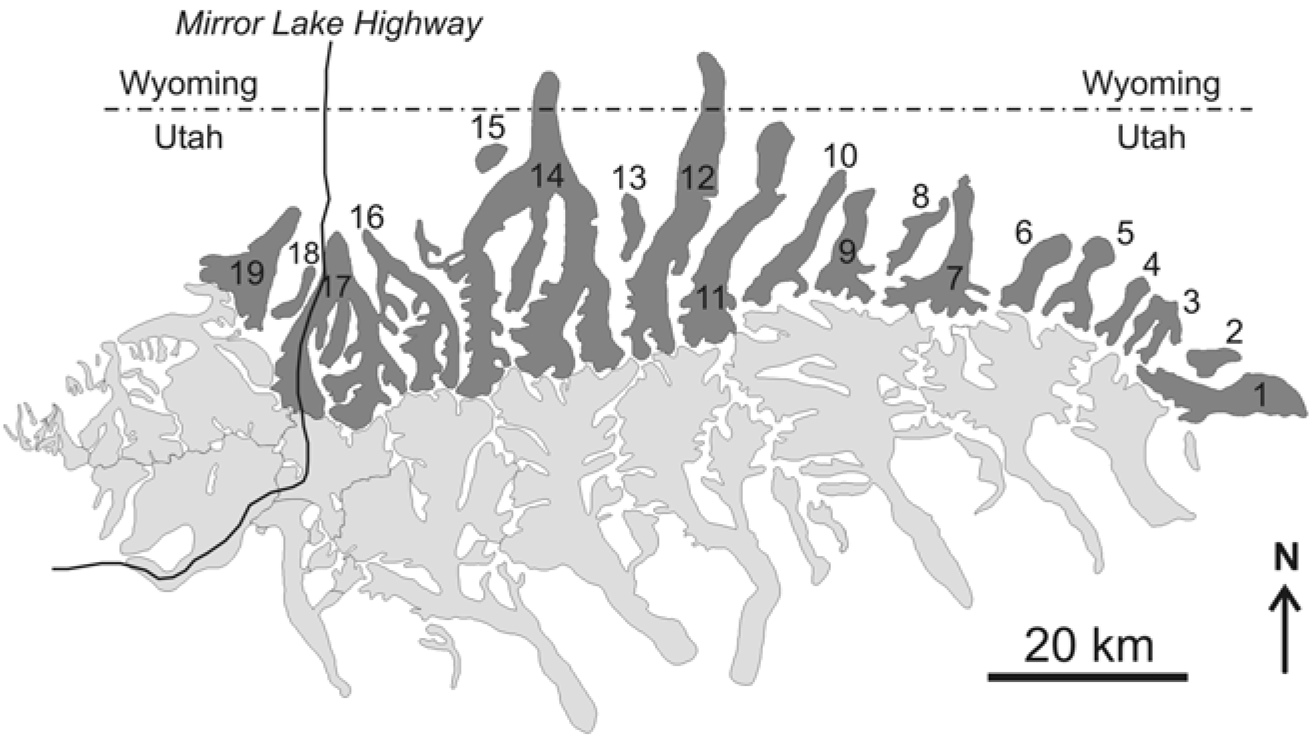
Most subsequent mapping of glacial deposits in the Uintas focused on the north slope of the range. Bradley (1936) recognized three separate ice advances in his report on the geomorphology of the north slope. From oldest to youngest, he named these the Little Dry, Blacks Fork, and Smiths Fork Glaciations, after localities where particular deposits are well preserved. Later, in his summary report on glaciation of the Rocky Mountains, Richmond (1965) correlated the Little Dry deposits with the Illinoian and pre-Illinoian glacial stages of the U. S. Midwest, the Blacks Fork with the early Wisconsin, and the Smiths Fork with the late Wisconsin Glaciation. Richmond also correlated the Blacks Fork advance with the Bull Lake Glaciation in the Wind River Range, and the Smiths Fork with the Pinedale. More recent work has also considered the Blacks Fork to be correlative with the Bull Lake Glaciation (Richmond, 1986; Bryant, 1992), which has been shown to pre-date the early Wisconsin (Sharp and others, 2003).

Several unpublished graduate theses have also reported aspects of the glacial geology of the northern Uintas. Schoenfeld (1969) investigated the Quaternary geology of the Burnt Fork drainage, part of the area studied by Bradley (1936). Barnhardt (1973) reported on the glacial and periglacial geomorphology of the Bald Mountain area along the Mirror Lake Highway. He employed both lichenometry and radiocarbon dating to develop a chronology for the late stages of the Pinedale Glaciation and local manifestations of the Neoglaciation. Grogger (1974) divided the Blacks Fork Glaciation into two stades, the Smith Fork into four stades, and the Neoglaciation into four stades on the basis of till weathering, vegetation, soil development, stratigraphic relations, and erosional modification. Gilmer (1986) focused on aspects of the geomorphology of a section of the north slope just north of the Utah-Wyoming border. Schlenker (1988) studied the geomorphology of the Blacks Fork area and corroborated the mapping of Atwood (1909). Zimmer (1996) revisited the glacial geology of the Smiths Fork drainage in his study of soil development on glacial landforms. Finally, Douglass (2000) investigated the complex of end moraines located where the West Fork Beaver Creek passes through the Madison Limestone hogback south of Lonetree, Wyoming. On the basis of geomorphic and soils evidence he was able to subdivide the Smiths Fork-age glacial deposits into an early and late advance.

Scope and Methods of This Study

Glacial deposits across the entire north slope of the Uinta Mountains were investigated through field mapping between 1998 and 2000 (Munroe, 2001). This work was part of a geomorphic mapping initiative undertaken by the U.S. Forest Service as part of the land systems inventory for the Wasatch-Cache National Forest. Lateral moraines, end moraines, ice-marginal drainages, and heads of outwash were identified and mapped at 1:24,000 scale on parts of twenty 7.5-minute quadrangles. Glacial limits were correlated between adjacent valleys on the basis of position, elevation, crosscutting relationships of outwash deposits confined to valley floors (valley trains), and the extent of weathering. All field mapping and air photo interpretations were compiled in a GIS, and Smiths Fork-age glaciers were reconstructed from the distribution of ice-marginal features (figure 1). A map illustrating the surficial geology of the glaciated valleys of the northern Uintas appears as a plate in the digital archive of this volume (Munroe-Plate in CD).

A variety of metrics were calculated for the reconstructed glaciers (table 1), including length, area, thickness (a minimum estimate from the difference in elevation between lateral moraines and



Smiths Fork-age Glacier	Number
South Fork Ashley Creek	1
North Fork Ashley Creek	2
East Fork Carter Creek	3
West Fork Carter Creek	4
South Fork Sheep Creek	5
Middle Fork Sheep Creek	6
Burnt Fork	7
Thompson Creek	8
Middle Fork Beaver Creek	9
West Fork Beaver Creek	10
Henrys Fork	11
Smiths Fork	12
West Fork Smiths	13
Blacks Fork	14
Little West Fork Blacks	15
East Fork Bear	16
Hayden Fork Bear River	17
Gold Hill	18
West Fork Bear	19

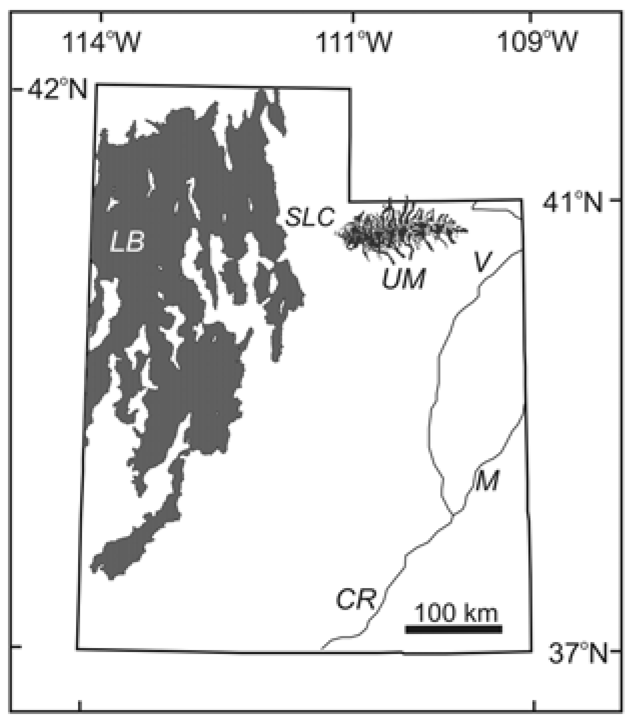


Figure 1. *Reconstructed Smiths Fork-age glaciers of the Uinta Mountains. North slope glaciers considered in this report (from Munroe, 2001) are shown in darker gray. Glaciers shaded in lighter gray are from Shakun (2003)--south slope, and Oviatt (1994)--west of the Mirror Lake Highway. North slope glacier numbers are keyed to the list below. The Utah-Wyoming stateline and the Mirror Lake Highway are shown for reference. Inset shows the Smiths Fork-age glaciers of the Uinta Mountains (UM) in relation to the state of Utah and the extent of the Lake Bonneville highstand (LB, from the Utah Automated Geographic Reference Center, http://agrc.its.state.ut.us/) ca. 15 ka BP. SLC is Salt Lake City; V is Vernal; M is Moab, and CR is the Colorado/Green River system.*

Table 1. *Parameters of Reconstructed Smiths Fork-age Glaciers*

Glacier	Glacier*	Area km^2	Length km	Mean Thickness m	Elevation High m	Elevation Low m	Elevation Drop m	Mean Gradient m km-1	Perimeter km	Complexity **	Driving Stress kPa ***	ELA m #
South Fork Ashley Creek^	1	51.4	18.4	83	3460	2760	700	38.0	45.0	1.77	29	3115
North Fork Ashley Creek	2	9.9	5.5	32	3340	3000	340	62.0	14.0	1.26	14	3142
East Fork Carter Creek^	3	19.1	7.1	71	3420	2800	620	87.9	32.9	2.12	45	3084
West Fork Carter Creek^	4	12.4	7.8	70	3440	2820	620	79.8	20.0	1.61	50	3130
South Fork Sheep Creek^	5	24.5	10.2	38	3500	2780	720	70.7	32.3	1.84	24	3135
Middle Fork Sheep Creek^	6	23.4	9.7	68	3500	2880	620	63.9	24.0	1.40	32	3129
Burnt Fork	7	46.5	17.4	142	3600	2680	920	53.0	56.8	2.35	69	3176
Thompson Creek	8	13.9	10.1	105	3480	2820	660	65.5	23.9	1.81	55	3136
Middle Fork Beaver Creek	9	33.6	11.8	125	3660	2840	820	69.5	37.7	1.83	72	3231
West Fork Beaver Creek	10	34.4	16.9	118	3780	2740	1040	61.6	42.3	2.03	56	3227
Henry's Fork	11	76.4	24.2	151	3760	2680	1080	44.6	64.9	2.09	36	3187
Smiths Fork	12	105.1	32.3	117	3740	2640	1100	34.1	87.1	2.40	28	3114
West Fork Smiths	13	10.2	7.1	64	3400	2960	440	61.6	16.1	1.42	39	3142
Blacks Fork	14	243.3	36.3	141	3660	2620	1040	28.7	221.6	4.01	31	3089
Little West Fork Blacks	15	6.0	7.1	61	3060	2880	180	25.2	9.4	1.09	29	2976
East Fork Bear	16	56.7	19.6	83	3540	2620	920	46.9	88.8	3.33	32	3075
Hayden Fork Bear River	17	118.7	21.3	128	3500	2580	920	43.3	126.1	3.27	38	3059
Gold Hill^^	18	7.4	6.9	--	3050	2780	270	39.2	15.3	1.59	--	2905
West Fork Bear	19	47.9	12.8	98	3100	2580	520	40.5	40.1	1.63	40	2861
Mean	--	49.5	14.9	94.2	3473	2761	712	53.5	52.5	2.04	39.9	3101
Standard Deviation	--	56.9	8.8	36.2	218	123	279	17.5	51.0	0.76	15.4	97

* Glacier number keyed to Figure 1
** Ratio of glacier perimeter to that of a circle with the same area.
*** Average value computed at ~60-m vertical increments along lateral moraines, using thickness and gradient values from this table, an ice density of 910 kg/m3, and a shape factor F after Ackerly, 1989.
\# Weighted average of Accumulation Area Ratio (0.65, 0.50 for piedmont glaciers^), Toe-Headwall Altitude Ratio (0.40), uppermost elevation of lateral moraines, and elevation of the lowest northeast-facing cirque floor (see Munroe and Mickelson, 2002).
^^ Poor preservation of lateral moraines precluded estimation of former ice thickness and basal driving stress for the Gold Hill glacier.

valley floors), highest and lowest elevation (from inflection point on cirque headwall, and terminal moraine altitude, respectively), gradient (from high-low elevation difference and glacier length), perimeter, and complexity (ratio of glacier perimeter to that of a circle with the same area). Basal driving stress (in kPa) was calculated for each reconstructed glacier using lateral moraine slopes as a proxy for former glacier surface slope, estimated ice thickness, an ice density of 910 kg/m^3, and a shape factor dependent on valley width (after Ackerly, 1989).

To investigate paleoclimate conditions in the northern Uintas during the Smiths Fork Glaciation, equilibrium-line altitudes (ELAs) were calculated for each of the reconstructed glaciers using four separate methods: Accumulation Area Ratio (AAR), Toe-Headwall Altitude Ratio (THAR), uppermost elevation of lateral moraines (LM), and cirque floor altitudes (CIR). Briefly, the AAR method is based on the observation that the accumulation area comprises approximately 65% of the surface area of modern glaciers in equilibrium. Similarly, equilibrium-lines are usually located at an elevation above the terminus that is equivalent to ~40% of the total elevation difference between the terminus and top of the glacier, providing a basis for the THAR method. Because the transition from net erosion to net deposition occurs at the equilibrium line of glacier in mass balance, lateral moraines should extend only from the terminal moraine to the equilibrium line. Thus, the uppermost elevation of continuous lateral moraines in a deglaciated valley approximates the equilibrium line altitude during construction of the terminal moraine. Finally, the floor elevations of cirques with similar aspects provide an additional constraint on the elevation of the former equilibrium line. Estimates obtained from these four methods were combined in a weighted average, with the greatest weight placed on the AAR estimate and the least on the CIR method, reflecting the estimated accuracy of these methods and the success other researchers have had in their application (e.g. Meierding, 1982; Locke, 1990). The resulting ELA estimates were compared with the modern climate of the Uintas, as recorded by automated weather stations and Snowpack Telemetry (SNOTEL) sites. Additional details of this process are presented in Munroe and Mickelson (2002).

OVERVIEW OF GLACIAL CHRONOLOGY AND GEOMORPHOLOGY

Moraines of the Smiths Fork Glaciation

Bradley (1936) named the Last Glacial Maximum (LGM) in the Uintas the Smiths Fork Glaciation for the well-expressed terminal moraine along the East Fork Smiths Fork just inside the Wasatch-Cache National Forest boundary, in sections 4 and 5, T. 12 N., R. 115 W. (Buck Fever Ridge quadrangle). Moraines considered correlative with the Smiths Fork type locality are present in each of the major glaciated valleys of the northern Uintas (Munroe-Plate). The Smiths Fork moraines are presumably coeval with the Pinedale deposits of the Wind River Mountains, which have been dated by cosmogenic ^{26}Al and ^{10}Be to between 23 and 16 ka BP (Gosse and others, 1995). Preliminary cosmogenic surface exposure ages from the southern Uintas support this correlation (Laabs and others, 2003). Smiths Fork-age moraines in the northern Uintas commonly feature steep frontal slopes, narrow crests, and an overall unweathered appearance, and are located at elevations between 2580 and 3000 m (figure 2). Outwash valley trains are often graded to the frontal slopes of these terminal moraines, and some of the larger recessional moraines are also associated with outwash terraces.

Several types of diamicton are commonly found in the end moraines, representing a variety of sedimentary mechanisms and environments. Massive boulders, some partially buried by silt, are often found at the moraine surface. These boulders were likely deposited directly from a supraglacial position during wasting of the underlying ice, or by direct rolling and sliding to the glacier terminus. The capping silt layer is interpreted as loess, given its ubiquity, uniform grain size, and fairly constant thickness. An unstratified sandy till is present below the silt cap and boulder mantle in all exposures. Fabric is rarely well developed in this sediment, mainly due to the lack of elongate clasts. The sandy till is interpreted as meltout till deposited at the ice margins. Exposures along the north side of the Little West Fork Blacks Fork contain mudflow diamicton deposited by saturated debris flows off the glacier surface (Schlenker, 1988). Basal till is rarely exposed at the surface in the northern

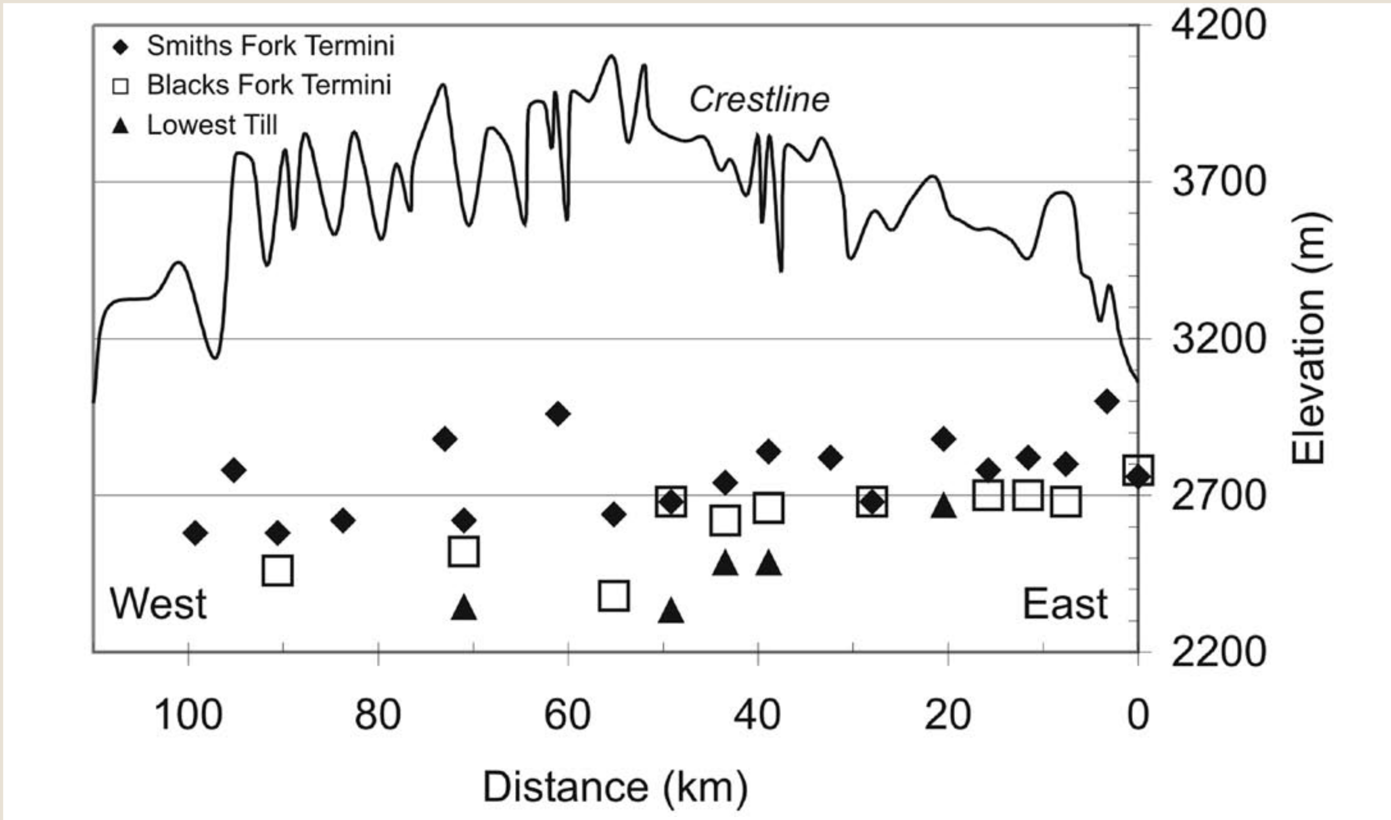


Figure 2. *Elevations of terminal moraines across the north slope and east-west profile of the Uinta ridgecrest. The mean elevation of the 19 Smiths Fork-age moraines is 2760 m, with a standard deviation (σ) of 123 m. The mean of 11 Blacks Fork-age moraines is 2620 m (σ 120 m). Pre-Blacks Fork-age till is found in five valleys, at elevations down to 2330 m, with a mean of 2460 m. Note that in some valleys the Smiths Fork and Blacks Fork-age moraines are found at the same elevation.*

Uintas, although diamicton containing striated bullet clasts exposed by landsliding along the Little West Fork Blacks Fork may represent lodgement till. Overconsolidated till with well-developed columnar jointing is also exposed along the West Fork Beaver Creek upstream from the Smiths Fork-age terminal moraine complex. Although striated clasts were not observed in this diamicton, it is likely also basal till given its extreme compaction.

Uinta Mountain quartzite is the dominant clast lithology in the end moraines, even in locations downslope from outcrops of Paleozoic rocks. Douglass (2000) found no limestone clasts in his study of the end moraine complex along the West Fork Beaver Creek, despite being within 2 km of Madison Limestone exposures. In some valleys (i.e., Mill Creek, distributary from the West Fork Blacks Fork glacier), till is finer-grained reflecting a greater component of the matrix being derived from the Wasatch Formation. However, the dominance of quartzite clasts in the Oligocene Bishop Conglomerate, which locally caps the Wasatch Formation, ensures that these tills resemble those derived entirely from the Precambrian rocks. Clasts of diorite from a Cambrian dike crossing the eastern High Uintas (Ritzma, 1983) are occasionally found on the surface of end moraines at the eastern end of the north slope. Although rare, these clasts are easily recognizable by their weathered character.

Soils on the Smiths Fork-age moraines are poorly developed Typic Cryochrepts with weak E horizons and slightly iron-enriched Bw horizons (Bockheim and others, 2000; Douglass, 2000). The sandy B horizons of these soils are quite well drained, and features indicative of reducing conditions are absent. The A and sometimes E horizons in these profiles are developed in the capping loess layer, giving them physical and chemical properties that contrast strongly with the lower solum (Bockheim and Koerner, 1997; Bockheim and others, 2000).

Older Moraines and Till Sheets

Bradley (1936) named the penultimate glaciation in the northern Uintas the Blacks Fork Glaciation after the older moraines and outwash terraces along the Blacks Fork River, southwest of Robertson, Wyoming. The Blacks Fork stage in the Uintas is considered correlative with the Bull Lake Glaciation in the Wind River Mountains, stades of which have been dated to between 100 and 130 ka BP by cosmogenic surface-exposure dating using ^{36}Cl and ^{10}Be (Phillips and others, 1997; Chadwick and others, 1997). Blacks Fork moraines are more weathered than the younger Smiths Fork-age landforms, and are found at lower elevations (2380 to 2780 m) (figure 2; Munroe-plate in CD).

Till sheets of possible pre-Blacks Fork age were also mapped in five of the north slope valleys. The most extensive of these features is found along the Henrys Fork southwest of Lonetree, Wyoming, where older till is present down to an elevation of 2330 m (figure 2; Munroe-plate in CD). Pre-Blacks Fork-age till forms subtle landforms typified by erratic surface boulders and subdued kettles. Along the South Fork Sheep Creek, the older till sheets can be recognized by the subrounded character of the surface clasts in contrast to the flaggy clasts characteristic of locally derived regolith. In the western Uintas, large quartzite erratics are the most compelling argument for deposits of pre-Blacks Fork age.

Glaciofluvial Deposits

Outwash surfaces graded to the Smiths Fork end moraines are present in all the north slope valleys (Munroe-Plate). Massive outwash valley trains graded to moraines of Blacks Fork age are also present farther downvalley. South of Lonetree, Wyoming, outwash from Henrys Fork combined with that from the West and Middle Forks of Beaver Creek to form a compound outwash surface more than 5 km wide (Munroe-plate in CD).

Glaciofluvial sediments are composed of rounded, sorted quartzite cobbles and pebbles, with small boulders common near moraine fronts. Clasts within 1-2 m of the surface frequently have thin (up to ~1 cm) carbonate coatings. Outwash terraces are sparsely vegetated due to the excessively well-drained soils, providing a strong contrast with the moraines, which are usually forested. Relict braided channel patterns are locally visible from the air, most notably on the outwash surface along the Bear River, south of Evanston, Wyoming.

Valley trains graded to recessional moraines are also present in some of the north slope valleys, although it can be difficult to distinguish these younger glaciofluvial deposits from modern alluvial plains. Outwash is also present in some drainages that carried meltwater away from the glacier margins. These drainages now contain underfit streams incapable of transporting the coarser sediment originally deposited by meltwater-enhanced flows (figure 3, Munroe-Plate).

Figure 3. *Mill Creek passing through a wide channel cut beside the Smiths Fork-age end moraine upstream from the junction of Forest Roads 058 and 061. Note the coarse lag of quartzite boulders. Some of these are derived from the lateral moraine of the Mill Creek glacier immediately left of the stream. Most of the boulders, however, were likely transported to this position by meltwater-enhanced flows that formed this channel during the Smiths Fork Glaciation.*

Scoured Bedrock Surfaces

Striated bedrock ledges and areas of thin till over bedrock are common in the higher cirque basins, especially at the western end of the Uintas (Munroe-plate in CD). These surfaces are often littered with a lag of scattered erratic cobbles and boulders, and striations are best preserved around the periphery of the exposed ledges where till has been most recently removed (figure 4). In most of the upper cirque basins, measurements of striations reveals a pattern of flow converging toward the valley axis before turning and

Figure 4. *Scoured and striated ledges of bedrock in the western Uintas south of McPheters Lake near the head of the Stillwater Fork of Bear River. Erratic blocks of quartzite likely represent a lag from an initial thin covering of till.*

continuing parallel with the deepest part of the glacial trough. Striated bedrock surfaces are more common in the western Uintas, possibly reflecting facies changes or variations in the degree of metamorphism within the Uinta Mountain Group bedrock along the range. Rocks at the western end of the Uintas appear to be more indurated and resistant to subaerial erosion, resulting in more extensive valley-bottom exposures of bedrock, and greater preservation of striations. Also, there appears to be more shale in cirque headwalls at the eastern end of the Uintas, contributing to a widespread mantle of fine colluvium in the upper cirque basins, reducing the area of exposed ledges.

Glacier Effects on Fluvial Systems

Repeated Pleistocene valley glaciations impacted drainages of the north slope in a variety of ways. The most obvious effect is the enlargement and deepening of drainages as they were transformed from V-shaped fluvial valleys into U-shaped glacial troughs. Hanging valleys, such as the Left Hand Fork of the East Fork Bear River and the Amethyst Basin of the Stillwater Fork Bear River, are elevated 60 m and more than 150 m respectively above their trunk valleys, providing a minimum estimate of the amount of glacial downcutting. Former ice-marginal drainages are also common components of the north slope landscape. Many of these channels now contain misfit streams (e.g., Dahlgreen Creek alongside the western lateral moraine of the Henrys Fork, see figure 10). This discrepancy is indicative of the large volumes of meltwater that flowed from the ice margin during the glacial maximum and subsequent deglaciation. Oversized channels cut into bedrock are present in some of these drainages, as are large gravel bars above the grade of modern streams. In other situations, lateral moraines redirected ice-marginal meltwater into other drainages, for instance upstream from The Narrows on the east side of the Henrys Fork. The massive right lateral moraine of the Henrys Fork glacier also contained a stagnant ice body that drained eastward

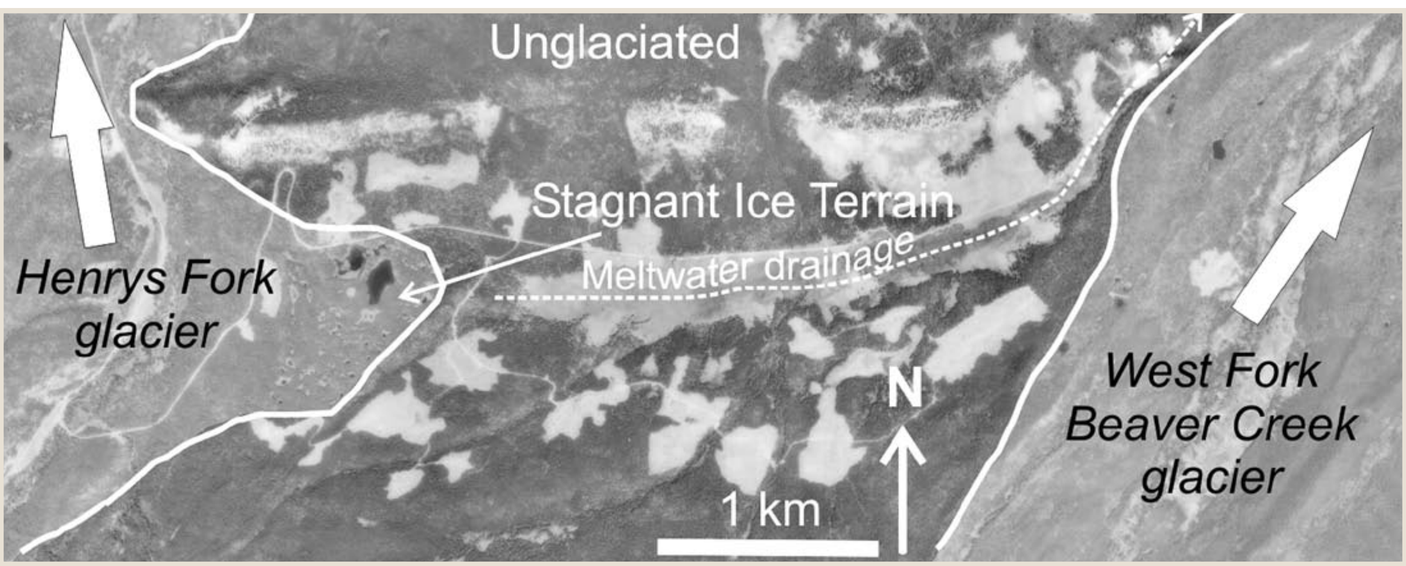


Figure 5. *Vertical aerial photo of the Bullocks Park area showing the Smiths Fork-age glaciers in the Henrys Fork (left) and West Fork Beaver Creek (right). Forest Road 058 is visible (thin gray line) ascending from the Henry's Fork through a tight switchback. The road passes just to the north of the stagnant ice terrain, marked by an abundance of kettles. Meltwater from the stagnant ice complex drained eastward (dashed arrow) into the West Fork Beaver Creek valley.*

into the West Fork Beaver Creek (figure 5).

These diversions are not necessarily permanent, as lateral moraines are vulnerable to post-glacial modification. Along the east side of the Henrys Fork south of figure 5, Joulious Creek, which follows a former ice-marginal drainage, has diverted from its eastward path into the West Fork Beaver Creek and now joins the Henrys Fork after a steep descent down the proximal slope of the lateral moraine. A similar relationship is seen near Meeks Cabin Reservoir where the Little West Fork Blacks Fork diverges from an ice-marginal channel (now occupied by Fish Creek) and merges with the Blacks Fork after a tumultuous descent through a steep canyon cut across Blacks Fork and Smiths Fork-age lateral moraines.

Finally, fluctuations of the glacier margins locally impounded ice-marginal drainages, forming temporary ice-marginal lakes. Dahlgreen Creek on the west side of the terminus of the Henrys Fork glacier (see figure 10) was dammed on the south side of Red Mountain, forming a lake in the NE¼ of section 27, T. 3 N., R 115 W. (Gilbert Peak NE quad). Other examples of ice-marginal lakes are present along the east side of the Burnt Fork in Lost Creek Sink (figure 6) and at Willow Park on the east side of the West Fork Beaver Creek.

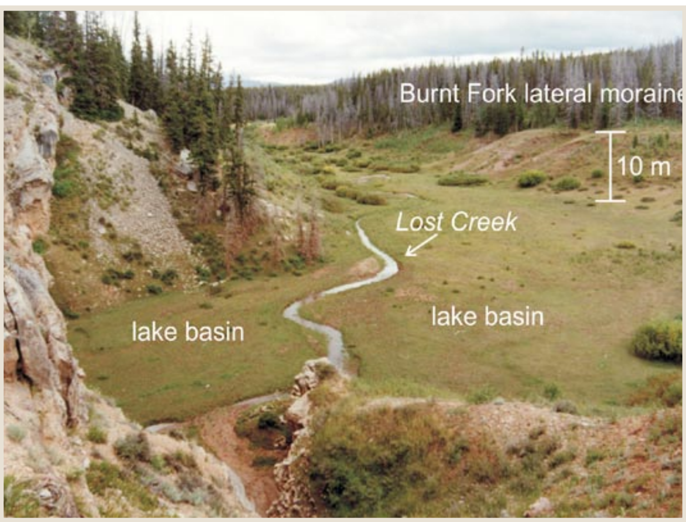


Figure 6. *View southward across Lost Creek Sink showing the lower part of the Madison Limestone hogback (left) and the distal slope of a lateral moraine of the Burnt Fork glacier. A lake existed in this location during the Smiths Fork Glaciation, filling the sink with sediment to the level of the base of the trees in the background. The modern sink is inset ~10 m into this fill, however the depth to bedrock within the sink is unknown.*

DYNAMICS OF THE SMITHS FORK GLACIATION

Study of the well-preserved terminal moraines from the Smiths Fork Glaciation in the northern Uintas provides information about the style of deglaciation and allows reconstruction of the paleoclimate responsible for the Smiths Fork advance. The geomorphic record supports division of the major north slope glaciers into distinct eastern, central, and western groupings. The eastern north slope glaciers include East and West Carter Creeks, South Fork Sheep Creek, and Middle Fork Sheep Creek. The South and North Forks of Ashley Creek are technically part of the south slope, but will briefly be considered in this report. The Burnt Fork, Thompson Creek, Middle Fork Beaver, West Fork Beaver Creek, and Henrys Fork represent the central north slope. Finally, the western north slope contains the East Fork Smiths Fork, West Fork Smiths Fork, Blacks Fork (including the Little East, East, Middle, and West Forks), Bear River (including the East Fork, Stillwater Fork, and Hayden Fork), and West Fork Bear River. Each of these three regions is considered separately in the discussion that follows.

Glaciers of the Eastern North Slope

The eastern glaciers were generally the smallest, ranging in area from 12 to 25 km^2 (mean of 20 km^2) and with lengths from 7.1 to 10 km (table 1; Munroe-Plate). Yet despite their short lengths, the four glaciers in Carter and Sheep Creeks all advanced beyond their deeply carved valleys to form unconfined terminal (piedmont) lobes. With steep frontal slopes rising 30 to more than 60 m above the outwash surfaces that front them, the moraine complexes formed by these glaciers are the most dramatic features of the eastern glacial valleys. These moraines are also extremely hummocky, indicating that their formation incorporated a large amount of buried ice (figure 7). Nowhere else in the northern Uintas did piedmont glaciers form during the Smiths Fork Glaciation, a situation that is likely due to the structural geology in this part of the range. Because the glacial valleys dip less steeply than the Uinta Mountain Group in this area, the valleys shallow progressively to the north, allowing the glaciers to

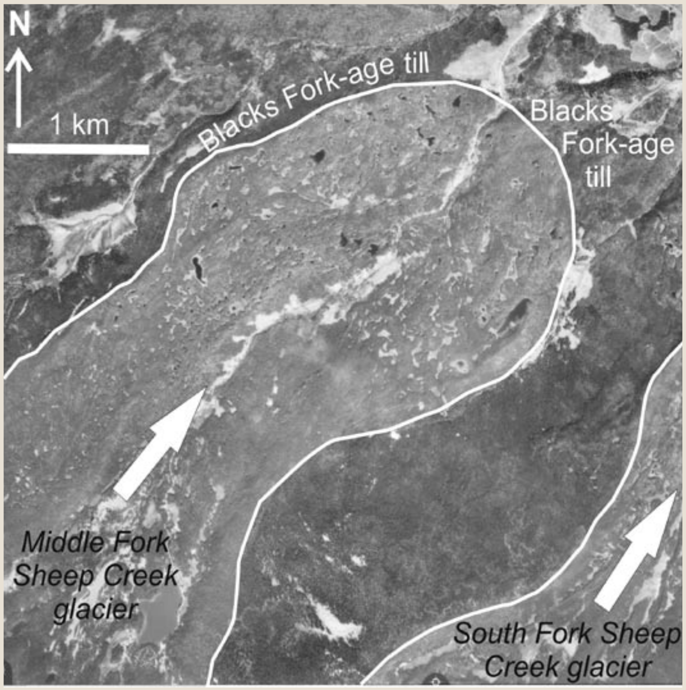


Figure 7. *Vertical air photo of the Middle Fork Sheep Creek showing the hummocky nature of the terminal moraines in this part of the northern Uintas. The piedmont form of the Middle Fork Sheep Creek glacier is clearly visible (glacier shaded in gray, moraine crest highlighted in white). Blacks Fork-age till is present beyond the termini of both glaciers, however it is only visible along the Middle Fork in this photograph.*

broaden into unconfined piedmont lobes on the north limb of the Uinta anticline. Furthermore, all four of these glaciers terminated upslope from the prominent hogback of Paleozoic sedimentary rocks that encircles the range, so they were not influenced by this topographic obstacle as were glaciers farther west.

It is also worth considering the South Fork Ashley Creek in this discussion; although it technically is part of the south slope, this drainage was included in this study of the north slope glaciers because of its proximity and geomorphic similarity. Located ~8 km southeast of the Carter Creek drainage, the South Fork Ashley Creek glacier advanced more than 18 km during the Smiths Fork Glaciation, forming a pair of lateral moraines unlike any others in the Uintas in terms of their size and surface form. Locally in excess of 1.5 km wide, each moraine rises more than 70 m above the valley bottom and is laterally continuous for 8-10 km (Munroe-Plate). The moraines are extremely hummocky, with internal relief in excess of 20 m. Four kilometers south of the Red Cloud Loop (Forest Road 018) crossing of the South Fork, the lateral moraines merge to form a terminal moraine that the stream breeches in a narrow, boulder-choked gorge. Outwash graded to this terminal moraine forms a sparsely vegetated valley train with a remnant braided-channel network that extends downstream for over 2 km. The lateral moraines are clear evidence for large-scale ice-marginal deposition of debris and stagnant ice, but there is relatively little evidence of ice stagnation along the valley axis. Instead, ice appears to have actively backwasted from the position of the maximum Smiths Fork advance, pausing long enough to form a recessional moraine and associated outwash system at Hicks Park.

An explanation for the genesis of these unusually hummocky moraines may be found in the bedrock geology of the Lakeshore Basin area at the head of the South Fork Ashley Creek drainage, where shale outcrops are particularly abundant. Munroe and Douglass (1998) suggested the presence of these weak shales facilitated lateral erosion of the Smiths Fork-age glacier resulting in large-scale mass wasting onto the glacial surface. This abundant supraglacial debris would have been incorporated into the lateral moraines, where its later ablation would have caused local surface collapse, forming the hummocky moraine surface.

Glaciers of the Central North Slope

The five glaciers in the central north slope were generally larger than those to the east, with areas ranging from 14 to 76 km^2 with a mean of 41 km^2 (table 1). They were also long enough (10.1 to 24.2 km) that they all reached the hogback of Paleozoic rocks (Munroe-plate in CD). The Middle Fork Beaver Creek and Thompson Creek both terminated on the upslope side of the hogback, the Burnt Fork and West Fork Beaver Creek both terminated in narrow gaps eroded through the hogback, and the Henrys Fork advanced 7 km north of the hogback. However, despite the difference in their terminal positions, the dynamics of these glaciers were all controlled by bedrock structure.

The Burnt Fork and Thompson Creek Glaciers had distinct accumulation areas, but both glaciers advanced toward the same gap in the limestone (figure 8). As they neared this gap, the larger Burnt Fork Glacier deflected the smaller Thompson Creek

Glacier, although meltwater from Thompson Creek was routed along the northwest side of the Burnt Fork Glacier. This meltwater apparently enhanced ablation of the Burnt Fork glacier, constraining the position of the glacier terminus. As a result, the Burnt Fork Glacier became thicker and steeper, generating a greater basal shear stress and increasing the flux of ice into the narrow gap. At the peak of the Smiths Fork Glaciation, the Burnt Fork Glacier had the greatest maximum thickness (210 m), and the second largest average basal driving stress (69 kPa) of glaciers in the northern Uintas (table 1). Regardless, ice was unable to advance beyond the Paleozoic hogback, as demonstrated by the location of the Burnt Fork terminal moraine between the double ridges of Madison limestone and Weber sandstone (figure 8). Meltwater produced along the east side of the Burnt Fork Glacier drained underground through Lost Creek Sink (see figure 6).

The two forks of Beaver Creek advanced toward wider gaps in the Paleozoic hogback. The West Fork Beaver Creek Glacier terminated just north of the hogback and just south of Blacks Fork-age and pre-Blacks Fork-age deposits (figure 9). The close spacing of end moraines from multiple glaciations indicates that this location marks the most likely glacier equilibrium position given the geometry of the upper West Fork Beaver Creek accumulation area and the range of late Quaternary glacial conditions. Field evidence, including soil development and weathering rind thickness, suggests that the Smiths Fork terminal moraine at this location may be a compound feature produced by two glacial advances separated by a fairly long weathering interval (Douglass, 2000).

The neighboring Middle Fork Beaver Creek Glacier appears to have been influenced more by structural geology than the West Fork. The Middle Fork terminal moraine is a conspicuous zone of hummocky terrain immediately south of Beaver

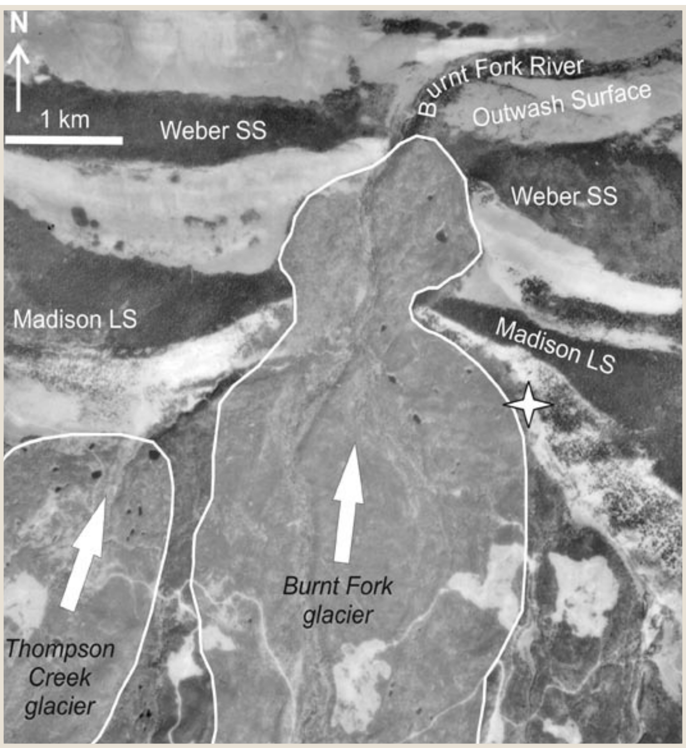


Figure 8. *Vertical air photo showing the Smiths Fork-age terminal moraine in the Burnt Fork valley located between hogbacks of the Paleozoic Madison limestone and Weber sandstone. Meltwater from the Thompson Creek glacier flowed alongside the western margin of the Burnt Fork glacier to reach the outwash surface north of the hogbacks. The star marks the location of the Lost Creek Sink (see figure 6).*

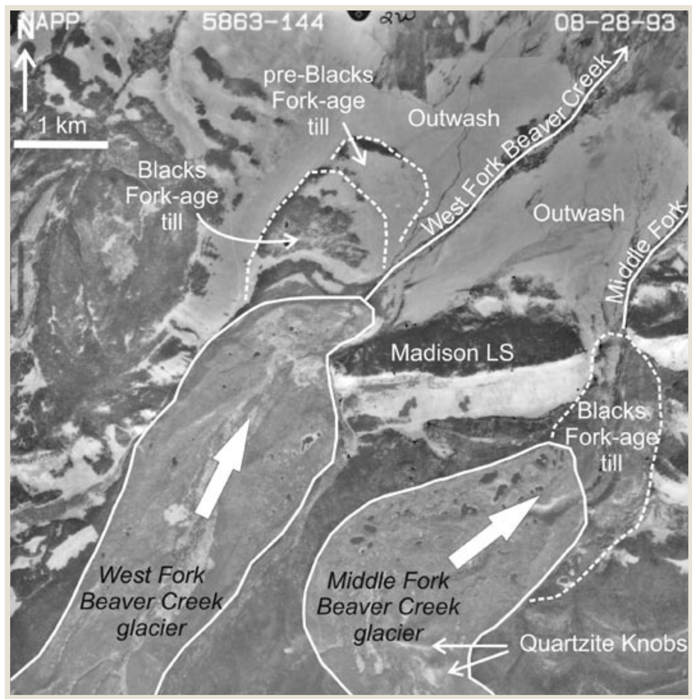


Figure 9. *Vertical air photo showing the reconstructed Smiths Fork-age glaciers of the West Fork and Middle Fork of Beaver Creek. Terminal moraines of Blacks Fork age (dashed) are located north of the Smith Fork-age termini, and a small area of pre-Blacks Fork-age till is present along the West Fork. Meltwater from both glacier systems contributed to the formation of an extensive compound outwash surface that covers much of the northern half of the photograph. Note the location of the prominent quartzite knobs considered responsible for stagnation of the terminus of the Middle Fork glacier.*

Mountain on the Hole in the Rock quadrangle (figure 9). The numerous large kettles (some with lakes up to 15 m deep) in this moraine indicate that the terminus of the Middle Fork Beaver Creek Glacier stagnated. The twin quartzite knobs highlighted on figure 9 were likely responsible as they are located near the valley center and their summit elevations are <20 m below the crests of the lateral moraines. In this situation, even a small amount of surface downwasting would dramatically reduce the flux of ice past the ridges into the terminal zone of the glacier. Thus, the terminus would have stagnated at the start of deglaciation, forming a terminal moraine with abundant buried ice blocks.

Finally, the Henrys Fork Glacier, presumably by virtue of its large, high-elevation accumulation area, was able pass beyond the hogback of Madison limestone at The Narrows. Quartzite erratics sit above striated limestone on both sides of The Narrows, and their elevations are concordant with the lateral moraines farther north, indicating that ice overtopped the limestone during the LGM. After passing The Narrows, the glacier formed a sprawling terminal moraine loop to the east of Red Mountain (figure 10). Once again, the complex surface of this moraine indicates that a large amount of ice was buried in the process of moraine building, and it appears that structural geology played a role. The hogback ridge, which rises ~250 m from the valley floor, reduced the width of the glacier from 2.3 km at 10,000 ft (just upstream from the hogback) to 1.4 km at 9900 ft. As a result, a zone of stagnant ice formed along the east side of the glacier on the up-ice side of the limestone. This ice was presumably separated from flowing ice in the valley by a shear margin, effectively reducing the active width of the glacier. Nonetheless, the glacier cross-section within the hogback (at 9900 ft) was ~25% smaller than that immediately upvalley (at 10,000 ft). Because the Henrys Fork Glacier was only ~200 m thick where it passed through The Narrows, a small amount of surface downwasting at the onset of deglaciation would have separated the ice in the terminal zone from the active ice in the valley to the south. Such an unforgiving geometry would have caused the terminal zone of the Henrys Fork glacier to stagnate at the start of deglaciation, even when glaciers in valleys farther west were able to keep pace with ameliorating conditions by actively retreating upvalley in response to shifting mass balance.

Interestingly, Atwood mapped this massive terminal moraine loop as a Blacks Fork-equivalent, and located the Smiths Fork-equivalent terminal moraine near Alligator Lake, 6 km upvalley from The Narrows. However, the overall fresh form of this massive moraine loop, its steep frontal slope, its elevation, and its dramatically kettled surface argue that it was formed during the Smiths Fork

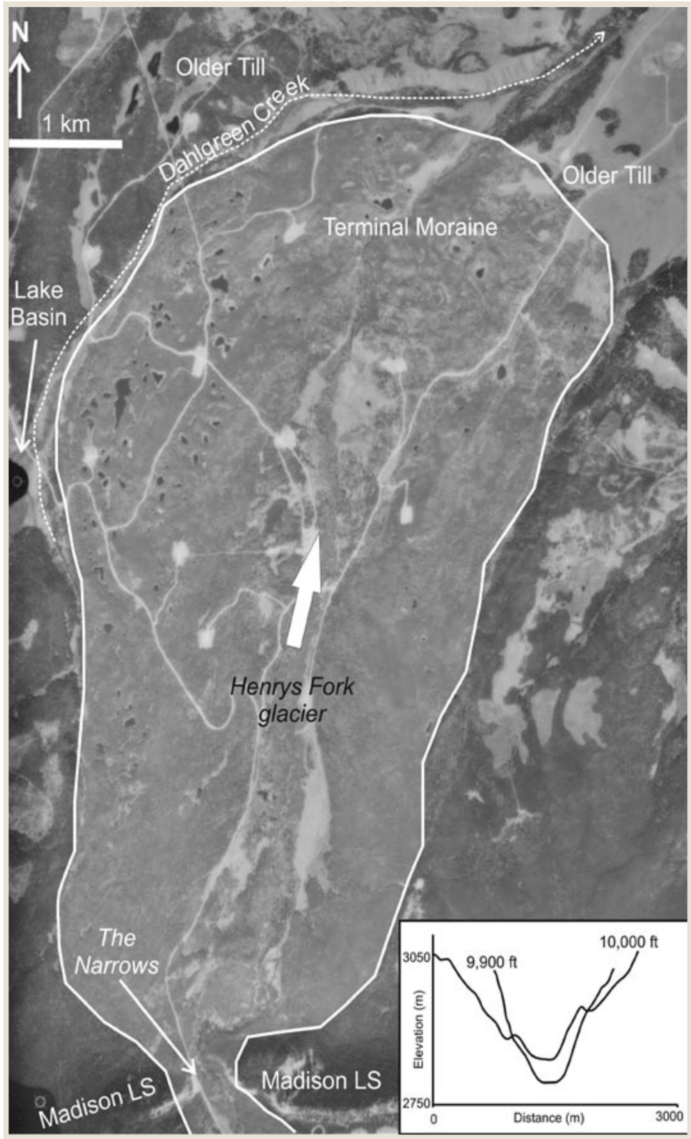


Figure 10. *Vertical air photo showing the terminus of the reconstructed Smiths Fork-age Henrys Fork glacier downstream from The Narrows. Dahlgreen Creek, which flows in an ice-marginal channel along the western side of the terminal moraine, was temporarily impounded to form an ice-marginal lake during the Smiths Fork Glaciation. The inset shows the relative dimensions of cross-sections through the Henrys Fork valley at elevations of 10,000 ft (top of lateral moraines, upstream from The Narrows, just off the southern boundary of this photograph) and 9,900 ft (within The Narrows). See text for discussion.*

Glaciation. Outwash graded to this moraine also merges downslope with Smiths Fork-age outwash from the West and Middle Forks of Beaver Creek. The moraine near Alligator Lake is most likely a rock-cored recessional feature formed when the retreating Smiths Fork-age glacier was temporarily pinned on a bedrock ledge.

Glaciers of the Western North Slope

Glaciers at the western end of the north slope were the largest (48 to 243 km^2, mean of 114 km^2) and longest (up to 36 km) in the northern Uintas, with three exceptions (West Fork Smiths Fork, West Fork Blacks Fork, Gold Hill) (table 1; Munroe-plate in CD). All of these large glaciers advanced to terminal positions far beyond the Paleozoic hogback, deposited fairly small, non-hummocky terminal moraines, and actively retreated upvalley during deglaciation. These characteristics indicate that glaciers at the western end were not influenced by the presence of the Paleozoic rocks, as these rocks do not generally form valley constrictions in this area.

The dominance of west-to-east moisture transport in this region, in the present and presumably in the past, generates greater precipitation over the western summits and supported larger and more erosive glaciers during each Quaternary glaciation. Additional evidence for generally greater erosion at the western end of the north slope comes from the overall drainage pattern. Some glaciers of the central and eastern north slope drained multiple cirques, but none of them had major tributary valleys. In contrast, glaciers in the Blacks Fork and Bear River valleys drained ice from four and five tributary valleys respectively (note the complexity values in table 1). These more elaborate upper drainages suggest that glacial erosion has been more successful at breaking down divides between western valleys over time.

The largest end moraine complex in the northern Uintas is present immediately downstream of the confluence of the Hayden and Stillwater Forks of the Bear River (Munroe-plate in CD). This end moraine complex, referred to informally as the Manor Lands moraine, is over 6.4 km wide and rises ~100 m above the outwash surface (figure 11). Massive boulders deposited directly from a supraglacial transport position dot the moraine front, as visible from the Mirrow Lake Highway (Highway 150) as it approaches the moraine from the north. Sandy till with abundant quartzite boulders is widespread near the valley center, whereas finer-grained diamicton is common near the valley margins. This gradation reflects proximity to the Wasatch Formation in the valley walls, and is well displayed along Forest Road 058 eastward toward Mill Creek from its junction with the Mirror Lake Highway. The Manor Lands moraine is fronted by an outwash surface up to 1.6 km wide, known as Hillard Flat, which extends northward into Wyoming. Remnant braided channel patterns exist on the outwash surface. Lone Mountain, a prominent upland in the center of the valley train at the Utah-Wyoming border, is a remnant of an older outwash surface related to an early Quaternary glaciation (Munroe-plate in CD).

The age of the Manor Lands moraine is unclear. One theory is that the moraine was deposited during the Smiths Fork Glaciation (e.g. Bryant, 1992). This explanation is supported by the good preservation of the lateral moraines and hummocky surface of the terminal moraine, but requires that the Smiths Fork-age Bear River Glacier advanced at least as far as the Blacks Fork-age glacier in this valley, a situation that was not matched in neighboring drainages. The Manor Lands moraine is also much larger than other Smiths Fork-age moraines on the north slope, and contains considerably more internal relief (~100 m). The alternate explanation is that the Manor Lands moraine was deposited during the Blacks Fork Glaciation and that the maximum

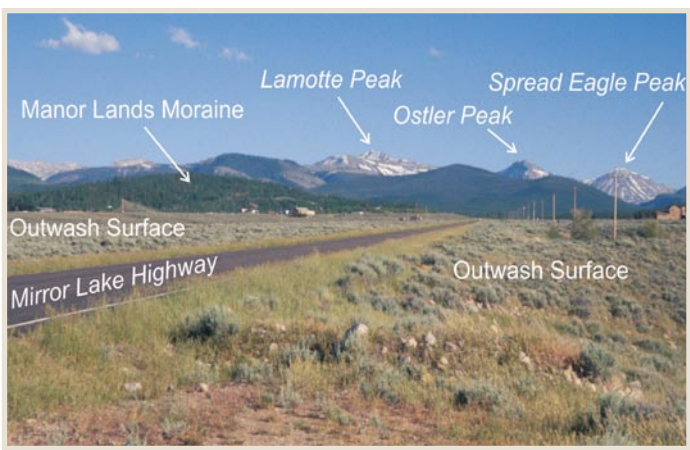


Figure 11. *The Manor Lands end moraine viewed from the Mirror Lake Highway, just south of the Wyoming/Utah state line. The moraine, which is considered Blacks Fork-age, rises almost 100 m above the outwash surface.*

Smiths Fork advance is recorded by end moraines farther upvalley near the confluence of the East Fork and Main Bear River (e.g. Atwood, 1909). This theory is consistent with the lack of other moraines downvalley from the Manor Lands moraine, and the presence of carbonate-enriched horizons in soil profiles visible in roadcuts along the Manor Lands road network. A slightly lower outwash terrace along the Mirror Lake Highway that wraps around the Manor Lands moraine and grades to the upvalley moraines, and narrow outwash valley trains that cut through the western sector of the Manor Land moraine, also support this interpretation. For now the Manor Lands moraine and Hillard Flat are considered Blacks Fork-age features, while the upvalley moraines and the lower outwash terrace represent the Smiths Fork Glaciation. Samples of erratic boulders for cosmogenic surface-exposure dating (^{10}Be and ^{26}Al) to resolve this issue were taken from the Manor Land moraine and upvalley moraines in 2003 and 2004, and analysis of these samples is underway in 2005.

Smiths Fork Equilibrium Line Altitudes and Moisture Sources

The equilibrium line is a hypothetical line across a glacier surface that divides the glacier into an upper accumulation zone where mass is gained and a lower ablation zone where mass is lost during the course of a balance year. The altitude of a glacier equilibrium line is climatically controlled, specifically by summer temperature and winter precipitation, and reconstruction of past equilibrium line altitudes (ELAs) provides information about paleoclimate. ELAs calculated for Smiths Fork-age glaciers of the northern Uintas are presented in figure 12 along with additional data points calculated from the mapping of Oviatt (1994) at the extreme western end of the range. The ELAs calculated in this study range from 2860 to 3230 m with a mean of 3100 m and a standard deviation of ~100 m (table 1). ELAs are lowest at the western end of the study area (~2900 m) while ELAs for the three glaciers in the center of the north slope are 80 to 130 m above the mean value. The ELA gradient over the western 60

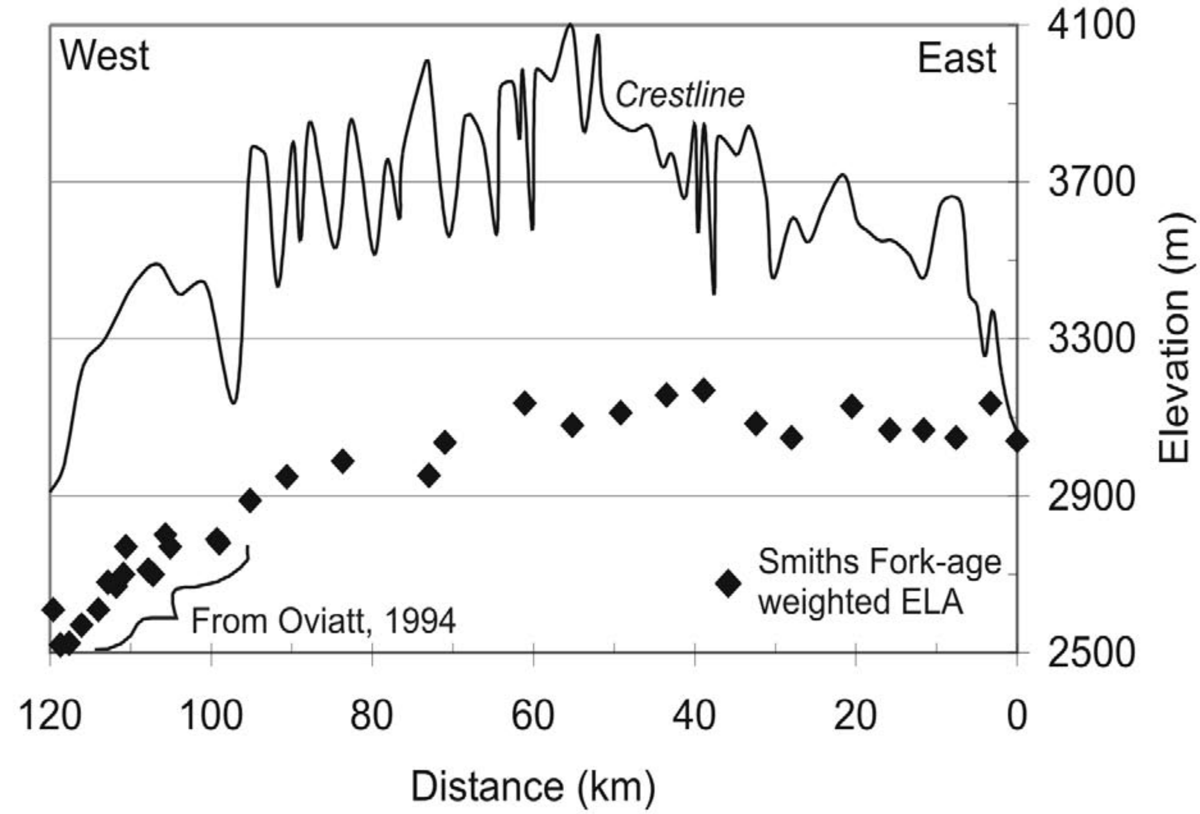


Figure 12. *Reconstructed Smiths Fork-age equilibrium line altitudes (ELAs) across the north slope, including the work of Oviatt (1994) west of the Mirror Lake Highway (modified from Munroe and Mickelson, 2002). The precipitous drop in ELA to the west probably reflects proximity to Lake Bonneville, which would have provided an extensive moisture source during the Smiths Fork Glaciation (Munroe and Mickelson, 2002).*

km of the study area is 6 m per km (370 m elevation change).

The reconstructed ELA during the Smiths Fork Glaciation compares favorably with the mean value of 3000 m proposed for the Uintas by Porter and others (1983) and with the range of 2620 to 3310 proposed by Schlenker (1995). Similarly, Flint (1971) indicated that the lowest elevation of cirque floors in the Uintas is about 3050 m. However, the east-west gradient in the ELA dataset (figure 12), which is reinforced by the mapping of Oviatt (1994), was not previously recognized. Because ELAs are a function of climate, this westward decrease indicates that either summer temperatures were cooler at the western end of the Uintas or that the western end of the range received more winter precipitation than the eastern part. Consideration of modern temperature gradients in the Uintas indicates that it is unlikely that temperature changes can explain this pattern. Using the modern atmospheric lapse rate of 6.7°C/km, mean summer temperatures at the lowest western ELAs should be 4°C warmer than those at the highest ELAs (600 m elevation difference). Thus, temperatures at ELAs in the western Uintas must have been decreased more than 4°C relative to similar elevations farther east to explain the observed ELA pattern by temperature changes alone, and there is no reason to suspect that lapse rates within the range at the LGM varied this dramatically.

It is far more likely that the ELA gradient reflects changes in precipitation across the north slope, with glaciers in the western Uintas receiving more winter snowfall to compensate for the warmer temperatures at their lower equilibrium lines. Orographic precipitation resulting from prevailing west-to-east moisture transport does impact the western Uintas and Moran (1974) and Zielinski and McCoy (1987) concluded that circulation patterns across the northern Great Basin during the late Pleistocene were similar to those today. However, the modern difference in winter precipitation between the central and western Uintas is much less than the difference required to explain the 600-m drop in ELA.

An additional moisture source available to Smiths Fork-age glaciers in the Uintas was Lake Bonneville, which was located ~100 km upwind. Lake Bonneville reached its maximum surface elevation of 1550 m around 15,000 ^{14}C yrs BP, covering an area of 51,500 km^2 (Currey, 1990; O'Connor, 1993). Although the relative timing of the Bonneville highstand and the Smiths Fork Glaciation is unclear, the lake was present during the Smiths Fork Glaciation, and was certainly larger than the modern Great Salt Lake (Oviatt, and others, 1992; Oviatt, 1997). Numerical simulations of the LGM paleoclimate in the Bonneville Basin suggest that mean maximum temperatures were below 0°C only in January and December (Craig and others, 1997; Timofeyeva and others, 1999) and that open water was present throughout the winter (Hostetler and others, 1994). This moisture source, combined with the abrupt topography of the Wasatch Front and western Uintas, would have produced tremendous lake effect snow. While modern lake effect snows are generally restricted to the western slope of the Wasatch Mountains (Carpenter, 1993), lake effect snow enhanced by the larger surface area of Lake Bonneville may have extended into the western Uintas, allowing glaciers there to remain in mass balance at lower, warmer elevations (Munroe and Mickelson, 2002).

NEOGLACIATION IN THE NORTHERN UINTA MOUNTAINS

Two areas of the north slope have extensive and complicated records of glaciation during the late Holocene "Neoglacial" period: the Dead Horse Lake area at the head of the West Fork Blacks Fork and the large cirque at the head of the East Fork Blacks Fork. The Dead Horse Lake area contains the best evidence for Neoglacial ice with multiple, distinct moraines present on the alpine upland west of Dead Horse Lake and along the foot of the steep headwall west of Dead Horse Pass (figure 13). The moraines range from 3 to 15 m in height, and are composed of quartzite blocks up to 5 m in length. Two additional end moraines are submerged in Dead Horse Lake where they form shoals visible from Dead Horse Pass under favorable light conditions. Mapping of the moraines identified crosscutting relationships that allowed for establishment of a relative age chronology (figure 13, table 2). In an attempt to determine the ages of these moraines, Munroe (2002) measured ***Rhizocarpon geographicum (L.)*** lichens at eight study sites: two sites on bare ledges in front of the moraines, five on discrete moraine ridges,

Table 2. *Site Relative Age and Absolute Age from Largest Rhizocarpon#*

Relative Order	Diameter mm	Age* yr	Age** yr	Age*** yr
7	56	1370	1300	1470
6	58	1430	13606	1530
5	59	1460	1400	1550
landslide	60	1500	1430	1580
4	60	1500	1430	1580
3	63	1590	1530	1660
2	65	1650	1590	1710
1	85	2250	2250	2240

Modifed from Munroe, 2002
* 14 mm/100 yr for first century, then 3.3 mm/century (Benedict, 1967)
**y = 32.6636x - 504 for x in mm and lichens > 20 mm (Benedict, 1993)
***0.038 mm/yr (Mahaney, 1987)

and one on a landslide deposit that crosscuts the moraine loops (figure 13). Two growth-rate curves derived for ***R. geographicum*** in the Colorado Front Range (Benedict, 1967; 1993) and one for the Wind River Range (Mahaney, 1987) were applied to the Dead Horse Lake lichens to estimate moraines ages. Although the equations were not developed specifically for the Uintas, ***R. geographicum*** growth rates are relatively constant between alpine areas (Locke and others, 1979). This approach suggests that the moraines west of Dead Horse Lake are at least 1300 to 2250 years old. In contrast, Gannett Peak (Little Ice Age) moraines in the Wind River Range are considered to be only a few centuries old (Dahms and Birkeland, 2000), and Matthes (Little Ice Age) moraines in the Sierra Nevada were deposited since ~650 yr BP, (Wood, 1977; Clark and Gillespie, 1997). Thus, the moraines studied at Dead Horse Lake were all deposited during an episode of Neoglaciation that predated the Little Ice Age, roughly synchronous with deposition of the Black Joe Alloformation (2000 to 1500 BP) in the Wind River Range (Dahms, 2002).

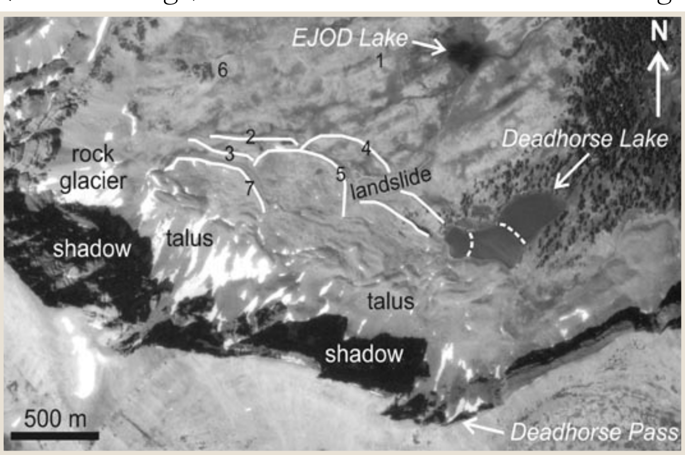


Figure 13. *Vertical air photo showing post-Smiths Fork-age end moraines (white lines) in the Deadhorse Lake area at the head of the West Fork Blacks Fork (modified from Munroe, 2002). Lichen sample sites are identified with their relative age (see Table 2). The narrow landslide cross-cuts some of the moraines and has maximum lichen diameters identical to Site 4.*

The deep cirque at the head of the East Fork Blacks Fork also contains an intriguing Neoglacial record. Multiple moraines from 3 to 20 m tall are present on the floor of the cirque, and the lowest one impounds a small basin floored by fine-grained sediment. Upslope from this basin, numerous end moraines cross the valley between continuous lateral moraine ridges. ***R. geographicum*** lichens on these landforms have diameters similar to those measured at Dead Horse Lake, suggesting that small cirque glaciers fluctuated in each location simultaneously in the late Holocene.

Small moraines are also present at lower elevations and greater distances from cirque headwalls in north slope valleys, including the Henrys Fork where a moraine crosses the valley near Dollar Lake ~4 km from the headwall. A similar moraine impounds a small basin at the head of the Left Hand Fork Bear River. Peat and wood *(Salix)* fragments obtained from behind both moraines were radiocarbon dated to before ~10 ka BP (Munroe, 2002), indicating that these lower cirque floor moraines were formed during episodes of cooler climate that interrupted the transition from the Smiths Fork Glaciation to the Holocene.

The post-Smiths Fork glacial history of the northern Uintas, therefore, includes at least two episodes of renewed glaciation: one in the latest Pleistocene, and a second during the late Holocene Neoglaciation. Given that glaciers formed and/or advanced during the Little Ice Age (ca. A.D. 1250-1850) in many parts of Rocky Mountains, it is notable that correlative landforms have not been positively identified in the northern Uintas. Munroe (2003) used rephotography to determine that summer temperatures in the northern Uintas were depressed ~1°C below modern values in A.D. 1870, a change that certainly would have been favorable for the formation of small glaciers in shaded locations. To explain the lack of evidence for Little Ice Age glacial activity in the Blacks Fork, Munroe (2002) hypothesized that conditions during the Little Ice Age were too dry in the northern Uintas for the formation of active cirque glaciers. Instead, the slightly cooler and drier climate enhanced periglacial processes, giving rise to the rock glaciers that are ubiquitous at higher elevations (>3000 m) in the northern Uintas (figure 14). A minority of these rock glaciers have unvegetated frontal slopes standing beyond the angle of repose, meandering longitudinal and transverse furrows, surface meltwater ponds, and lateral springs that discharge silty meltwater at 0°C in late summer—characteristics suggesting that they are active today (Wahrhaftig and Cox, 1959). The majority of rock glaciers in the Uintas, however, appear dormant. The combination of a few active rock glaciers and a large number of apparently dormant features suggests that climatic conditions at higher elevations of the Uintas are close to being suitable for widespread rock glacier activity. The entire population of rock glaciers, both dormant and active, may, therefore, have originated during the slightly cooler Little Ice Age, with only those situated in the most ideal locations remaining active today.

Figure 14. *Rock glacier near Weyman Lakes at the head of the South Fork Sheep Creek viewed from the ridgeline above Pearl Lake. The rock glacier is 600 m long, and has a terminus 250 m wide. The vertical drop from the rocks in the foreground to the rock glacier surface is 250 m. Note the prominent longitudinal furrows and the turbid water in the pond below the terminus, both of which are characteristics of active rock glaciers.*

SUMMARY

The northern Uinta Mountains contain a fascinating geomorphic record of Quaternary glaciations. The type localities for the last two glacial cycles in the Uintas are located in the Smiths Fork and Blacks Fork valleys, and moraines representing these glaciations are present in valleys throughout the range. During the most recent Smiths Fork Glaciation, glaciers at the eastern end of the north slope were relatively small, yet formed broad piedmont lobes where they advanced beyond glacial canyons. The terminal moraines of these glaciers are dramatically hummocky, indicating that large amounts of buried ice were involved in their formation. Similarly, the formation of hummocky Smiths Fork-age moraines in the South Fork Ashley Creek appears to have been influenced by an unusually large load of supraglacial debris. Smiths Fork-age glaciers in the central part of the north slope advanced to positions where they were influenced by the prominent hogback of Paleozoic limestone and (locally) sandstone. Some of these glaciers were unable to advance beyond the Paleozoic rocks, and deposited terminal moraines in close association with the hogback. Others were able to advance beyond this topographic obstacle, but quickly stagnated during deglaciation as surface downwasting reduced the flux of ice into their terminal regions. In contrast, glaciers of the far western north slope were several times larger than those farther east during the Smiths Fork Glaciation, and were unaffected by the Paleozoic rocks, which had been completely eroded away during prior Quaternary glaciations. Glaciers in the western Uintas during the Smiths Fork Glaciation also received enhanced snow accumulation, as evidenced by an equilibrium line altitude drop of 600 m from the center of the north slope to the western end. Glacial activity after the Smiths Fork Glaciation was limited to a few small cirque-floor moraines that formed during the latest Pleistocene, and other high-elevation moraines, particularly in the Blacks Fork drainage, that formed during the late Holocene Neoglaciation. Conditions during the latest Holocene Little Ice Age were hypothetically too dry for cirque glaciers to form and advance. However, rock glaciers, which are common components of the high elevation landscape across the north slope, may have developed at this time.

ACKNOWLEDGMENTS

Support for my research in the northern Uintas came from the Ashley National Forest and the Wasatch-Cache National Forest, the Department of Geology & Geophysics at the University of Wisconsin-Madison, the Graduate School of the University of Wisconsin-Madison, Middlebury College, the Geological Society of America, USGS EDMAP, Sigma Xi, the NSF-AMS Radiocarbon lab at the University of Arizona-Tucson, and the American Alpine Club. My work in the Uintas would not have been possible without the assistance of D. Koerner of the Ashley National Forest. E. Carson, M. Devito, D. Douglass, J. Martin Duque, L. Martin de Frutos, B. Laabs, D. Munroe, S. Munroe, C. and N. Oprandy, C. Rodgers, and J. Shakun all provided field assistance. I appreciate the guidance of my graduate advisors D. Mickelson and J. Bockheim. E. Carson, C. Dehler, B. Laabs, and J. Pederson provided reviews that were useful in improving the original manuscript.

REFERENCES

Ackerly, S.C., 1989, Reconstruction of mountain glacier profiles, northeastern United States: Geological Society of America Bulletin, v. 101, p. 561-572.

Atwood, W.W., 1909, Glaciation of the Uinta and Wasatch Mountains: United States Geological Survey Professional Paper 61, 96 p.

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

Benedict, J.B., 1967, Recent glacial history of an alpine area in the Colorado Front Range, U.S.A. I. Establishing a lichen-growth curve: Journal of Glaciology, v. 6, p. 817-832.

– 1993, A 2000-year lichen-snowkill chronology for the Colorado Front Range, U.S.A.: The Holocene, v. 3, p. 27-33.

Bockheim, J.G., and Koerner, D., 1997, Pedogenesis in alpine ecosystems of the eastern Uinta Mountains, Utah, U.S.A.: Arctic and Alpine Research, v. 29, no. 2, p. 164-172.

Bockheim, J.G., Munroe, J.S., Douglass, D. C., and Koerner, D., 2000, Soil development along an elevational gradient in the southeastern Uinta Mountains, Utah, USA: Catena, v. 39, p. 169-185.

Bradley, W.A., 1936, Geomorphology of the north flank of the Uinta Mountains: United States Geological Survey Professional Paper 185-I, p. 163-199.

Bryant, B., 1992, Geologic and structure maps of the Salt Lake City 1° x 2° quadrangle, Utah and Wyoming: USGS Miscellaneous Investigations Series Map I-1997, scale 1:125,000.

Carpenter, D.M., 1993, The lake effect of the Great Salt Lake – overview and forecast problems: Weather and Forecasting, v. 8, p. 181-193.

Chadwick, O.A., Hall, R.D., and Phillips, F.M., 1997, Chronology of Pleistocene glacial advances in the central Rocky Mountains: Geological Society of America Bulletin, v. 109, no. 11, p. 1443-1452.

Clark, D.H., and Gillespie, A.R., 1997, Timing and significance of late-glacial and Holocene cirque glaciation in the Sierra Nevada, California: Quaternary International, v. 38/39, p. 21-38.

Craig, R.G., Timofeyeva, M.M., and Clement, S.M., 1997, Simulation of climate and runoff for the Bonneville Basin at the last glacial maximum [abs]: Geological Society of America Abstracts with Programs, v. 29, no. 6, p. 254.

Currey, D.R., 1990, Quaternary palaeolakes in the evolution of semidesert basins, with special emphasis on Lake Bonneville and the Great Basin: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 76, p. 189-214.

Dahms, D.E., 2002, Glacial stratigraphy of Stough Creek Basin, Wind River Range, Wyoming: Geomorphology, v. 42, no. 1-2, p. 59-83.

Dahms, D.E., and Birkeland, P.W., 2000, Glacial stratigraphy of Stough Creek basin and a latest Quaternary glacial and periglacial stratigraphy for the Wind River Range, Wyoming [abs]: Geological Society of America Abstracts with Programs, v. 32, no. 7, p. 16.

Douglass, D.C., 2000, Glacial History of the West Fork of Beaver Creek, Uinta Mountains, Utah: Madison, University of Wisconsin, M.S. thesis 64 p.

Flint, R.F., 1971, Glacial geology and the Pleistocene epoch: New York, John Wiley and Sons,.892 p.

Gilmer, D.R., 1986, General geology, landsliding, and slope development of a portion of the north flank of the Uinta Mountains, south-central Uinta County, Wyoming: Laramie, University of Wyoming, M.S. thesis, 93 p.

Gosse, J.C., Klein, J., Evenson, E.B., Lawn, B., and Middleton, R., 1995, Beryllium-10 dating of the duration and retreat of the last Pinedale glacial sequence: Science, v. 268, p. 1329-1333.

Grogger, P.K., 1974, Glaciation of the High Uintas Primitive Area, Utah, with emphasis on the northern slope: Salt Lake City, University of Utah, Ph.D. dissertation, 209 p.

– 1975, Neoglaciation of the northern slope of the High Uintas Primitive Area, Utah: Great Plains Rocky Mountain Geographical Journal, v. 4, p. 37-43.

Hostetler, S.W., Giorgi, F., Bates, G.T., and Bartlein, P.J., 1994, Lake-atmosphere feedbacks associated with paleolakes Lahontan and Bonneville: Science, v. 263, no. 5147, p. 665-669.

Laabs, B.J.C., Shakun, J.D., Munroe, J.S., Mickelson, D.M., Singer, B.S., and Caffee, M., 2003, Cosmogenic-exposure age limits on the last glacial maximum in the south-central Uinta Mountains, northeastern Utah [abs]: Geological Society of America Abstracts with Programs, v. 35, no. 6, p. 541.

Locke, W.W., Andrews, J.T., and Webber, P.J., 1979, A manual for Lichenometry: British Geomorphological Research Group Technical Bulletin No. 26. 47 p.

– 1990, Late Pleistocene glaciers and the climate of western Montana, U.S.A.: Arctic and Alpine Research, v. 22, p. 1-13.

Mahaney, W.C., 1987, Tentative growth curve for Rhizocarpon geographicum s.l. in Stroud Basin, Wind River Range, western Wyoming: Northwest Science, v. 61, no. 7, p. 13-19.

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

Moran, J.M., 1974, Possible coincidence of a modern and a glacial-age climatic boundary in the montane west, United States: Arctic and Alpine Research, v. 6, p. 319-321.

Munroe, J.S., 2001, Late Quaternary History of the northern Uinta Mountains, northeastern Utah: Madison, University of Wisconsin, Ph.D. dissertation, 398 p.

– 2002, Timing of post-glacial cirque reoccupation in the northern Uinta Mountains, northeastern Utah, U.S.A.: Arctic, Antarctic and Alpine Research, v. 34, no. 1, p. 38-48.

– 2003, Estimates of Little Ice Age climate inferred through historical rephotography, northern Uinta Mountains, U.S.A.: Arctic, Antarctic and Alpine Research, v. 35, no. 4, p. 489-498.

Munroe, J.S., and Douglass, D.C., 1998, Physical and chemical evidence for eolian additions to soils of the eastern Uinta Mountains, northeastern Utah [abs]: Geological Society of America Abstracts With Programs, v. 30, no. 7, p. 138.

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, v. 48, no. 161, p. 257-266.

O'Connor, J.E., 1993, Hydrology, Hydraulics, and Geomorphology of the Bonneville Flood: Geological Society of America Special Paper 274, 83 p.

Oviatt, C.G., Currey, D.R., and Sack, D., 1992, Radiocarbon chronology of Lake Boneville, eastern Great Basin, U.S.A.: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 99, p. 225-241.

Oviatt, C.G., 1994, Quaternary geologic map of the Upper Weber River drainage basin, Summit County, Utah: Utah Geological Survey Map 156, 1:50,000 scale.

– 1997, Lake Bonneville fluctuations and global climate change: Geology, v. 25, no. 2, p. 155-158.

Phillips, F.M., Zreda, M.G., Gosse, J.C., Klein, J., Klein, J., Evenson, E.B., Hall, R.D., Chadwick, O.A., and Sharma P., 1997, Cosmogenic 36Cl and ^{10}Be ages of Quaternary glacial and fluvial deposits of the Wind River Range, Wyoming: Geological Society of America Bulletin, v. 109, no. 11, p. 1453-1463.

Porter, S.C., Pierce, K.L., and Hamilton, T.D., 1983, Late Wisconsin mountain glaciation in the western United States, in Wright, H.E., editor, Late-Quaternary environments of the United States, Volume 1, The late Pleistocene: Minneapolis, University of Minnesota Press, p. 71-111.

Richmond, G.M., 1965, Glaciation of the Rocky Mountains, in Wright, H.E., Jr. and D.G. Frey editors, The Quaternary of the United States: Princeton, Princeton University Press, p. 217-230

– 1986, Stratigraphy and correlation of glacial deposits of the Rocky Mountains, the Colorado Plateau and the ranges of the Great Basin: Quaternary Science Reviews, v. 5, p. 99-127.

Ritzma, H.R., 1983, Igneous dikes of the Eastern Uinta Mountains, Utah and Colorado: Utah Geological and Mineral Survey, Special Studies 56, 23 p.

Schlenker, G.C., 1988, Glaciation and Quaternary geomorphology of the Blacks Fork drainage, High Uinta Mountains, Utah and Wyoming: Salt Lake City, University of Utah, M.S. thesis, 87 p.

– 1995, Reconstruction of late-Pleistocene equilibrium-line altitudes and paleoenvironments: Uinta Mountains, Utah and Wyoming: College Station, Texas A&M University, Ph.D. dissertation, 145 p.

Schoenfeld, M.J., 1969, Quaternary geology of the Burnt Fork area, Uinta Mountains, SummitCounty, Utah: Laramie, University of Wyoming, M.S. thesis, 80 p.

Shakun, J.D., 2003, Last glacial maximum equilibrium-line altitudes and paleoclimate, northeastern Utah: Middlebury College, B.A. thesis, 55 p.

Sharp, W.D., Glaser, L.L., Ludwig, K.R., Chadwick, O.A., and Amundson, R., 2003, Dating fluvial terraces by ^{230}Th/U on pedogenic carbonate, Wind River Basin, Wyoming: Quaternary Research, v. 59, no. 2, p. 139-150.

Timofeyeva, M., Clement, S., and Craig, R., 1999, Estimates of Glacial Climate and Runoff for the Bonneville Basin, in Wilson R. and L. Buffaloe, editors, Proceedings of the Fifteenth Annual Pacific Climate (PACLIM) Workshop, April 27 - 30, 1998: Interagency Ecological Program, Technical Report 64, State of California Department of Water Resources, p. 9-19.

Wahrhaftig, C., and Cox, A., 1959, Rock glaciers in the Alaska Range: Geological Society of America Bulletin, v. 70, p. 383-436.

Wood, S.H., 1977, Distribution, correlation, and radiocarbon dating of late Holocene tephra, Mono and Inyo Craters, eastern California: Geological Society of America Bulletin, v. 88, p. 89-95.

Zielinski, G.A., and W.D. McCoy, 1987, Paleoclimatic implications of the relationship between modern snowpack and late Pleistocene equilibrium-line altitudes in the mountains of the Great Basin, Western U.S.A.: Arctic and Alpine Research, v. 19, no. 2, p. 127-134.

Zimmer, T.M., 1996, Pedogenesis on quartzite-rich Pleistocene moraines, Smith's Fork drainage, Uinta Mountains: Logan, Utah State University, M.S. thesis, 118 p.