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Timing of Postglacial Cirque Reoccupation in the Northern Uinta Mountains, Northeastern Utah, U.S.A.

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Abstract

Radiocarbon-dated organics from lacustrine cores and colluvial basins, combined with measurements of lichens growing on moraines previously interpreted as Neoglacial in age, argue that the extent of post-Pleistocene ice in the northern Uinta Mountains was limited. An AMS date of 12,190 ± 120 ^{14}C yr BP on basal organic sediment from a tarn indicates that deglaciation was locally complete by ~14 ka BP. Additional radiocarbon ages of ~10 ka BP from two colluvial basins behind end moraines suggest that these moraines date to the latest Pleistocene. Cross-cutting moraines from a fourth site indicate that an active cirque-glacier system was present at some point after the terminal Pleistocene deglaciation. Maximum diameters of *Rhizocarpon geographicum* lichens on these moraines, however, suggest they may predate Little Ice Age deposits in the Wind River Range and Sierra Nevada. The convergence of evidence suggests that glaciers in the northern Uintas during the Holocene were extremely limited in extent, and restricted to the middle Neoglacial period. This preliminary conclusion contradicts some previous interpretations of the upper subalpine geomorphology of the range, which postulated widespread post-Pinedale glaciation, and multiple stades of Holocene Neoglaciation.

Introduction

Understanding the magnitude of Holocene climate changes is critical to predictions of how climate may change in the future, including the remainder of the current interglacial period. Studies over the past 50 yr have recognized that Holocene climate changes, while minor in comparison with those of the latest Pleistocene, were nevertheless sufficient to cause alpine glacier rejuvenation in cirques that had become ice-free during the middle Holocene warm period known as the "Altithermal" (Antevs, 1948; Grove, 1990). The overall expansion or rebirth of alpine glaciers during the mid-to-late Holocene is termed the "Neoglaciation," after Moss (1951) who first applied this expression to cirque glaciers in the Wind River Range, and Porter and Denton (1967) who employed the term more widely. Although global synchronicity of Neoglacial advances remains unproven, peaks of Neoglaciation occurred in many mountain ranges during the Little Ice Age, between approximately A.D. 1500 and 1850 (Matthes, 1939; Porter and Denton, 1967; Lamb, 1977).

Davis (1988, 2000) surveyed evidence for Holocene glacier fluctuations in the western U.S., building on previous work by Birkeland et al. (1971) and Burke and Birkeland (1983). Davis (1988) acknowledged abundant evidence for Holocene glacial advances, especially after ~5 ka BP, but cautioned that correlation of these advances could only be addressed through the acquisition of additional radiometric ages. Other investigators have developed chronologies for Neoglaciation in specific mountain ranges. In Colorado, relative age parameters and radiocarbon dating have been used to support a tripartite division of Neoglacial deposits (Benedict, 1973, 1985). In the Wind River Range, Wyoming, Dahms and Birkeland (2000) and Dahms (2002) used a combination of field mapping and both relative and numeric age techniques to outline a three-fold Neoglacial sequence with deposition sometime after 5.0 ka BP, about 1000 BP, and during the past few centuries. In the Sierra Nevada, California, the Little Ice Age is represented by moraines of the "Matthes" advance, which were deposited sometime after ~650 yr BP based on tephrachronology (Wood, 1977). Multiple AMS radiocarbon dates demonstrate that the "Recess Peak" advance of Birman (1964), previously considered Holocene in age, actually occurred before ~11,000 ^{14}C yr BP (Clark and Gillespie, 1997). This result indicates that the Sierra Nevada were ice free through the Younger Dryas subchron, and until the latest Holocene, a conclusion that runs contrary to reports of Younger Dryas advances in the Wind River Range and Colorado Front Range (Zielinski and Davis, 1987; Clark and Gillespie, 1997; Menounos and Reasoner, 1997; Davis et al., 1998).

Although evidence is converging on a unified Neoglacial model for the mid-latitude mountains of the western U.S., study of the Neoglacial record from additional ranges is necessary to better understand the possible equivalence and correlation of advances across the region. The northern Uinta Mountains are ideally situated to fill a gap between the relatively well-studied ranges of Wyoming, Colorado, and the Sierra Nevada. This paper presents an investigation of upper subalpine glacial deposits in the northern Uintas that postdate the latest Pleistocene. Radiocarbon dating of lacustrine sediments yields a minimum limit on the timing of deglaciation in the range, while organics retrieved from colluvial basin-fills and lichens growing on end moraines provide information about the record of glaciation in the range during the Holocene.

GENERAL SETTING

The Uinta Mountains reach elevations of over 4 km along a 150-km-long trend across northeastern Utah (Fig. 1). Structurally the range is a block of slightly metamorphosed Precambrian sediments that was uplifted during the Laramide Orogeny. No

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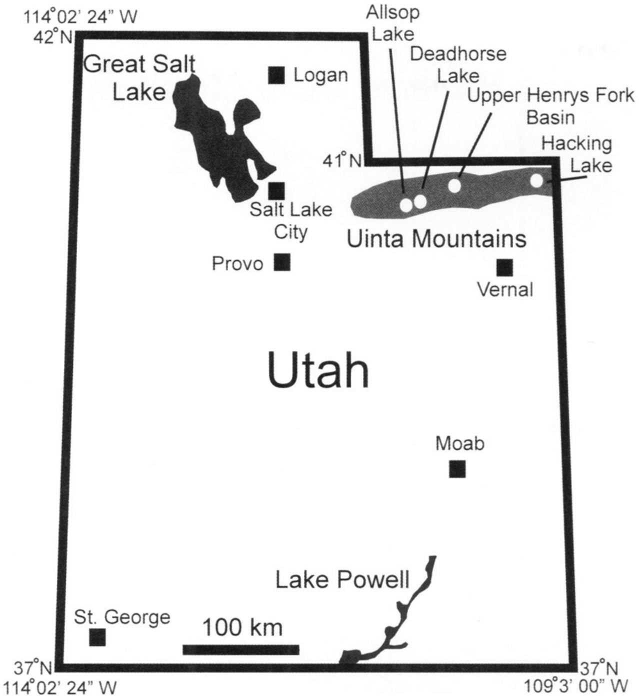

FIGURE 1. Location of the Uinta Mountains in northeastern Utah, and the four localities investigated in this study. Major cities and water bodies are shown for reference.

glaciers are present in the Uintas today, but the range was extensively glaciated by alpine-style glaciers during the Pleistocene. Cirque-floor moraines are present at elevations from 3100 to 3400 m a.s.l. (~500 m above the late Pleistocene terminal moraines) in most of the valleys of the northern Uintas, indicating that glaciers locally persisted or reformed after the terminal Pleistocene deglaciation.

The locations of the four sites investigated in this study are highlighted in Figure 1. Hacking Lake, a tarn at the head of the easternmost glaciated drainage in the Uintas, was cored to provide a limit on the timing of deglaciation in the range. Local basins behind cirque-floor moraines in the upper Henrys Fork basin and above Allsop Lake were studied to determine the age of the moraines and the sedimentary history of the upper cirque basins. Finally, lichenometry was applied to determine the relative age of high-elevation moraines near Deadhorse Lake.

PREVIOUS WORK

There have been few previous studies of the Holocene glacial record in northern Utah. Madsen and Curry (1979) determined that the most recent glacial activity in Little Cottonwood Canyon in the Wasatch Mountains southeast of Salt Lake City occurred sometime before 7515 ± 180 ^{14}C yr BP. They identified two episodes of cooler climate corresponding to the Neoglacial period from palynological evidence, and concluded that these episodes were accompanied by formation of rock glaciers and protalus ramparts, but not true glaciers. Slightly farther south, Anderson and Anderson (1981) used weathering rinds developed on calcareous quartzarenite clasts to argue that only rock glaciers occupied cirques on Mount Timpanogos during the Neoglaciation.

In the Uintas, Schoenfeld (1969) investigated glacial deposits at the eastern end of the range. He concluded that the Little Ice Age was represented only by periglacial features, and that the cirque-floor moraines, found 2 to 3 km from the headwalls, were of Temple Lake age. At the time, Temple Lake deposits were considered early Neoglacial, but subsequent work has concluded that the type Temple Lake deposits and their correlates in the Wind River Range actually date to the latest Pleistocene (Zielinski and Davis, 1987; Davis et al., 1998).

Barnhardt (1973) reported on the glacial and periglacial geomorphology of the Bald Mountain area in the western Uintas. He employed both lichenometry and radiocarbon dating to develop a chronology for the late stages of the Pinedale and the Neoglaciation. Barnhardt concluded that glaciers last occupied the cirques of Bald Mountain sometime before 8235 ± 345 ^{14}C yr BP, and that abundant talus cones, avalanche boulder tongues, and protalus ramparts represent the Neoglacial climate deterioration.

The most extensive study of cirque-floor moraines and putative Neoglacial features in the Uintas was undertaken by Grogger (1974, 1975) who divided these deposits into groups on the basis of geomorphology, vegetative cover, soil development, stratigraphic relations, and degree of erosional modification. Grogger concluded that semiquantitative analysis of these parameters supported a four-fold division of the Neoglacial deposits in the northern Uintas: three stades in which actual glaciers were present, and a fourth and final stade marked only by intense periglacial activity.

Methods

FIELD MAPPING AND SAMPLING

Geomorphic features investigated as part of this study were identified on stereo air photos and 1:24,000 scale topographic maps with a 40-ft (12-m) contour interval. All localities were field checked to evaluate the map and air photo interpretations. Mapping was compiled on a rectified orthophoto base, and digitized into a Geographic Information System.

The two local basins sampled in this project were discovered behind cirque-floor moraines during field reconnaissance. Inspection of stream cutbanks through the basin fills revealed organic material, including *Salix* fragments, at the contact between the fill and the underlying diamicton. Samples of these organic sediments were collected in sterile plastic bags, frozen within 2 d, and kept frozen until submission for radiocarbon dating less than 2 mo later.

CORING

The lacustrine sediment core was retrieved in March 1998, using a Livingstone corer with a 2.5-cm-diameter barrel (Livingstone, 1955). The water depth at the coring site was 345 cm, and a total of 200 cm of core was retrieved in four segments spanning 260 cm (77% recovery). Core segments were sealed in plastic and foil for transport back to the University of Wisconsin-Madison where they were processed. Organic samples for AMS radiocarbon dating were refrigerated in distilled water in glass vials. Radiocarbon ages were calibrated using CALIB 4.3 (Stuiver et al., 1998, 2000). For clarity, uncalibrated radiocarbon ages in this report are designated by ^{14}C.

LICHENOMETRY

Studies have documented that the diameter of lichens increases with the time since lichen germination and that the size of lichens growing on late Holocene glacial deposits can be considered an indicator of the age of the deposits (Beschel, 1961;

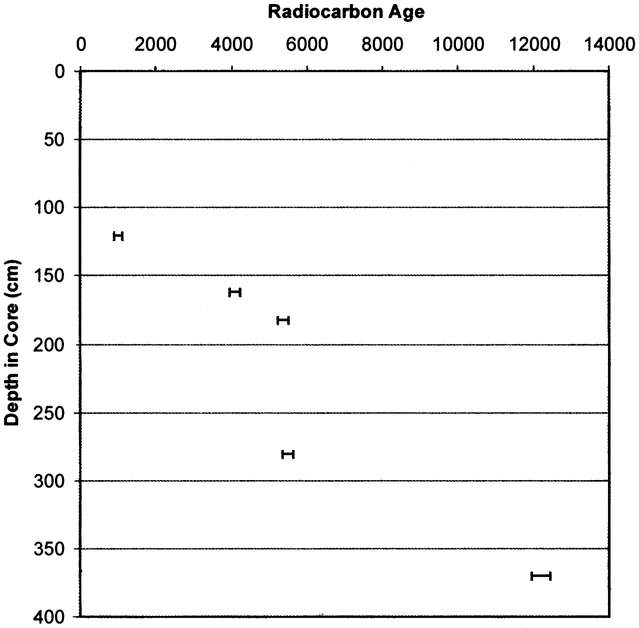

FIGURE 2. Depth-Age relationship for the Hacking Lake core. Depths are in cm below the sediment/water interface, and ages are in ^{14}C yr BP with 2-sigma error bars.

Locke et al., 1979; Bull et al., 1994; Konrad and Clark, 1998). Assuming that growth conditions are uniform between the surfaces compared, the largest lichen on a surface is an indicator of relative surface age because colonization occurs soon after the surface is exposed (Webber and Andrews, 1973; Bull and Brandon, 1998). Measurement of the diameters of lichens on surfaces of known ages has also allowed the construction of growth-rate curves for specific lichen species (Benedict, 1967, 1993; Curry, 1968, 1969). Therefore, lichens can provide age control on otherwise inorganic glacial deposits where other dating methods are inapplicable.

For this project, *Rhizocarpon geographicum* (L.) lichens were measured on high-elevation moraines near Deadhorse Lake (Fig. 1) considered Neoglacial by a previous study (Grogger, 1975). Eight study sites were established: two on bare ledges in front of the moraines, five on discrete moraine ridges, and one on a rock avalanche deposit that crosscuts the moraine loops. At each of these sites, the maximum diameter of the largest *Rhizocarpon,* identified through a 30-min search, was determined to the nearest 1.0 mm. Following the methods of Locke et al. (1979), only circular or nearly circular lichens growing without contact with neighboring thalli were measured. Thirty-three additional lichens were measured at each study site to investigate the population distribution of lichen sizes and to identify lichens with anomalously large diameters that may have been growing on boulders before they were detached from the cirque headwall. Two lichen growth-rate equations presented by Benedict (1967, 1993) for *Rhizocarpon geographicum* in the Colorado Front Range, and a third equation derived by Dahms (2002) for the Wind River Range, were applied to estimate the ages of the lichen-covered surfaces.

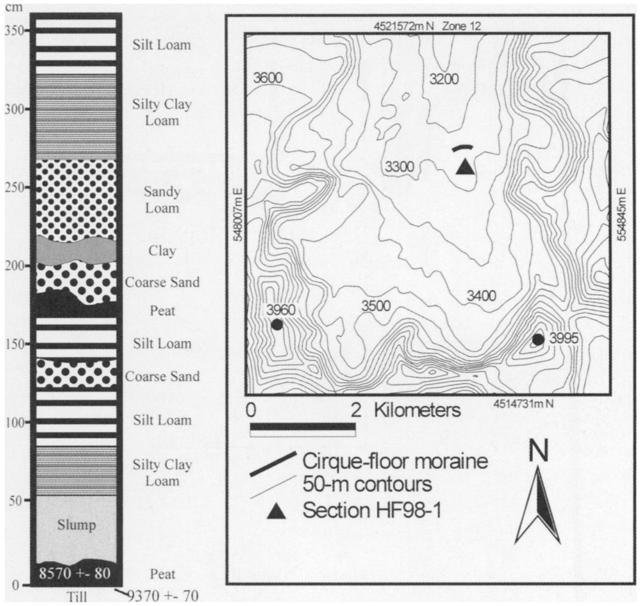

FIGURE 3. Map of the upper Henrys Fork Basin and stratigraphy of the Henrys Fork site behind the cirque-floor moraine. See Figure 1 for the location of the upper Henrys Fork Basin in the Uinta Mountains. The stratigraphic column for this exposure (scale in cm) is shown at left. The two basal radiocarbon dates are given in ^{14}C yr BP.

Results

RADIOCARBON DATES

Daphnia ephippia (sample HL-715) isolated from a sand layer near the base of the Hacking Lake core returned an age of 12,190 ± 120 ^{14}C yr BP, which calibrates to between 15.0 and 13.9 ka BP (Table 1). This is the oldest radiocarbon date obtained from anywhere in the Uinta Mountains. Other dates on terrestrial macrofossils and organic fragments from higher levels

TABLE 1

AMS radiocarbon dates from cirques in the Uinta Mountains

Sample	^{14}C age ± 1 sig (yr BP)	Sample material	1 Sigma range	Lab no.
HF98-1-26	9370 ± 70	Wood	9662–9478	Beta-134567
HF99-01	8570 ± 80	Woody peat	10415–10305	Beta-125001
AL99-01	8950 ± 70	Wood	10197–9925	Beta-134565
AL99-02	8730 ± 90	Organic silt	9888–9556	Beta-134566
HL-465	1000 ± 55	Wood	966–796	AA32838
HL-506	4105 ± 70	Conifer needle	4810–4522	AA32839
HL-528	5365 ± 75	Wood	6272–6001	AA32840
HL-625	5495 ± 75	Organic fragment	6397–6200	AA32842
HL-715	12190 ± 120	*Daphnia ephippia*	15019–13847	AA35250

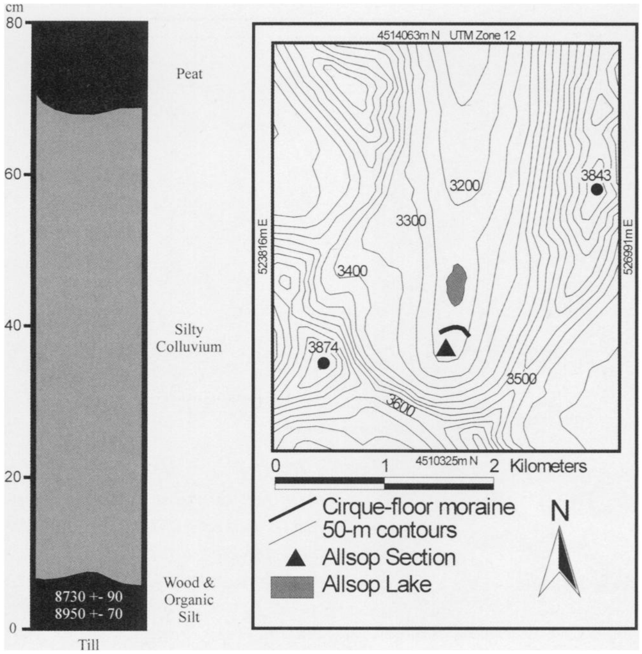

FIGURE 4. Location and stratigraphy of the Allsop Lake site. See Figure 1 for the location of Allsop Lake in the Uinta Mountains. The stratigraphic column for this exposure (scale in cm) is shown at left. The two basal radiocarbon dates are given in ^{14}C yr BP.

within the core are stratigraphically consistent, although sedimentation rates appear to have varied (Table 1, Fig. 2). No evidence of slumping or other disturbance was noted in the core stratigraphy, which consisted of 122 cm of organic silt over 88 cm of finely laminated silty clay.

At the Henrys Fork site (Fig. 3) a peat layer containing fragments of *Salix* wood is exposed directly on top of glacial diamicton in a 3.5-m-high stream-cut bank. The diamicton presumably relates to a cirque-floor moraine ~300 m to the north (down-glacier), which impounded the basin before stream incision successfully breached the moraine. No other glacial deposits are present higher in the exposure, which is dominated by fluvial and wetland sediments recording lateral migration of the Henrys Fork during basin filling. Two samples of *Salix* from the basal peat were dated to provide a limiting age on the formation of the cirque-floor moraine. One sample (HF98–1–26) returned an age of 9370 ± 70 ^{14}C yr BP while the other (HF99–01) returned an age of 8570 ± 80 ^{14}C yr BP (Table 1).

At the Allsop Lake site (Fig. 4) a basin is impounded behind a cirque-floor moraine at an elevation of 3220 m a.s.l. The total thickness of colluvium is ~80 cm, and a layer of organic-rich silt containing wood fragments is present at the base of the fill, directly above glacial diamicton. A wood fragment (AL99–01) and the organic silt containing the wood (AL99–02) were dated to provide a limiting age on the cirque-floor moraine. The wood returned an age of 8950 ± 70 ^{14}C yr BP, while the silt dated to 8730 ± 90 ^{14}C yr BP (Table 1).

LICHENOMETRY

Figure 5 is a vertical air photo presenting the moraine complex at the Deadhorse Lake site and the lichen sample sites (numbered). No radiocarbon-datable material was found in association with the moraines, which range from 3 to 15 m in height, and are composed of quartzite blocks up to 5 m in length (Fig. 6). Soil cover is nonexistent except on the lowest, outermost moraine. Mapping of the moraines identified crosscutting relationships that allowed for establishment of a relative age chronology. Five separate moraines were sampled, although lower elevation moraines are present at the western end of Deadhorse Lake, and visible within the lake as curvilinear shoals. In addition, a linear deposit produced by a rock avalanche was sampled (Site 8). This feature crosscuts one of the moraines, and is in turn overrun by a younger moraine (see Fig. 5).

The lichen diameters were combined with the crosscutting relationships noted in the field mapping to produce a chronology for the eight deposits (Table 2). The largest lichens in the study area were found at Site 1, an area of striated and polished bedrock ledges in front of the outermost moraine (Fig. 5). The significantly larger lichen diameters at Site 1 indicate that these ledges have been exposed longer than the quartzite blocks in the moraines, assuming equivalent growing conditions. Although these ledges were probably covered by a thin layer of till after

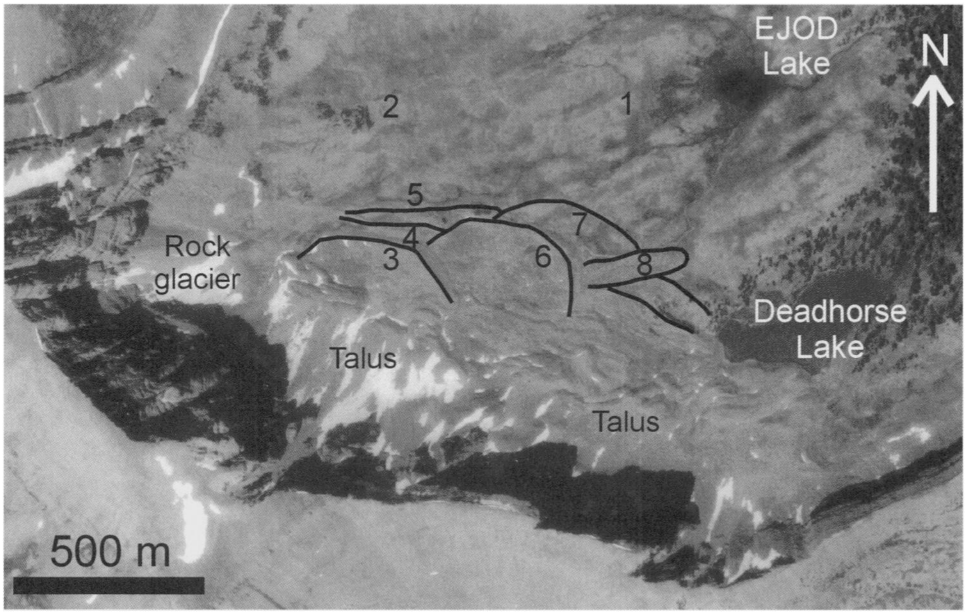

FIGURE 5. Vertical aerial photo of the Deadhorse Lake cirque, showing major moraine loops (delineated), lichen sampling sites (numbered), and the major rock glacier/snowbank system discussed in the text (labeled). See Figure 1 for the location of Deadhorse Lake within the Uinta Mountains.

FIGURE 6. Distal slope of terminal moraine west of Deadhorse Lake. Lichen sample Site 6 is located at the top of this moraine, which is 13 m high (note figure for scale.) For location see Figure 5.

deglaciation, the removal of this material, likely facilitated by the small stream entering EJOD Lake (Fig. 5), was not a recent occurrence.

The outer and inner western moraines (Sites 5 and 4, respectively) follow next in the relative age sequence, indicating that glaciation occurred in the western sector of the cirque before the formation of the outermost eastern moraine (Site 7). The moraine at Site 7, which has a smaller maximum lichen diameter than Site 4, is crosscut by the rock avalanche deposit (Site 8). However, the maximum lichen diameter measured on the rock avalanche (Site 8) was identical to that measured at Site 7, indicating that the rock avalanche occurred soon after deposition of the outer moraine (Site 7). The rock avalanche is clearly overrun by the inner moraine at Site 6, indicating that this glacial advance occurred after the rock avalanche. Lichen diameters show that the second set of exposed ledges (Site 2) is younger than all but Site 3. These ledges were almost certainly striated at the same time as those at Site 1, but presumably remained debris-covered until sometime after the deposition of the older moraines. The thickness of unconsolidated sediment overlying these ledges may have initially been greater, or the rate of overburden removal (likely related to water availability) may have been slower at this higher elevation site.

Finally, the geomorphic position and measured lichens of the innermost moraine (Site 3) show that it is the most recent feature studied. Rock glaciers and talus along the base of the headwall inside the innermost moraines must be even younger, however lichens were not measured on surfaces inside the moraines at Sites 3 and 6.

Three growth-rate curves derived for *Rhizocarpon geographicum* in the Colorado Front Range (Benedict, 1967, 1993) and Wind River Range (Dahms, 2002) were applied to the lichens from the Deadhorse Lake moraines. Although the equations were not developed specifically for the Uintas, *Rhizocarpon geographicum* growth rates are relatively constant between alpine areas (Locke et al., 1979). Therefore, it is assumed that these equations are sufficient to estimate the age of the lichens given the similar climate and geographic proximity of the two ranges. Benedict (1967) determined a growth rate of 14 mm during the first 100 yr after colonization, followed by a slower growth rate of 3.3 mm/century for the remainder of the lichen lifespan. Application of this relationship to the Deadhorse lichens yields ages from

TABLE 2

Site relative age and absolute age from largest Rhizocarpon

Site No.	Relative order	Diameter (mm)	Age[a] (yr)	Age[b] (yr)	Age[c] (yr)
3	7	56	1373	1299	2545
2	6	58	1433	1364	2636
6	5	68 (59)	1736 (1463)	1691 (1397)	3090 (2682)
8	4	60	1494	1430	2727
7	4	60	1494	1430	2727
4	3	63	1585	1528	2864
5	2	65	1645	1593	2955
1	1	85	2252	2246	3864

[a] 14 mm yr^{-1} for first 100 yr, followed by 3.3 mm/century (Benedict, 1967).

[b] $y = 32.6636x - 504$ for x in mm and lichens >20 mm (Benedict, 1993).

[c] 0.022 mm yr^{-1} (Dahms, 2002).

1370 to 2250 yr (Table 2). Benedict (1993) slightly modified this growth rate for lichens >20 mm in diameter, yielding the equation $y = 32.6636x - 540$, where x is lichen diameter in mm and y is lichen age. Application of this equation yields ages from 1300 to 2250 yr. Dahms (2002) determined a growth rate of 2.2 mm/century for *Rhizocarpon geographicum* in the Stough Creek Basin in the southern Wind River Range. This growth rate produces ages of 2500 to 3900 yr (Table 2).

Discussion

TIMING OF DEGLACIATION

The basal AMS date from Hacking Lake indicates that this cirque was entirely deglaciated sometime before 15 to 14 ka BP. For comparison, the Lowder Creek Glacier on the Markagunt Plateau in southern Utah had disappeared by 16.4 to 14.3 ka BP (Anderson et al., 1999). Retreat of valley glaciers in the Colorado Front Range was complete by 15 to 12 ka BP as indicated by sediments from Glacial Lake Devlin (Madole, 1986). Limiting minimum ages from 11.2 ± 180 to 13.7 ± 110 ka BP on basal organics from other locations within the Front Range, and a date of 13.8 ± 810 ka BP from the base of a kettle on the western slope of the Colorado Rockies, support this interpretation (Madole, 1976; Madole, 1980; Nichols et al., 1984; Madole, 1986).

In Wyoming, bog-bottom radiocarbon dates of 6100 ± 100 and 9300 ± 80 ^{14}C yr BP provide a crude limit on the time of formation of the type-Pinedale end moraines (Sorenson, 1987). Surface exposure dating using cosmogenic isotopes better constrains these moraines to between 21,700 ± 700 and 15,800 ± 500 ^{10}Be yr BP, or 23,000 to 16,000 ^{36}Cl yr BP (Gosse et al., 1995a; Phillips et al., 1997). Although there is uncertainty in the cosmogenic nuclide production rates used in these calculations, these results support the consensus opinion that the Pinedale maximum was reached between 23 and 18.5 ka BP (Porter et al., 1983).

The Hacking Lake basal age indicates that retreat of glaciers in the northern Uintas was roughly simultaneous with deglaciation in the Colorado Rockies and Wind River Range. However, radiocarbon dates on aquatic organics can be biased by old carbon effects even in areas of noncalcareous bedrock like the Uintas (e.g., Davis et al., 1998). Therefore, this basal date must be viewed as a maximum estimate on the time of deglaciation in the Hacking Lake cirque. More limiting radiocarbon dates augmented by surface exposure dates are needed to determine exactly how the timing of deglaciation in the northern Uintas related to the deglaciation in neighboring mountain ranges.

AGE OF CIRQUE-FLOOR MORAINES

In the Wind River Range, work by Zielinski and Davis (1987), Gosse et al. (1995b), and Davis et al. (1998) has demonstrated that the Temple Lake and Titcomb Basin moraines, which are located in positions similar to the cirque-floor moraines in the northern Uintas, were deposited during the Younger Dryas subchron, 12.8 ± 0.2 to 11.5 ± 0.3 ka BP. Lacustrine sediments and fossil pollen from the Colorado Front Range also indicate renewed cirque glacier activity and treeline depression during the Younger Dryas (Menounos and Reasoner, 1997; Reasoner and Jodry, 2000). Figure 7 plots the one (black) and two (outline) sigma calibrated age ranges of the five radiocarbon dates that limit the cirque-floor moraines in the northern Uintas. The Younger Dryas subchron and the pre-Younger Dryas Recess Peak advance in the Sierra Nevada are also plotted. The basal date from Hacking Lake (HL) indicates that that cirque was deglaciated prior to 15 to 14 ka BP. The radiocarbon dates from the Henrys Fork (HF) and Allsop Lake (AL) sites suggest that cirque-floor moraines in these basins were deposited in the latest Pleistocene, sometime before ~10 ka BP. Because the dated organics were deposited directly on glacial diamicton with no evidence for the passage of considerable time between deglaciation and colonization by vegetation, it is assumed that these dates provide a limiting minimum age on the formation of the moraines. The duration of the glacial episode that produced these moraines cannot be determined, but the cirque-floor moraines may represent the final pulse of a latest Pleistocene glacial episode that began during the Younger Dryas and persisted until shortly before 10 ka BP. Alternatively, glaciers may have developed in these cirques prior to 10 ka BP after an ice-free Younger Dryas period. Additional dates are required to test these preliminary interpretations and determine the relationship between the cirque-floor moraines of the northern Uintas and Younger Dryas advances in other ranges of the western U.S.

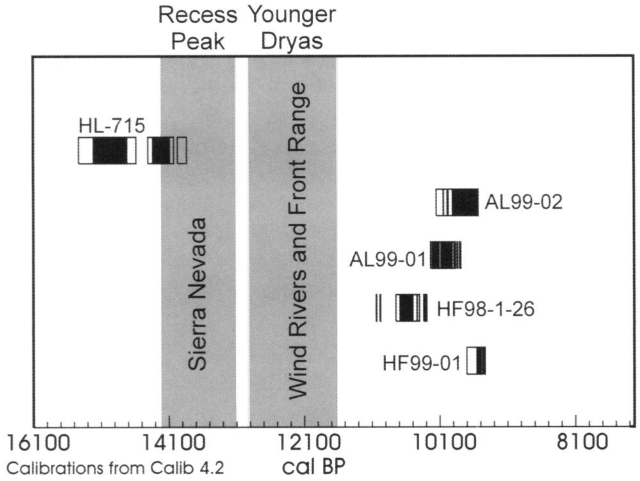

FIGURE 7. Calibrated radiocarbon ages from the northern Uinta cirques compared with post-LGM glaciation in the Sierra Nevada and Wind River Range.

AGE OF THE DEADHORSE LAKE MORAINES

The age of the study sites determined from the lichen growth equations matches the relative order inferred from the field relations for all sites except Site 6 (Table 2). This moraine, determined to be the second oldest from field relationships, has an anomalously old estimated age. This result is due to the largest measured lichen at this site, 68 mm, which is larger than all but four of the 34 *Rhizocarpon* thalli sampled at Site 1. It is possible that this lichen represents an inherited individual that was growing on the exposed bedrock of the cirque headwall before the block it colonized detached and fell onto the glacier surface. Although there is only a 16% (1 in 6) chance of a cubic block coming to rest with its originally exposed side facing upwards, this scenario remains a possibility.

This explanation is illustrated by Figure 8, which plots the measured lichens ($n = 34$) at Site 6 partitioned into 5-mm size classes. The 68-mm-diameter lichen is clearly an outlier to the sample set because no other lichens at this site were measured in the 60- to 64-mm class. If it is assumed that the 68-mm lichen was inherited from the headwall, then the first lichen that started growing after formation of the moraine is only 59 mm in di-

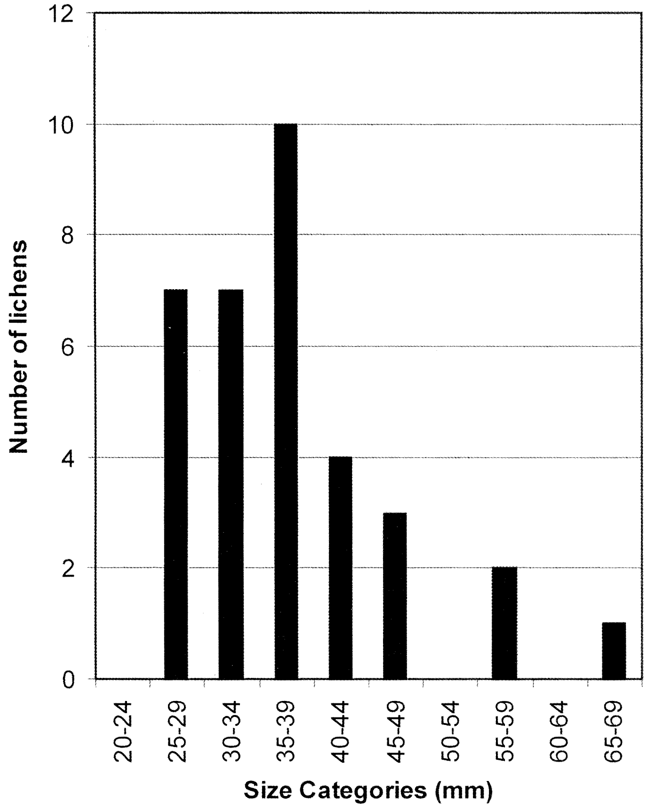

FIGURE 8. Frequency of Rhizocarpon geographicum *thalli diameters measured at Site 6.*

ameter. This size yields an age of 1463 yr through the lichen growth-rate curve of Benedict (1967), 1397 yr from Benedict (1993), and 2682 from Dahms (2002). These results agree better with the observed field relationships (Table 2).

It is possible that the lichen growth-rate curves derived for the Colorado Front Range and Wind Rivers are not applicable to the north slope of the Uintas despite the similar alpine climate of the ranges. Benedict (1967, 1993) and Dahms (2002) measured lichens growing on Precambrian igneous and metamorphic rocks, which may yield different rates of lichen growth than the quartzite in the Uintas. To test the possibility that the moraines at Deadhorse Lake are in fact Little Ice Age features, and that the Uinta lichens are simply growing faster than those in Colorado and Wyoming, lichen-growth rates were calculated for the Deadhorse Lake lichens assuming that the moraines are actually between 700 and 200 yr old. Assuming the same constant used in the growth equation of Benedict (1993, $b = 530$), the growth rates yielded by this analysis are from 4.6 to 11.6 mm/century, considerably greater than the 3 mm/century determined by Benedict (1967, 1993) and the 2.2 mm/century determined by Dahms (2002). Eliminating the constant, and assuming linear growth from the time of colonization yields even higher rates of 8 to 42 mm/century. These values are greater than all others for *Rhizocarpon geographicum* for similar time periods (2 to 4 mm/century, $n = 5$) compiled by Webber and Andrews (1973), although Miller (1969) determined a rate of 8 mm/century for the moister North Cascades of Washington.

It is unlikely that lichen growth rates on quartzite in the Uintas exceed rates on crystalline rocks in the Front Range and Wind Rivers to this extent. This assertion is supported by Benedict (1967), who states that rock texture appears to exert a greater control on lichen growth rates than rock chemistry. Because coarse-grained igneous rocks contain more water than finer-grained or glacially polished rocks, lichens should grow faster on coarser substrates. This inference suggests that growth rates on the denser Uinta Mountain quartzite might actually be slower, not faster, than those in the Front Range and Wind Rivers, and that application of the growth-rate curves from these other ranges is more likely to underestimate substrate age in the Uintas.

Although lichenometry is best considered a relative dating tool, and the direct applicability of the *Rhizocarpon geographicum* growth curves for the Colorado Front Range and Wind Rivers to the Uintas is debatable, it is worth noting that the estimated lichen ages are all greater than ~1300 to 2500 yr (Table 2). 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). In the Sierra Nevada, Konrad and Clark (1998) used a lichen growth-rate equation for *Rhizocarpon superficiale* to argue that lichens more than 700 yr old began growing before the Matthes advance, and that younger lichens represent Little Ice Age deposits. From this logic, the largest lichens on the Deadhorse Lake moraines all began growing before the Little Ice Age.

The moraines at Deadhorse Lake are considerably higher (~3333 m) and closer to the headwall (<1 km) than the cirque-floor moraines in the upper Henrys Fork and at Allsop Lake (~3240 and 1 to 4 km). Although lower moraines at the west end of Deadhorse Lake, and moraine loops submerged beneath the modern water level, may correlate with cirque-floor moraines in other basins, the higher Deadhorse Lake moraines must postdate formation of these features. Therefore, the Deadhorse Lake moraines represent a limited pulse of cirque glaciation at the highest elevations of the Uintas sometime after the cirque-floor moraines formed in the latest Pleistocene, but before the Little Ice Age.

THE LITTLE ICE AGE IN THE UINTAS

If the cirque-floor moraines are latest Pleistocene features and the Deadhorse Lake moraines predate the Little Ice Age, then what is the geomorphic expression of the Little Ice Age climate deterioration known to have occurred in neighboring ranges and assumed to have occurred in the Uintas? The likely answer to this question is that the numerous rock glaciers, which are ubiquitous at higher elevations (>3000 m a.s.l.) of the northern Uintas, formed during the Little Ice Age when a slight deterioration of the climate increased both the supply of frost-shattered rock and the potential for perennial ice to exist in talus interstices. Many of these features appear active today, or were likely active in the recent past, on the basis of characteristics suggestive of motion including unvegetated frontal slopes standing beyond the angle of repose, meandering longitudinal furrows, and transverse furrows (Wahrhaftig and Cox, 1959). Some rock glaciers also have surface meltwater ponds, and springs discharging silt-laden 0°C meltwater in late summer, implying the presence of interstitial ice.

As noted above, numerous rock glaciers and abundant talus present inside the youngest moraines at Deadhorse Lake must postdate the last episode of glacial activity in this cirque. The largest rock glacier is located in the southwestern corner of the cirque, in front of a quasi-perennial snowbank (Fig. 9). The front of this feature is over 30 m high (Fig. 10) and 0°C water was recorded emerging from the frontal slope in early August 1999. The snowbank behind the rock glacier extends up the headwall

FIGURE 9. Rock glacier and snowfield complex west of Deadhorse Lake, 7 August, 1999.

for ~100 m in late summer, and is several meters thick at its upper edge, where it separates from the headwall along a furrow eroded by meltwater dripping from cornices above. Multiple linear cracks suggestive of crevasses, but likely due to snow creep instead of ice flow, are visible near the top of the snowbank in late summer. Excavations into the snowbank in August 1999 revealed several decimeters of winter 1998–99 snow overlying 1 to 5 m of firn representing at least the penultimate winter's snowfall.

The persistence of this snowbank, its volume, and the presence of firn and linear tension cracks indicate that this accumulation of snow is quite close to forming the nucleus of a cirque glacier. Indeed, the snowbank and rock glacier complex is quite similar to that present on the north side of Wheeler Peak in Nevada, considered to be the only modern glacier in the Great Basin (Heald, 1956; Osborn, 1988; Osborn and Bevis, 2001). The Deadhorse Lake snowbank and rock glacier complex likely represents a lingering example of a periglacial system that was common at higher elevations throughout the northern Uintas during the Little Ice Age. Most of these features are now represented by dormant and collapsed rock glaciers because warming has resulted in loss of interstitial ice and the snowbank meltwater sources from all but the most favorable locations.

PALEOCLIMATE IMPLICATIONS

These results have implications for our understanding of Neoglacial climate changes across the mountains of the western U.S., and the geomorphic responses to these changes. Seltzer (1994) determined the sensitivity of a glacier equilibrium-line altitude (ELA) to changes in various climatic variables, including mean summer temperature (95 m °C^{-1}) and winter snow

FIGURE 10. View of the 30-m high steep frontal slope of the rock glacier shown in Figure 9.

accumulation (-0.3 m mm^{-1}). Seltzer's equations suggest that, if changes in solar insolation were negligible, an ELA drop of 300 to 500 m (from the modern ELA to the upper cirque-floors) could have been driven by a decrease of mean summer temperature of 3.2 to 5.3°C if precipitation remained unchanged, or by an increase of winter snow accumulation of 1000 to 1667 mm snow water equivalent (SWE) if summer temperatures remained unchanged. These estimates are extreme values and temperature depression would certainly have been accompanied by increased winter precipitation because temperature depression would convert late-spring and early-fall rain to snow (Leonard, 1989).

Snow redistribution into cirques can total 4 to 8 times the actual winter snowfall (Outcalt, 1965), allowing small glaciers to exist as rejuvenated or drift glaciers below the local ELA in areas of excessive snow drifting and avalanching (Hambrey, 1994; Larson, 1996). Therefore, relatively modest increases in snow accumulation of ~250 to 400 mm SWE per winter (roughly twice the modern winter snow accumulation at Deadhorse Lake) could result in an additional ~1000 to 2000 mm of drifted accumulation within the highest cirques (Munroe, 2000; NRCS, 2000). Thus, enhancement of winter precipitation by ~250 to 400 mm SWE, accompanied by depression of the mean summer temperature 0.5 to 2°C, would likely be sufficient to generate and sustain small cirque glaciers in the most favorable locations.

Such changes are not as extreme as the end-member values yielded by assuming that one variable (precipitation or temperature) remained constant, but they are still in excess of changes that accompanied the Little Ice Age in neighboring mountain ranges. For example, Zielinski (1989) estimated that a temperature depression of 0.2°C and a 2% increase in snow accumulation accompanied the Little Ice Age in the Wind River Range. In the Sierra Nevada, Graumlich (1993) used tree-ring data from subalpine conifers to infer that summer temperatures were approximately 0.5°C lower than modern from A.D. 1450 to 1850. These comparisons suggest that the magnitude of climate deterioration during the Little Ice Age was insufficient to generate cirque glaciers in the Uintas.

REGIONAL CORRELATIONS

Depending on which growth-rate equation is used, the lichen data suggest that the Deadhorse Lake moraines were formed approximately 1300 to 3000 yr BP. This interval is roughly synchronous with deposition of the Black Joe Alloformation in the Wind River Range, a pre-Gannett Peak (Little Ice Age) unit defined by Dahms and Birkeland (2000) and Dahms (2002) on the basis of numerous relative age criteria. This correlation must be considered preliminary because there is no lichen growth-rate curve specific to the northern Uintas. But this relationship still suggests that the conditions responsible for positive glacier mass balance in the middle Neoglacial were regional in extent. Therefore, something must have changed to drive the disappearance (or extreme spatial diminution) of the Uinta glaciers during the Little Ice Age, a time when glaciers expanded in the Wind Rivers, Colorado Rockies, and Sierra Nevada.

Small glaciers are present in these other ranges today, indicating that some part of the modern climate equation, probably winter precipitation, still favors positive glacier mass balance. In contrast, conditions in the Uintas are likely too dry in the winter due to orographic shielding by upwind mountain ranges. These ranges include the Wasatch Mountains, which have the steepest modern precipitation gradient in the Great Basin (1.37 m km^{-1}), demonstrating highly effective interception of westerly-derived moisture (Zielinski and McCoy, 1987).

During the middle Neoglacial period, when glaciers existed at Deadhorse Lake, winter moisture may have reached the Uintas by a slightly different path than the modern (and Little Ice Age) storm systems, which come from the west along a major zonal boundary known as the "winter boundary" (Mitchell, 1976). The winter boundary forms annually along the northern border of Utah, and separates an air mass over the northern Rocky Mountains dominated by Pacific air, from a southerly air mass over the Great Basin dominated by drier southwesterly air flow. A shift in the storm track from westerly to more northwesterly would have allowed more moisture to bypass the Wasatch Front and reach the high-elevation cirques of the northern Uintas, facilitating glacier formation at Deadhorse Lake. Positive glacier mass balance can also result from cooler summer temperatures, but it is unlikely that middle Neoglacial cirque glacier formation was driven solely by regional temperature depression in excess of Little Ice Age values. Such a change would have produced simultaneous middle Neoglacial glacier expansion in the Wind Rivers and Sierra Nevada, yet there is no evidence of post-Recess Peak/pre-Matthes glaciers in the Sierra Nevada, and glaciers of Gannet Peak-age at least locally overran older deposits in the Wind Rivers (Burke and Birkeland, 1983). This theory should be tested by additional reconstructions of late Holocene paleoclimate in the Rocky Mountains. Effects of a northwesterly to westerly shift in storm track during the late Holocene might be recorded in the western Wasatch Range where winter precipitation would have increased as the Uintas became drier. Nonetheless, the record from Deadhorse Lake underscores the sensitivity of marginal glacier systems to minor climatic shifts, and provides an upper limit on the magnitude of precipitation and temperature changes that accompanied the Little Ice Age.

Conclusion

Field evidence and radiocarbon dating indicate that glaciers were present, but extremely limited in extent, in the northern Uintas during the Holocene. An AMS date of 12,190 ± 120 ^{14}C yr BP on basal organic sediment from a tarn, combined with four stratigraphically consistent dates from higher in the core, indicates that deglaciation was locally complete before ~15 to 14 ka BP, and that this cirque has been ice-free since the latest Pleistocene. Additional radiocarbon dates on organic sediments retrieved from the bases of two colluvial basins, directly atop glacial diamicton and behind cirque-floor moraines, cluster ~10 ka BP indicating that these moraines date to the latest Pleistocene and are not Neoglacial features.

There are higher elevation sites that appear to be preferable loci for Neoglacial cirque glacier rejuvenation, but materials suitable for radiocarbon dating have not been recovered from them. At one of these sites, measurements of lichen diameters indicate multiple advances of small glaciers at some point after the last deglaciation. However, the maximum diameters of *Rhizocarpon geographicum* suggest colonization approximately 1300 to 3000 yr BP, more than twice the age of Little Ice Age deposits in the Wind River Range and Sierra Nevada. The convergence of evidence indicates that Neoglacial ice was extremely limited in extent and restricted to the middle Neoglacial period; cirques remained ice free during the Little Ice Age. This conclusion contradicts some previous interpretations of the upper subalpine geomorphology of the Uintas and provides additional information about the magnitude of Holocene climate changes in the western U.S.

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References Cited

Anderson, L. W. and Anderson, D. S., 1981: Weathering rinds on quartzarenite clasts as a relative-age indicator and the glacial chronology of Mount Timpanogos, Wasatch Range, Utah. *Arctic and Alpine Research,* 13: 25–31.

Anderson, R. S., Hasbargen, J., Koehler, P. A., and Feiler, E. J., 1999: Late Wisconsin and Holocene subalpine forests of the Markagunt Plateau of Utah, southwestern Colorado Plateau, U.S.A. *Arctic, Antarctic, and Alpine Research,* 31: 366–378.

Antevs, E., 1948: The Great Basin, with emphasis on glacial and post-glacial times: climate changes and pre-white man. *Bulletin of the University of Utah,* 38: 168–191.

Barnhardt, M. L., 1973: Late Quaternary geomorphology of the Bald Mountain area, Uinta Mountains, Utah. MS thesis, University of Utah at Salt Lake City. 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,* 6(48): 817–832.

Benedict, J. B., 1973: Chronology of cirque glaciation, Front Range. *Quaternary Research,* 3: 584–599.

Benedict, J. B., 1985: Arapaho Pass: Glacial geology and archeology at the crest of the Colorado Front Range. *Research Report No. 3,* Center for Mountain Archeology, Ward, Colorado. 197 pp.

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

Beschel, R. E., 1961: Dating rock surface by lichen growth and its application to glaciology and physiography (lichenometry). *In* Raasch, G. O. (ed.), *Geology of the Arctic.* Vol. 2. Toronto: University of Toronto Press, 1044–1062.

Birkeland, Crandell and Richmond, 1971: Status of correlation of Quaternary stratigraphic units in the western conterminous United States. *Quaternary Research,* 1: 208–227.

Birman, H. J., 1964: Glacial geology across the crest of the Sierra Nevada, California. *Geological Society of America Special Paper,* 75. 80 pp.

Bull, W. B., King, J., Kong, F., Mountoux, T., and Phillips, W. M., 1994: Lichen dating of coseismic landslide hazards in alpine mountains. *Geomorphology,* 10: 253–264.

Bull, W. B. and Brandon, M. T., 1998: Lichen Dating of Earthquake-Generated Regional Rockfall Events, Southern Alps, New Zealand. *Geological Society of America Bulletin,* 110: 60–84.

Burke, R. M. and Birkeland, P. W., 1983: Holocene glaciation in the mountain ranges of the western United States. *The Holocene,* 2: 3–11.

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,* 38/39: 21–38.

Curry, R. R., 1968: Quaternary climate and glacial history of the Sierra Nevada, California. Ph.D. dissertation, University of California at Berkeley. 238 pp.

Curry, R. R., 1969: Holocene climatic and glacial history of the central Sierra Nevada, California. *In* Schumm, S. A. and Bradley, W. C. (eds.), *United States Contributions to Quaternary Research. Geological Society of America Special Paper,* 123: 1–47.

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. *Geological Society of America Abstracts with Programs,* 32: 60.

Dahms, D. E., 2002: Glacial stratigraphy of Stough Creek Basin, Wind River Range, Wyoming. *Geomorphology,* 42: 59–84.

Davis, P. T., 1988: Holocene glacier fluctuations in the American cordillera. *Quaternary Science Reviews,* 7: 129–157.

Davis, P. T., 2000: Five decades of alpine and glacial research in the late Pleistocene and Holocene: Type examples from Wyoming and Colorado, U.S.A. *Geological Society of America, Abstracts with Programs,* 32: 60.

Davis, P. T., Gosse, J., Romito, M, Sorenson, C. C., Klein, J. U., Dahms, D., Zielinski, G., and Jull, J. T., 1998: Younger Dryas age for type Titcomb Basin and type Temple Lake moraines, Wind River Range, Wyoming, U.S.A. *Geological Society of America Abstracts with Programs,* 30: 66.

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

Gosse, J. C., Evenson, E. B., Klein, J., Lawn, B., and Middleton, R., 1995b: Precise cosmogenic ^{10}Be measurements in western North America; support for a global Younger Dryas cooling event. *Geology,* 23: 877–880.

Graumlich, L. J., 1993: A 1000-year record of temperature and precipitation in the Sierra Nevada. *Quaternary Research,* 39: 249–255.

Grove, J. M., 1990: *The Little Ice Age.* New York: Routledge. 498 pp.

Grogger, P. K., 1974: Glaciation of the High Uintas Primitive Area, Utah, with emphasis on the northern slope. Ph.D. dissertation, University of Utah at Salt Lake City. 210 pp.

Grogger, P. K. 1975: Neoglaciation of the northern slope of the High Uintas Primitive Area, Utah. *Great Plains Rocky Mountain Geographical Journal,* 4: 37–43.

Hambrey, M., 1994: *Glacial Environments.* Vancouver: UBC Press. 296 pp.

Heald, W. F., 1956: An active glacier in Nevada. *American Alpine Journal,* 10: 164–167.

Konrad, S. K. and Clark, D. H., 1998: Evidence for an Early Neoglacial glacier advance from rock glaciers and lake sediments in the Sierra Nevada, California, U.S.A. *Arctic and Alpine Research,* 30: 272–284.

Lamb, H. H., 1977: *Climate: Present, Past and Future. Vol 2.* London: Methuen. 835 pp.

Larson, P. R., 1996: Glacial geomorphology of a portion of the Wasatch Plateau, central Utah. Ph.D. dissertation, University of Utah at Salt Lake City. 186 pp.

Leonard, E. M., 1989: Climatic change in the Colorado Rocky Mountains: estimates based on modern climate at late Pleistocene equilibrium-lines. *Arctic and Alpine Research,* 21: 245–255.

Livingstone, D. A., 1955: A lightweight piston sampler for lake deposits. *Ecology,* 36: 137–139.

Locke, W. W., Andrews, J. T. and Webber, P. J., 1979: A manual for lichenometry. *British Geomorphological Research Group, Technical Bulletin,* 26. 47 pp.

Madole, R. F., 1976: Bog stratigraphy, radiocarbon dates, and Pinedale to Holocene glacial history in the Front Range, Colorado. *U.S. Geological Survey Journal of Research,* 4: 163–169.

Madole, R. F., 1980: Time of Pinedale deglaciation in north-central Colorado: Further considerations. *Geology,* 8: 118–122.

Madole, R. F., 1986: Lake Devlin and Pinedale glacial history, Front Range, Colorado. *Quaternary Research,* 25: 43–54.

Madsen, D. B. and Currey, D. R., 1979: Late Quaternary glacial and vegetation changes, Little Cottonwood Canyon area, Wasatch Mountains, Utah. *Quaternary Research,* 12: 254–270.

Matthes, F. E., 1939: Report of the committee on glaciers. *American Geophysical Union Transactions,* 518–523.

Menounos, B. and Reasoner, M. A., 1997: Evidence for cirque glaciation in the Colorado Front Range during the Younger Dryas chronozone. *Quaternary Research,* 48: 38–47.

Miller, C. D., 1969: Chronology of Neoglacial moraines in the Dome Peak area, North Cascade Range, Washington. *Arctic and Alpine Research,* 1: 49–66.

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

Moss, J. H., 1951: Late glacial advances in the southern Wind River Mountains, Wyoming. *American Journal of Science,* 249: 865–883.

Munroe, J. S., 2000: Unpublished analysis of SNOTEL data for the northern Uinta Mountains.

Nichols, H., Short, S. K., Elias, S. and Harbor, J., 1984: Holocene sedimentation, palynology, and paleoecology of alpine and subalpine lakes and peat bogs, Colorado Front Range. *American Quaternary Association, Eighth Biennial Meeting,* Guidebook for Field Trip 7, 21 pp.

NRCS, 2000: National Resource and Conservation Service National Water and Climate Center SNOTEL Database: URL: ⟨www.wcc.nrcs.usda.gov/water/w_data.html⟩.

Osborn, G., 1988: Minimum age of the Wheeler Peak rock glacier, Great Basin National Park, Nevada. *American Quaternary Association, Tenth Biennial Meeting, Abstracts with Programs,* 10: 144.

Osborn, G. and Bevis, K., 2001: Glaciation in the Great Basin of the Western United States. *Quaternary Science Reviews,* 20: 1377–1410.

Outcalt, S. I., 1965: Analysis of the surface, derived by plotting annual height changes on a drift glacier as a function of time and elevation. *Geological Society of America Special Paper,* 82: 372–373.

Phillips, F. M., Zreda, M. G., Gosse, J. C., Klein, J., Evenson, E. B., Hall, R. D., Chadwick, O. A., and Sharma P., 1997: Cosmogenic ^{36}Cl and ^{10}Be ages of Quaternary glacial and fluvial deposits of the Wind River Range, Wyoming. *Geological Society of America Bulletin,* 109: 1453–1463.

Porter, S. C. and Denton, G. H., 1967: Chronology of Neoglaciation in the North American Cordillera. *American Journal of Science,* 265: 177–210.

Porter, S. C., Pierce, K. L., and Hamilton, T. D., 1983: Late Wisconsin mountain glaciation in the western United States. *In* Wright, H. E. (ed.), *Late-Quaternary Environments of the United States.* Vol. 1. Minneapolis: University of Minnesota Press, 71–111.

Reasoner, M. A. and Jodry, M. A., 2000: Rapid response of alpine timberline vegetation to the Younger Dryas climate oscillation in the Colorado Rocky Mountains, USA. *Geology,* 28: 51–54.

Schoenfeld, M. J., 1969: Quaternary geology of the Burnt Fork area, Uinta Mountains, Summit County, Utah. MS thesis, University of Wyoming at Laramie. 75 pp.

Seltzer, G. O., 1994: Climatic interpretation of alpine snowline variations on millennial time scales. *Quaternary Research,* 41: 154–159.

Sorenson, C. J., 1987: Soils map of the Fremont Lake South Quadrangle, Sublette County, Wyoming. *U.S. Geological Survey Map I-1800,* scale 1:24,000.

Stuiver, M., Reimer, P. J., Bard, E., Beck, W. E., Burr, G. S., Hughen, K. A., Kromer, B., McCormac, F. G., v.d. Plicht, J., and Spurk, M., 1998: INTCAL98 radiocarbon age calibration 0–24,000 BP. *Radiocarbon,* 40: 1041–1083.

Stuiver, M., Reimer, P. J., and Reimer, R. W., 2000, CALIB 4.3: [WWW program and documentation] URL:⟨http://www.calib.org⟩.

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

Webber, P. J. and Andrews, J. T., 1973: Lichenometry: A commentary. *Arctic and Alpine Research,* 5: 295–302.

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

Zielinski, G. A., 1989: Inferred climatic conditions based on equilibrium-line altitudes of former glaciers at type localities in the Wyoming Wind River Range and Colorado Front Range. *Geological Society of America Abstracts with Programs,* 21: 211.

Zielinski, G. A. and Davis, P. T., 1987: Late Pleistocene age of the type Temple Lake moraine, Wind River Range, Wyoming, U.S.A. *Géographie physique et Quaternaire,* XLI: 397–401.

Zielinski, G. A. and McCoy, W. D., 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,* 19: 127–134.

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