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Authors: Refsnider, Kurt A., and Brugger, Keith A.

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Rock Glaciers in Central Colorado, U.S.A., as Indicators of Holocene Climate Change

Kurt A. Refsnider† and Keith A. Brugger**

*Geology Discipline, University of Minnesota, Morris, 600 East 4th Street, Morris, Minnesota 56267, U.S.A.
†Present address: Institute of Arctic and Alpine Research *and* Department of Geological Sciences, University of Colorado, 1560 30th Street, Boulder, Colorado 80309, U.S.A.
Kurt.Refsnider@colorado.edu

Abstract

We measured thalli diameters of the lichen *Rhizocarpon* subgenus *Rhizocarpon* on 48 individual lobes of 18 rock glaciers and rock glacier complexes in the Elk Mountains and Sawatch Range of central Colorado. Cumulative probability distribution and K-means clustering analyses were used to separate lichen thalli measurements into statistically distinct groups, each interpreted as representing a discrete episode of rock glacier activity driven by an interval of cooler climate. Lichen ages for these episodes were assigned using a growth curve developed for *Rhizocarpon geographicum* in the nearby Front Range. An early Neoglacial episode, ca. 3080 yr BP, is correlative to other glacial and periglacial activity in the southern Rocky Mountains and surrounding areas and broadly corresponds to an interval of climatic deterioration evident in several other proxies of Holocene climate. The younger two episodes, ca. 2070 and 1150 yr BP, are also coeval with regional (Audubon) glacial and periglacial activity but are thus far not widely recognized in other climate proxies.

Introduction

Rock glaciers are lobate or tongue-shaped masses of rock and ice that form under cold climates in alpine regions and at high latitudes. Such climates promote the splitting of exposed bedrock in valley and cirque walls by frost action, and in some cases, the accumulation of interstitial ice in talus fields. In many alpine settings, rock glaciers are ubiquitous elements of the landscape and can contribute significantly to its geomorphic evolution by transporting large volumes of debris downslope by creep facilitated by ice deformation (Barsch, 1977; Giardino and Vitek, 1988; Humlum, 2000). Historically, two general models of rock glacier genesis have been proposed. One is a periglacial model (*sensu* Clark et al., 1998) wherein the *in situ* freezing of rain and meltwater occurs within the interstices of pre-existing talus fields (Capps, 1910; Warhhaftig and Cox, 1959; Outcalt and Benedict, 1965). The other is a glacigenic model in which an existing glacier is buried by rock debris. This debris is derived either from cliffs upslope or by the accumulation of debris melting out of the ice, subsequently insulating the underlying ice (Brown, 1925; Outcalt and Benedict, 1965; Potter, 1972). More recent work suggests these two models may represent end members of a continuous spectrum of the processes responsible for rock glacier development (Corte, 1987; Whalley and Martin, 1992; Elconin and LaChapelle, 1997; Brazier et al., 1998; Clark et al., 1998), but this conclusion is not universally accepted (e.g., Haeberli, 1985).

Regardless of their mode of formation, rock glaciers and/or rock glacier activity have been used in varying manners to reconstruct paleoclimates. Previous investigations (e.g., Birkeland, 1973; Miller, 1973; Morris, 1987; Nicholas and Butler, 1996) have used relict rock glaciers to simply establish qualitative chronologies of Holocene climate change because intervals of rock glacier activity often correlate chronostratigraphically with those of Neoglacial glacier expansion, thus broadly implying colder periglacial conditions. Several studies have used relict rock glaciers to derive more quantitative estimates of paleoclimate based on their implications for depression of snow, or equilibrium-line altitudes (ELAs), the lower limit of rock glaciers in relation to zones of continuous and discontinuous permafrost, or based on more specific relationships between modern climate and active rock glaciers (e.g., Kerschner, 1978; Clark et al., 1994; Brazier et al., 1998; Sailer and Kerschner, 1999; Hughes et al., 2003). Rock glaciers do not, however, always show simple and consistent relationships to climatological parameters (Baroni et al., 2004), underscoring the need for a better understanding of the interactions between rock glaciers of a specific genetic origin (i.e., periglacial or glacigenic) and climate, debris supply, and topography (Olyphant, 1987; Kirkbride and Brazier, 1995; Humlum, 1998, 2000; Hughes et al., 2003) before their full potential as a climate proxy is realized. Nevertheless, some valuable insights regarding paleoclimates can be gleaned from rock glacier activity. In this study, we use lichenometry to date intervals of rock glacier activity to document late Holocene climate changes in central Colorado.

Methods

STUDY AREA

The Sawatch Range (Fig. 1) is a fault-bounded block consisting primarily of Precambrian crystalline rocks uplifted during the Late Mesozoic–Early Tertiary Laramide Orogeny (Tweto, 1987). In contrast, the Elk Mountains consist of thrust-faulted Paleozoic and Mesozoic sedimentary sequences. Tertiary intrusive rocks are common in both ranges. High peaks in the study area, many exceeding 4000 m, are separated by deeply incised valleys or intermontane parks. The high relief of the region is the product of Tertiary uplift and stream dissection. Subsequent modification by successive Pleistocene glaciations created the present alpine landscape.

Climate across the study area is generally cool and dry, but temperatures tend to decrease and precipitation increases from

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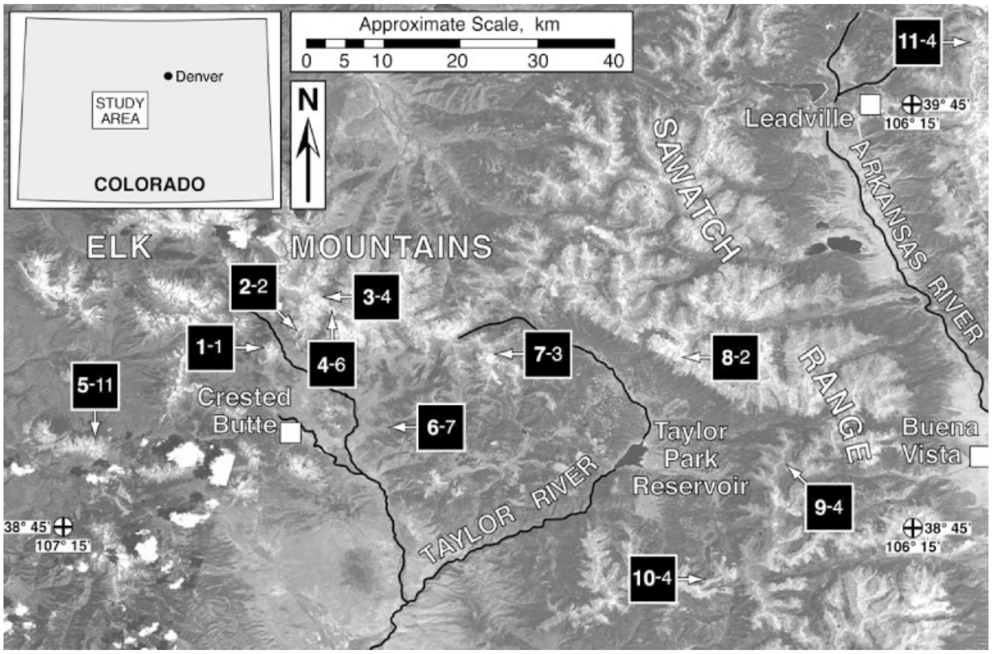

FIGURE 1. LANDSAT 7 (grayscale) image of the study area showing the locations of rock glaciers sampled. The first number corresponds to the following: 1—Gothic Mountain; 2—Virginia Basin; 3—Copper Creek; 4—Queen Basin; 5—East Beckwith Mountain; 6—Ferris Creek; 7—Italian Mountain; 8—Pieplant Creek; 9—Cottonwood Pass; 10—Cumberland Pass; and 11—Mount Democrat. The second number indicates the number of individual lobes upon which lichen measurements were made at each location.

east to west. Differences in elevation notwithstanding, values for mean annual temperature and precipitation at Buena Vista (2417 m; Fig. 1) are 6.2°C and 25.5 cm, whereas those at Crested Butte (2706 m) are 0.6°C and 60.6 cm (NCDC 1971–2000 norms from the Western Regional Climate Center [http://www.wrcc.dri.edu/summary/climsmco.html]).

Within the study area, moraines of two late Pleistocene glaciations are recognized in the Taylor Park area, adjacent parts of the Elk Mountains (Brugger and Goldstein, 1999; Brugger, unpublished), and the upper Arkansas River valley (Fig. 1; Nelson and Shroba, 1998). Cosmogenic ^{10}Be and ^{36}Cl zero-erosion exposure ages from boulders on last glacial maximum (LGM) terminal moraine complexes in the Taylor River valley (Fig. 1) range from 16.3 ± 1.6 to 22.2 ± 2.8 ka (Brugger, 2006) and suggest glacial advances during the LGM in the study area are generally correlative to others in the southern Rocky Mountain region (e.g., Gosse et al., 1995; Benson et al., 2005). ELA depression during the LGM suggests that mean summer temperatures in the region may have been ~7–9°C cooler than present (Brugger and Goldstein, 1999; Brugger, 2006). Late glacial and Holocene climate changes that followed were described by both Fall (1997) and Emslie et al. (2005) and will be discussed subsequently.

ROCK GLACIERS

Rock glaciers are abundant within the study area. Many of these are relict features, or least inactive, as indicated by morphological characteristics including frontal slopes much less than the angle of repose, soil development and extensive vegetative cover, and/or stable and lichen-encrusted boulders. Others may still be active, as indicated by steep and sharp-crested fronts, "bulldozed" turf rolls at the toes of frontal slopes, loose and unstable surface boulders above the toes, and a general lack of vegetation. Activity is also suggested by earlier measurements on two rock glaciers in the Elk Mountains (Bryant, 1971) and ongoing measurements on the East Beckwith rock glacier (Fig. 2A; Brugger, unpublished) where mean velocities are 43, 40, and 7 cm yr^{-1}, respectively.

Morphologies and topographic settings of rock glaciers throughout the region vary. The majority of the rock glaciers we studied are located in cirque basins, and of these, approximately half are situated on slopes extending from the valley side rather than the headwall; those not in cirques are situated below rockwalls. Continuous talus deposits connect all the rock glaciers to the debris source cliffs above. Despite their location, those situated in cirques are mostly of the "valley-wall" type (Outcalt and Benedict, 1965). These rock glaciers are smaller lobate forms, often occurring below avalanche chutes and fed by talus slopes along sidewalls and typically extend onto basin floors but do not fill the entire basin. Brazier at al. (1998) recognized analogous, non-glacigenic forms in cirque basins of the Ben Ohau Range of New Zealand. The exceptions are several rock glaciers in the Beckwith Range and on Italian Mountain (Fig. 1) that are best described as "cirque-floor" types. These larger tongue-shaped forms often fill the entire basin and emanate from talus slopes below cirque headwalls. However, these forms are not moraine-like deposits, nor do they have irregular surfaces associated with the stagnation of inactive, retreating, debris-covered glaciers.

The rock glaciers in the study area include both single- (Figs. 2A, lower left, and 2B) and multiple-lobed forms (Fig. 2C), where lobes are characterized as arcuate, steep-fronted features. Most multi-lobed forms exhibit stratigraphic relationships recording the advance of younger lobes over older ones (Fig. 2C). The tongue-shaped rock glacier on East Beckwith Mountain shown in the center of Figure 2A is the only feature included in this study with prominent concentric lobes with ridge-like morphologies.

Without detailed geophysical studies (e.g., Potter et al., 1998; Ikeda and Matsuoka, 2002), glaciological approaches (e.g., Potter,

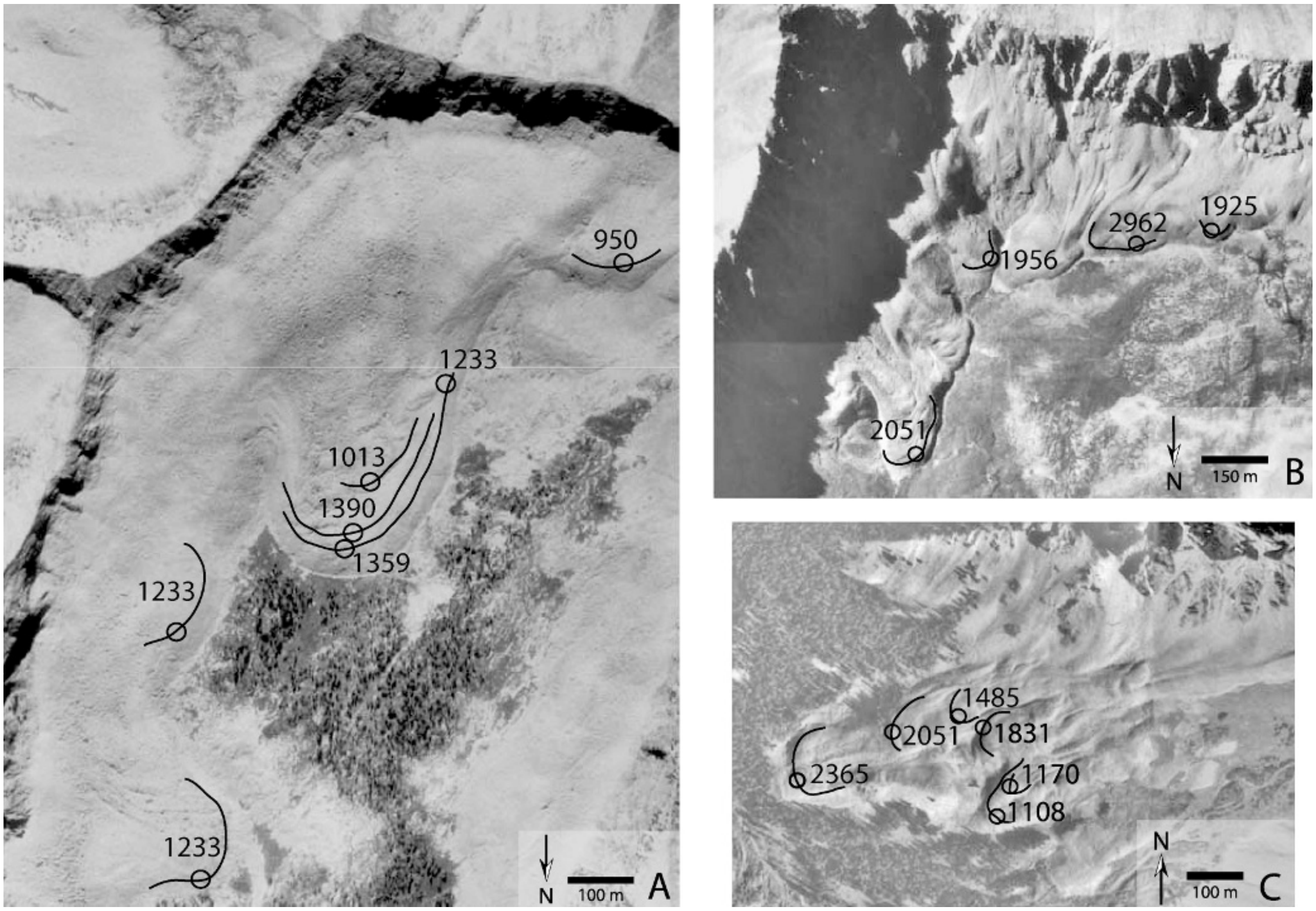

FIGURE 2. Examples of sampling locations (open circles) and lobe (lines) ages on several rock glaciers and rock glacier complexes; numbers indicate the deposit age at each sampling location. Note the general increase in deposit age with increasing distance from the talus source. (A) Rock glaciers in the northeastern cirque of East Beckwith Mountain. (B) A series of rock glaciers in the cirque immediately east of Mount Democrat. (C) The rock glacier complex in Queen Basin.

1972; Konrad and Clark, 1998), coring (Potter, 1972), or direct observations of their internal structure, it is difficult to be certain of the origin of the rock glaciers used in this study. However, we suspect, as have others (e.g., Outcalt and Benedict, 1965; Calkin et al., 1987; Clark et al., 1994, 1998; Kirkbride and Brazier, 1995) that the morphology and topographic setting of the valley–wall rock glaciers (or equivalent forms) reflect a periglacial genesis. Such a presumption is supported by evidence suggesting that many cirque basins were ice-free prior to the rock glacier activity documented in this study (Fall, 1997). No evidence has been found suggesting that glaciers re-formed during Holocene cool intervals. In contrast, the cirque-floor rock glaciers in the Beckwith Range (Fig. 2A, center) may be glacigenic. Within the largest of these rock glaciers, a discrete ice layer under the surface debris was evident in one exposure, and the ridge-like lobe-front morphology is possibly indicative of a glacigenic origin (Clark et al., 1998; Corte, 1987). Given the wetter and cooler climate in this part of the study area, glacial ice could have persisted (or re-formed) here longer than elsewhere in the study area.

SAMPLING

We selected 48 lobes on 18 rock glaciers or rock glacier complexes on the basis of location, elevation, and aspect. These rock glaciers fall within an east-west longitudinal transect (106.15–107.22°W) and range in terminus elevation from 3005 to 3755 m (Table 1). Rock glacier surfaces were dated using lichenometry (Beschel, 1961; Noller and Locke, 2000), as in numerous other studies of Holocene landforms (e.g., Birkeland, 1973; Calkin et al., 1987; Morris, 1987; Nicholas and Butler, 1996; Konrad and Clark, 1998). The lichen *Rhizocarpon* subgenus *Rhizocarpon* (generally synonymous with *Rhizocarpon geographicum* in many lichenometric studies) was used because of its slow and steady growth in alpine and arctic regions, relative abundance at higher elevations on a variety of lithologies, and ease of identification (Noller and Locke, 2000).

We measured thalli diameters of lichens on distinct lobes at various elevations and positions within each rock glacier. Although there is no consensus regarding the most effective sampling strategy for determining the maximum lichen size on rock substrates (Innes, 1984; McCarrol, 1994; Bull and Brandon, 1998; Noller and Locke, 2000), we employed a sampling protocol similar to that used by Innes (1984), because it is most suitable for dating rock glacier deposits. The lichen distribution on each lobe was examined to locate the areas with the largest thalli. Once such areas were found, we established a circular sampling site with a radius of 5 m. We sampled most lobes in two locations, and larger lobes and those with more variable lichen sizes were sampled in three areas. In nearly all cases, sampling sites were located just above the break in slope above the lobe

TABLE 1

Site and lichen data from each rock glacier lobe included in this study.

Location	Latitude (°N)	Longitude (°W)	Elev. (m)	Frontal slope (degrees)	Aspect	Lithology	Thallus size (mm)*	St. Dev. (mm)	Age (yrs BP)
East Beckwith Central	38.8546	107.2238	3333	46	NW	Granitic	53	2	1202
East Beckwith Central	38.8581	107.2211	3195	41	NW	Granitic	48	3	1038
East Beckwith East	38.8444	107.2185	3544	30	N	Granitic	44	3	907
East Beckwith East	38.8448	107.2159	3533	35	NNW	Granitic	46	4	973
East Beckwith East	38.8462	107.2150	3492	37	N	Granitic	58	5	1365
East Beckwith East	38.8475	107.2144	3435	22	N	Granitic	57	5	1332
East Beckwith East	38.8483	107.2142	3414	46	N	Granitic	54	1	1234
East Beckwith East	38.8499	107.2105	3428	30	NW	Granitic	53	4	1202
East Beckwith East	38.8528	107.2094	3362	28	NNW	Granitic	53	2	1202
East Beckwith East	38.8569	107.2092	3147	34	NW	Granitic	50	2	1104
East Beckwith East	38.8582	107.2084	3093	31	NW	Granitic	55	4	1267
Gothic Mountain	38.8595	107.0091	3567	38	NE	Granitic	61	11	1463
Virginia Basin	38.9727	106.9788	3332	36	WSW	Quartzitic	118	10	3325
Virginia Basin	38.9726	106.8624	3313	34	WSW	Quartzitic	76	4	1953
Copper Creek	38.9613	106.9701	3048	32	NW	Granitic	50	3	1104
Copper Creek	38.9609	106.9693	3054	42	NNW	Granitic	42	2	842
Copper Creek	38.9598	106.9684	3109	26	W	Granitic	41	1	809
Copper Creek	38.9596	106.9670	3161	40	NW	Granitic	43	5	875
Queen Basin	38.9750	106.9430	3482	37	SW	Granitic	79	2	2051
Queen Basin	38.9747	106.9407	3539	38	SW	Granitic	89	3	2378
Queen Basin	38.9760	106.9384	3574	36	WSW	Granitic	61	4	1463
Queen Basin	38.9762	106.9357	3644	43	WSW	Granitic	72	6	1822
Queen Basin	38.9763	106.9328	3722	46	W	Granitic	51	9	1136
Queen Basin	38.9763	106.9328	3723	48	W	Granitic	49	4	1071
Ferris Creek	38.8519	106.8848	3005	40	W	Quartzitic	142	17	4109
Ferris Creek	38.8518	106.8823	3053	25	NW	Quartzitic	119	11	3358
Ferris Creek	38.8504	106.8804	3086	31	NW	Quartzitic	167	14	4925
Ferris Creek	38.8567	106.8624	3061	32	WSW	Quartzitic	131	7	3750
Ferris Creek	38.8548	106.8523	3336	26	NNE	Quartzitic	130	25	3717
Ferris Creek	38.8548	106.8523	3330	32	NNE	Quartzitic	123	3	3488
Ferris Creek	38.8382	106.8520	3293	28	NNE	Quartzitic	75	5	1920
Italian Mountain	38.9511	106.7457	3661	45	NE	Granitic	205†	41	6167
Italian Mountain	38.9519	106.7444	3644	36	NE	Granitic	86	6	2280
Italian Mountain	38.9527	106.7431	3607	37	NNE	Granitic	80	8	2084
Pieplant Creek	38.9559	106.5473	3519	41	S	Granitic	92	3	2476
Pieplant Creek	38.9559	106.5473	3519	41	S	Granitic	82	7	2149
Cumberland Pass	38.6930	106.4696	3566	32	SE	Granitic	91	8	2443
Cumberland Pass	38.6925	106.4689	3560	39	SE	Granitic	101	7	2770
Cumberland Pass	38.6909	106.4616	3572	36	SE	Granitic	109	14	3031
Cumberland Pass	38.6911	106.4610	3572	22	SE	Granitic	109	13	3031
Cottonwood Pass	38.8178	106.4011	3541	31	SSE	Gneissic	121	12	3423
Cottonwood Pass	38.8029	106.3852	3340	36	NNW	Gneissic	115	5	3227
Cottonwood Pass	38.8016	106.3841	3336	36	N	Gneissic	129	15	3684
Cottonwood Pass	38.7990	106.3819	3385	39	NE	Gneissic	113	13	3162
Mount Democrat	39.3237	106.1614	3755	41	NW	Granitic	75	3	1920
Mount Democrat	39.3261	106.1595	3729	39	NW	Granitic	108	3	2998
Mount Democrat	39.3296	106.1562	3659	42	NW	Granitic	76	5	1953
Mount Democrat	39.3304	106.1549	3639	47	NNW	Granitic	79	1	2051

* Mean of the largest five thalli on each lobe.
† Data considered statistical outliers and excluded from statistical analyses.

terminus. Rock substrates were mostly of granitic lithologies, but some lichens were measured on gneissic and quartzitic boulders.

Within each sampling site, we measured the largest 25 *Rhizocarpon* subgenus *Rhizocarpon* thalli. These measurements were taken along the longest axis of each thallus, including the black prothallus rim. Lichens with very irregular shapes and those that appeared to have grown together were not measured. We avoided sampling in any areas that appeared to have suffered from snowkill, a phenomenon which can significantly affect the age distribution of lichens (Benedict, 1990, 1993).

Results

Lichen thalli measured on rock glacier lobes varied in diameter from 28 to 260 mm. Five of the largest of these, ranging in size from 160 to 260 mm, were found on boulders greater than 4 m across occurring on a single lobe on the eastern flank of Italian Mountain. Hamilton and Whalley (1995) concluded that anomalously large thalli represent lichens that may have been established on rock substrates prior to deposition on the rock glacier. Therefore, following their suggestion, these larger lichens were excluded from subsequent statistical treatments of the data

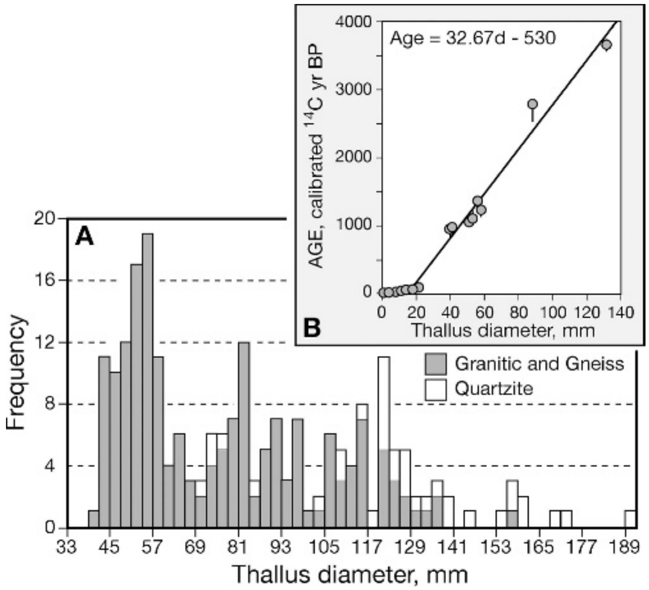

FIGURE 3. (A) Size-frequency distribution for lichen thalli diameters on rock glacier surfaces. A distinction is made between the distribution resulting from measurements made exclusively on granitic and gneissic substrates and those including measurements made on quartzitic substrates because growth rates on the latter may differ. (B) Inset shows the growth curve used to calculate the age of the peaks in the size-frequency plot (after Benedict, 1993). The linear regression ($r^2 = 0.985$) is valid for thalli greater than 20 mm on the long axis; lichen growth rate is ~0.031 mm yr^{-1}. The ages of radiocarbon-dated surfaces have been recalibrated using CALIB 5.0.1 (Stuvier et al., 2005). Error bars indicate 1σ uncertainty in the radiocarbon ages.

presented here. Three thalli on quartzite boulders from two lobes at Ferris Creek were between 168 and 190 mm, but the size-frequency distribution of thalli on these lobes does not allow these larger thalli to be distinguished as statistical outliers.

Innes (1984) found that the most accurate method for establishing maximum lichen size for a landform was to use the mean diameter of the largest five thalli from a sampling site. Using the five largest thalli from each lobe, the thalli size range becomes 39–190 mm (Fig. 3A). For lobes where lichens were measured in multiple locations, we only used the largest mean in the data analysis.

The oldest lichens on an individual lobe were nearly always found just above the break in slope above the terminus. Lichens become smaller, and therefore younger, with increasing distance upslope. Generally, the differences among the mean diameters of the largest five lichens in several different sampling sites located across the front of a lobe were less than 10 mm. Within a single rock glacier complex, such as those at East Beckwith East and Queen Basin (Figs. 2A and 2C), nested and overlapping lobes exhibit older ages with increasing distance from the talus source. These observations support our assumption that each individual lobe represents a separate period of rock glacier activity.

Discussion

SIZE AND AGE DISTRIBUTIONS

A local lichen growth curve has not yet been developed for the study area due to a lack of dated surfaces. Therefore, lichen ages in this study were calculated, as they have been elsewhere (e.g., Miller, 1973; Nicholas and Butler, 1996; Munroe, 2002), using a growth curve (Fig. 3B) developed for *Rhizocarpon geographicum* in the Front Range of Colorado (Benedict, 1967). Revisions to this growth curve (Benedict, 1993) include the addition of a lichen with a thallus diameter of 131 mm, so the curve now extends to nearly 4000 yr BP without extrapolation. We also recalibrated the ^{14}C age data used in this growth curve using CALIB 5.0.1 (Stuiver et al., 2005).

The applicability of Benedict's (1967, 1993) growth curve to the Elk Mountains and Sawatch Range warrants some additional discussion. Climate and substrate lithologies in the study area, with the exception of quartzitic boulders, are both very similar to those in the Front Range. Nonetheless, differences in climate, though very minor, may affect lichen growth rates. Of note, Birkeland's (1973) earlier work in the Elk Mountains suggested that the growth rate for *Rhizocarpon geographicum* here could possibly be lower than that in the Front Range. In addition, there are conflicting conclusions regarding the influence of lithology (and associated textures) on lichen growth (cf. Bull and Brandon, 1998; Noller and Locke, 2000); i.e., growth rates of lichens on quartzitic substrates *may* differ from those on the granitic and gneissic rocks upon which Benedict's (1967, 1993) growth curve is based. Munroe (2002), in particular, concluded that the use of Benedict's (1967, 1993) curve could potentially underestimate lichen ages on quartzite. Lacking a more robust data set to assess possible differences in growth rates on varied lithologies, only those lichen measurements from lobes comprised of granitic and gneissic boulders are emphasized in the analyses and conclusions that follow.

We applied two different statistical methods to separate the lichen age-frequency data into statistically different groups: K-means clustering analysis and cumulative probability analysis (CPA). K-means clustering analysis separates a data set into groups by maximizing between-group variation while minimizing within-group variation. Excluding measurements made on quartzitic boulders, this technique defined four groups (Fig. 4A) with mean ages of 1136, 2149, 3031, and 3552 yr BP. A one-way ANOVA test confirmed that all four groups are significantly different at a 95% confidence level. Inclusion of the quartzitic boulder data does not significantly change these results, though the mean of the third-oldest peak becomes 219 years older.

CPA sums the probability distributions of a data set and incorporates normally distributed errors. Using the ages of *each* of the five largest lichens on granitic and gneissic substrates from all lobes and an assigned common 1σ uncertainty of 150 yr (see below), three groups were identified with peak amplitude ages of 1178, 2012, and 3110 yr BP (Fig. 4B). These groups correspond very closely with the ages of the three youngest groups in the K-means clustering analysis. We note that including lichen measurements made on quartzite boulders in this analysis only has a sizeable affect on the oldest group, increasing it to 3328 yr BP (Fig. 4B). We repeated this analysis using the *mean* age of the largest five lichens from each rock glacier lobe, and again three very similar age groups were identified, with peak amplitude ages of 1132, 2044, and 3088 yr BP (Fig. 4C). Inclusion of the measurements made on quartzite in this analysis changes the peak amplitude ages by between 72 and 136 years. Innes (1984) concluded that using the mean age of the largest five lichens provides the most accurate age of a landform, but our results suggest differences in age assignment using the largest five or the mean of the largest five thalli are minimal (<4%).

For measurements made on granitic and gneissic boulders, three clusters of lichen ages for rock glacier surfaces are common to all three statistical analyses, having ages of approximately 1150, 2070, and 3080 yr BP. If lichen growth rates on quartzite are similar to those on granitic and gneissic substrates, rock glacier

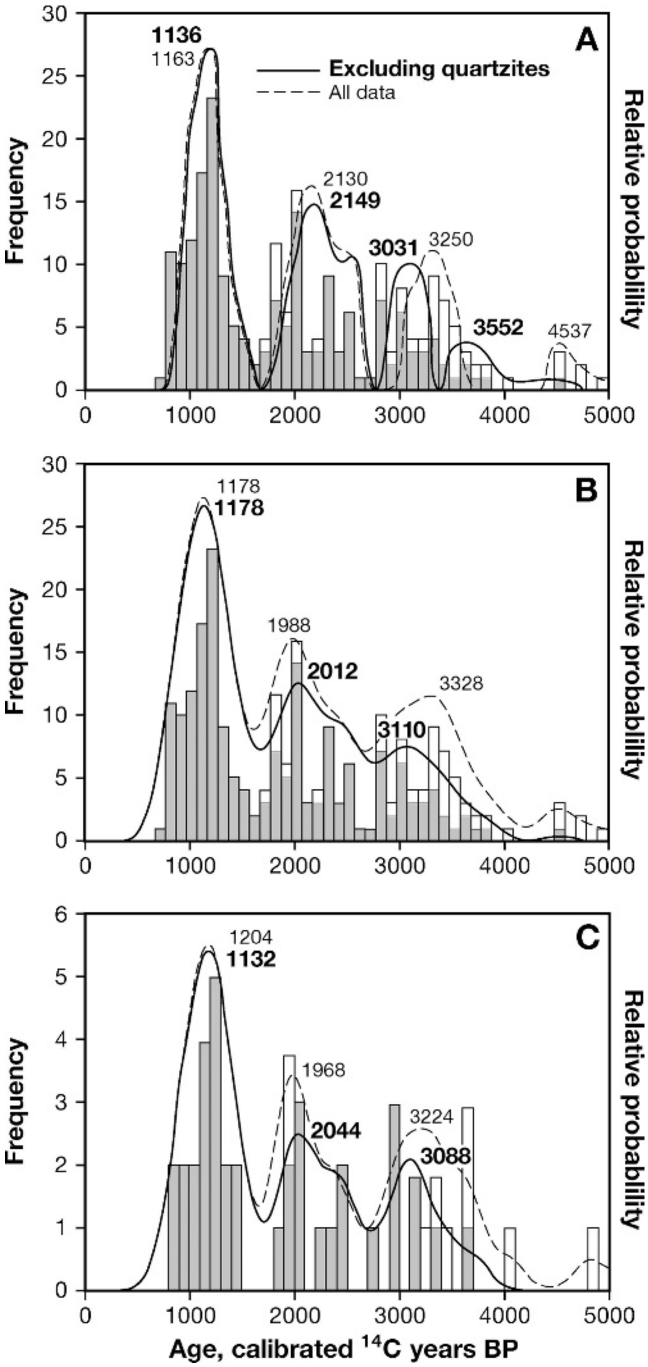

FIGURE 4. Frequency distribution and clustering of rock glacier ages. (A) K-means clustering analysis. Each peak on the curve is statistically distinct at a 95% confidence level. (B) Cumulative probability Gaussian distributions based on the largest five lichen thalli from each lobe and an assigned uncertainty (1σ) of 150 years in each age. (C) Cumulative probability Gaussian distributions based on the *mean* of the largest five thalli from each lobe, also using an assigned uncertainty (1σ) of 150 years. The histograms in (B) and (C) differ because (B) is based on five times more data points than (C).

surfaces cluster around ages of 1180, 2030, and 3270 yr BP. An older peak in the K-means clustering analysis (3550 yr BP; Fig. 4A) was not differentiated by CPA. To varying degrees, all analyses reveal a subtle peak or "shoulder" between 2360 and 2540 yr BP. This peak is not statistically significant, however, and it and the 3550 yr BP peak are therefore not considered further in this paper.

The total uncertainty in the thallus diameter of each statistically defined group is difficult to quantify. Uncertainties arising from sampling and the variability in the diameter of the largest five lichens on a specific lobe, as discussed above, are thought to be small (<5%). Additional uncertainty arises from the method by which the data are segregated into groups. A Euclidian distance metric was used in the K-means clustering analysis, but other metrics result in only slightly different (<1%) group mean ages. With due consideration of the greater uncertainties in the response time of rock glacier systems (Olyphant, 1987), lags in lichen colonization (probably fewer than 50 years; Beschel, 1961; Evans et al., 1999; Noller and Locke, 2000), and growth rates, a total uncertainty of 150 years is probably reasonable (cf. Kirkbride and Brazier, 1998). This uncertainty is most probably skewed toward older ages, making the assigned lichenometric ages minimum estimates.

LICHENOMETRIC AGE SIGNIFICANCE AND PALEOCLIMATIC IMPLICATIONS

It is generally accepted that the age of a rock glacier surface represents the time of debris production and deposition (Hamilton and Whalley, 1995; Kirkbride and Brazier, 1995; Konrad and Clark, 1998; Sloan and Dyke, 1998). Dated intervals of rock glacier activity are typically interpreted as periods of climate deterioration (e.g., Benedict, 1967; Birkeland, 1973; Miller, 1973; Calkin et al., 1987; Nicholas and Butler, 1996). However, few studies have explicitly considered how the paleoclimatic interpretation of these ages might differ depending on the mode of rock glacier genesis. Morris (1987), Morris and Olyphant (1990), Brazier et al. (1998), and Hughes et al. (2003) stress that while rock glaciers of a periglacial origin likely indicate cooler temperatures, glacigenic rock glaciers *may* form under a warming climate. Given our assumption that most of the rock glaciers investigated in this study are periglacial in origin, we suggest that the lichen-dated surfaces documented here reflect rock glacier activity associated with intervals of cooler climate within the study area. Such conditions would not only promote the formation of interstitial ice, but also potentially increase debris supply by virtue of increased mechanical weathering by frost cracking (Walder and Hallet, 1986; Matsuoka, 2001; Hales and Roering, 2005). At some point, a combination of sufficient debris thickness and surface slope would generate the driving stress required for mobilization of the ice-rock mixture (Warhhaftig and Cox, 1959; Kirkbride and Brazier, 1995). Rock glaciers remain active so long as driving stresses are sufficient and climate is such that interstitial ice could persist. In contrast, talus produced during warmer intervals would accumulate below rock walls, but because interstitial ice cannot form, neither can new rock glaciers develop nor can existing ones be reactivated.

The possibility exists, however, that rock glacier lobes were intermittently active during *sustained* periods of cooler climate (Kirkbride and Braizier, 1995). Rock glaciers might have become inactive as flow attenuated debris thicknesses and surface slopes and thus reduced driving stresses below some critical value. Continued debris accumulation might have eventually reactivated lobes as the threshold stress was restored. Under this scenario, rock glacier activity is not recording discrete intervals of cooling. Also, reactivation—caused by *either* increased talus deposition or a distinct cooling event—can result in a mantle of younger debris that buries older lobes or rock glacier surfaces. Consequently, Kirkbride and Brazier (1995) argued that (1) climate records derived from rock glacier activity may be incomplete, or

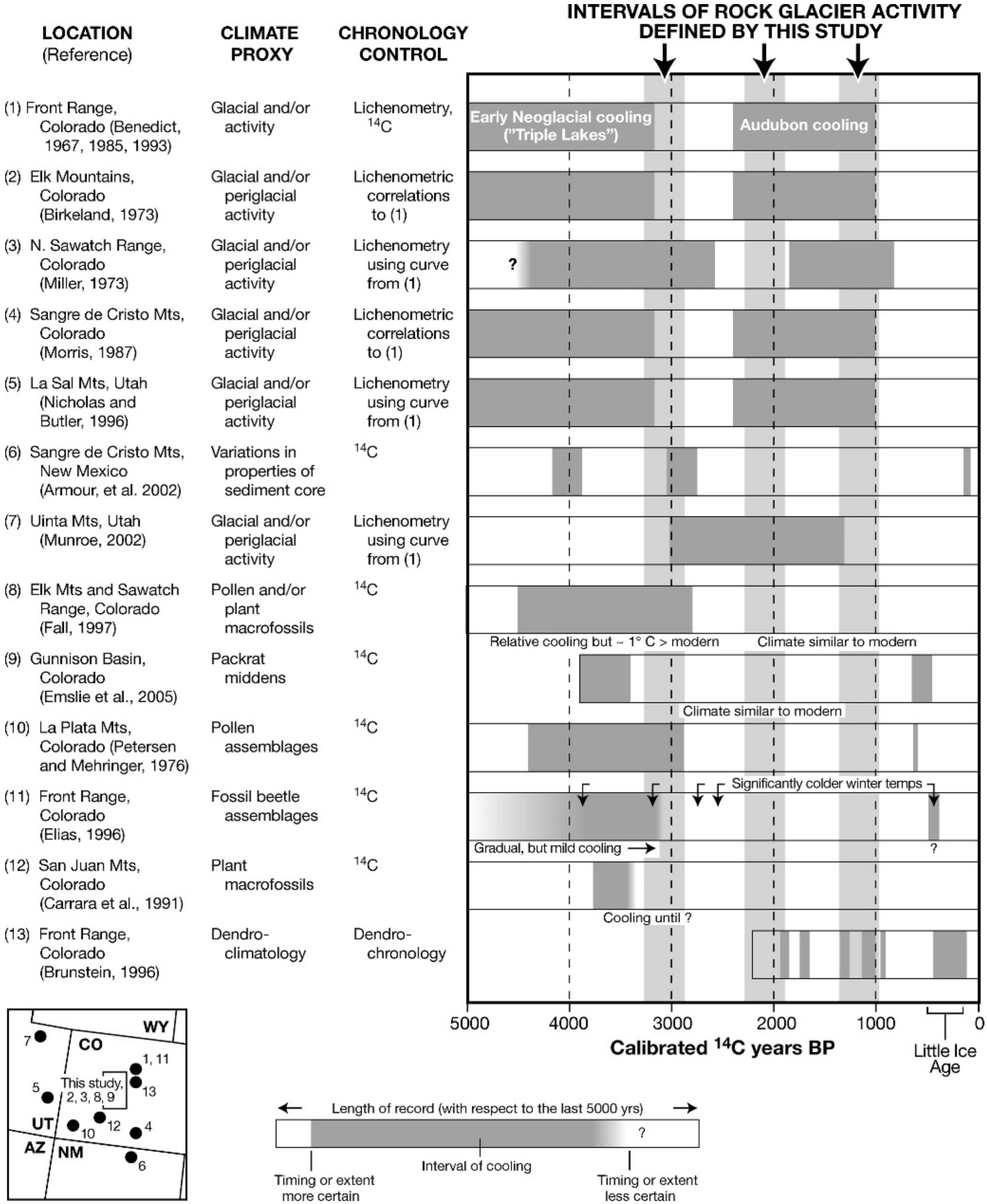

FIGURE 5. Comparison of intervals of rock glacier activity determined from this study (vertical bars) with selected paleoclimate proxies (horizontal bars) in the southern Rocky Mountains and surrounding areas. The approximate locations of the latter are shown in the inset map. Cool intervals (darker gray) are variously defined in these studies as cooler than present, cooler than some long-term mean, cooler mean annual temperatures, cooler summer temperatures, and so forth. See text for discussion.

complicated by non-climatically forced activity; (2) rock glaciers within individual basins may not show a coherent pattern of activity; and (3) regional climate signals deduced from rock glacier activity will necessarily be noisy.

Based on the foregoing, some caution should therefore be exercised when making climatic interpretations of rock glacier activity. However, we feel the lichen ages of 1150, 2070, and 3080 yr BP date discrete intervals of debris delivery and transport to the front of individual rock glacier lobes under cooler climates. This conclusion is based on the statistically significant clustering of lichen ages seen in Figure 4 and the close correlations between these ages and those of cool intervals inferred from other climate proxies documented elsewhere in the Rocky Mountain region (Fig. 5). Independent of climate, the stochastic nature of rockfall, hence debris deposition on rock glacier surfaces, should preclude a clustering of lichen ages. Following Kirkbride and Brazier (1995), we note that even given the possibility of rockfall-driven "pulses" of rock glacier activity during prolonged cool intervals, it is unlikely that the latter persisted for 2000 years (ca. 3080 to 1150 yr BP) during the late Holocene (see below). We suggest the greater frequency (or strength) evident in the youngest peak at 1150 yr BP may be an artifact of the aforementioned burial of older rock glacier lobes by younger advances. Finally, it bears mentioning that if data from those few rock glaciers that *may be* of glacigenic origin is omitted from the foregoing analyses, the ages of individual peaks do not change significantly.

Episodes of rock glacier activity in the Sawatch Range and Elk Mountains are compared schematically with other proxy records of Holocene climate change in the southern Rocky Mountains and adjacent regions in Figure 5. Direct comparisons and/or correlations must necessarily be tentative due to (1) the differences among individual studies of temporal resolution and age control of chronologies; (2) the requirement to recalibrate ^{14}C ages where necessary and possible (using CALIB 5.0.1, Stuiver et al., 2005); and (3) the methodological and/or sampling differences (e.g., ages obtained using the largest versus the mean of the largest five lichens). Nonetheless, intervals of cooler climate inferred from rock glacier activity within the study area correspond with the broader regional pattern of mid- to late Holocene climate change.

Within the immediate study area, Fall (1997) used pollen spectra and plant macrofossils to reconstruct Holocene climates and found evidence indicating Neoglacial cooling beginning 4000 ^{14}C yr BP (ca. 4500 cal. yr BP) and lasting for perhaps 2000 years. Fall (1997) estimated mean annual temperatures during that interval decreased by 0.8°C (but were still 0.6–1.2°C warmer than modern climate), and conditions were drier from 4000 to 2600 ^{14}C yr BP (ca. 4500 to 2800 cal. yr BP). Winter precipitation may have dominated during this interval, implying that summers were very dry. By ca. 2000 yr BP, the modern climate was established.

In the upper Gunnison Basin, plant material preserved in packrat middens also suggests a cooling period beginning 4000 ^{14}C yr BP and lasting until at least 3180 ^{14}C yr BP (ca. 4500 to 3500 cal. yr BP; Emslie et al., 2005). Pollen records from the San Juan Mountains in southwestern Colorado suggest cooling began by about 3500 ^{14}C yr BP (ca. 3780 cal. yr BP; Carrara et al., 1991). Analyses of fossil beetle assemblages in the Front Range of Colorado led Elias (1996) to conclude that a period with a gradual trend toward cooler summers was underway by 7800 yr BP but temperatures remained comparable to those today. However, between 2965 and 2680 ^{14}C yr BP (ca. 3150 and 2850 cal. yr BP), summer temperature could have been 1–2.5°C cooler than present. Mean January temperatures throughout most of the Holocene are thought to have been below modern values until the last millennia (Elias, 1996). In the La Plata Mountains, mean July temperatures are inferred to have been cooler between 4000 and 2500 ^{14}C yr BP (ca. 4450 to 2880 cal. yr BP) on the basis of pollen spectra (Petersen and Mehringer, 1976). The oldest episode of rock glacier activity revealed by our study, occurring ca. 3080 yr BP, appears to be coeval with the earlier stages of the Neoglacial cooling documented by these studies. In addition, this episode coincides with widespread glacial or periglacial activity documented in the region, in particular the northern Sawatch Range (Miller, 1973), Front Range (Benedict, 1973, 1985), Sangre de Cristo Range (Morris, 1987; Armour et al., 2002), the La Sal Mountains (Nicholas and Butler, 1996), and the Uinta Mountains (Munroe, 2002).

The younger two episodes of rock glacier activity in the study area are correlative to Audubon-aged (Benedict, 1973, 1985, 1993) glacial deposits and rock glacier surfaces found throughout Colorado (Benedict, 1973, 1985; Birkleland, 1973; Miller, 1973; Morris, 1987) and Utah (Nicholas and Butler, 1996; Munroe, 2002). The cool intervals associated with these episodes are not apparent in other climate proxies, with the exception of a record of latewood frost-rings from the Front Range (Brunstein, 1996). These younger rock glacier surfaces also tend to occur in, although they are not exclusive to, the western portion of the study area (Table 1). We cannot say whether this is a result of more persistent rock glacier activity due to spatial trends in regional climate, differences in microclimates, more complete burial of older surfaces during younger advances, or simply sampling bias. Future work will focus on these questions by expanding the database and developing a local lichen growth curve.

Conclusions

Rock glacier surfaces and/or lobes of three discrete ages occur in the Sawatch Range and Elk Mountains, each representing an episode of increased debris production, formation of interstitial ice, and subsequent flow under cooler climates. The earliest episode, dated at 3080 yr BP, is correlative with early Neoglacial glacial and periglacial activity elsewhere in the southern Rocky Mountains and adjacent areas and is closely associated with a cool interval identified in other climate proxies from the region. The younger two episodes, dated at 2070 and 1150 yr BP, also generally correspond to regional glacial and periglacial activity during the Audubon stade. As of yet, however, these intervals have not been recognized in any other local climate proxies. This may suggest that these events are related to very local topoclimatic modification of regional (or larger) scale climate forcings.

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