Surface-exposure ages of Front Range moraines that may have formed during the Younger Dryas, 8.2 cal ka, and Little Ice Age events

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Abstract

Surface-exposure (^{10}Be) ages have been obtained on boulders from three post-Pinedale end-moraine complexes in the Front Range, Colorado. Boulder rounding appears related to the cirque-to-moraine transport distance at each site with subrounded boulders being typical of the 2-km-long Chicago Lakes Glacier, subangular boulders being typical of the 1-km-long Butler Gulch Glacier, and angular boulders being typical of the few-hundred-m-long Isabelle Glacier. Surface-exposure ages of angular boulders from the Isabelle Glacier moraine, which formed during the Little Ice Age (LIA) according to previous lichenometric dating, indicate cosmogenic inheritance values ranging from 0 to ~3.0 ^{10}Be ka.[1] Subangular boulders from the Butler Gulch end moraine yielded surface-exposure ages ranging from 5 to 10.2 ^{10}Be ka. We suggest that this moraine was deposited during the 8.2 cal ka event, which has been associated with outburst floods from Lake Agassiz and Lake Ojibway, and that the large age range associated with the Butler Gulch end moraine is caused by cosmogenic shielding of and(or) spalling from boulders that have ages in the younger part of the range and by cosmogenic inheritance in boulders that have ages in the older part of the range. The surface-exposure ages of eight of nine subrounded boulders from the Chicago Lakes area fall within the 13.0–11.7 ^{10}Be ka age range, and appear to have been deposited during the Younger Dryas interval. The general lack of inheritance in the eight samples probably stems from the fact that only a few thousand years intervened between the retreat of the Pinedale glacier and the advance of the Chicago Lakes glacier; in addition, bedrock in the Chicago Lakes cirque area may have remained covered with snow and ice during that interval, thus partially shielding the bedrock from cosmogenic radiation.
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1. Introduction

Temperate alpine glaciers respond rapidly to climate change on both regional and global scales. In terms of climate forcing, fluctuations in the sizes of existing western North American Glaciers have been shown to be associated with changes in atmospheric circulation over the North Pacific Ocean and western North America; in particular, fluctuations in the size of modern glaciers have been correlated with changes in the Southern Oscillation Index and Northern Hemisphere air temperature (McCabe et al., 2000). During the past 45 years (yr), low-elevation coastal glaciers indicate a strong association between volume loss and increases in summer air temperature (Dyurgerov and McCabe, 2006). Thus, knowledge of the timing of past glacier fluctuations may aid in understanding past changes in atmospheric circulation and climate.

The purpose of this paper is to assess the timing of relatively small latest Pleistocene and Holocene glacial advances in the Front Range and to evaluate the accuracy of surface-exposure dating of those advances. In a recent

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[1]Surface-exposure ages in this paper are labeled ^{10}Be; radiocarbon ages are labeled ^{14}C ka, calendar and calibrated radiocarbon ages are labeled cal ka, and layer-based ice-core ages are labeled ka. ^{14}C ages, calibrated ^{14}C ages, and ice core ages are given relative to AD 1950, whereas ^{10}Be ages are given relative to the sampling date. Radiocarbon ages were calibrated using CALIB 5.01 and the INTCAL04 data base Stuiver et al. (2005). Ages estimated using CALIB 5.01 are shown in terms of their 1-sigma range.

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Table 1
Calibration of radiocarbon dates discussed in text

Locality	Material type	Organic fraction	^{14}C age (ka)	Calibrated age (cal ka)	Relative area
Fourth of July valley	Basal peat*	Humic	9.150±0.150	10.52–10.19	0.96
	Basal peat*	Humin	9.280±0.150	10.60–10.26	0.93
Caribou Lake valley	Humus, wood*	Humic	8.570±0.220	9.91–9.37	0.92
	Humus, wood*	Humin	9.010±0.140	10.30–9.90	0.97
Caribou Lake valley	Decomposed peat*	Humic + humin	9.315±0.105	10.66–10.38	0.96
Caribou Lake valley	Mucky peat*	Humic + humin	9.280±0.115	10.58–10.29	1.00
	Humus stained clay*	Humic + humin	9.320±0.115	10.68–10.38	0.95
Caribou Lake valley	Wood and spruce cones*	Humin	9.520±0.160	11.10–10.65	0.96
	Wood and spruce cones*	Humic	9.720±0.220	11.36–10.70	0.98
Caribou Lake valley	Basal peat*	Humic	9.700±0.215	11.33–10.69	1.00
	Basal peat*	Humin	9.915±0.165	11.72–11.19	1.00
Sky Pond	Lepidoptera mandibles	na	9.970±0.080	11.42–11.27	0.52
Sky Pond	Lepidoptera mandibles	na	10.410±0.090	12.40–12.12	0.80
Sky Pond	Lepidoptera mandibles	na	12.040±0.060	13.85	na
Satanta Peak base	Peat above clastics	Humin	7.440±0.125	8.39–8.16	0.97
Satanta Peak base	Muck below clastics	Humin	7.900±0.125	8.80–8.59	0.61

*In these instances, the material was collected from sediment immediately overlying the Satanta outwash. The total area under the probability curve was normalized to one and relative area represents the fraction of the total probability explained by the calibrated range. All ^{14}C ages were calibrated using CALIB 5.01 except the oldest age which was calibrated using Hughen et al. (2004). Radiocarbon dates for Satanta Peak base, Fourth of July valley, and Caribou Lake valley from Benedict (1973, 1981, 1985); radiocarbon dates for Sky Pond from Menounus and Reasoner (1997). na = not applicable.

study of early/middle Holocene glacier advance in western Austria, Kerschner et al. (2006) obtained a relatively narrow and consistent age range (8.69±0.41 to 8.01±0.69 ^{10}Be ka) for 1.0–1.9 m angular moraine-crest boulders deposited during the Kromer Stadial which suggests that moraines deposited by early/middle Holocene glaciers, having short (⩽2 km) transport distances, might yield consistent surface-exposure ages.

In this paper, we present new surface-exposure ages of three post-Pinedale glacial moraines in north-central Colorado and compare the results with previously published surface-exposure ages of moraines in west-central Wyoming (Gosse et al., 1995) and ^{14}C-based estimates of post-Pinedale/Tioga glacial events documented in north-central Colorado and the central Sierra Nevada (Benedict, 1981, 1985; Clark and Gillespie, 1997). We also compare the timing of Front Range early Holocene/latest Pleistocene glacial events with the Younger Dryas and 8.2 cal ka climate events (Barber et al., 1999). Our results indicate a wide range of ages for early and late Holocene advances and we discuss possible mechanisms that may account for the variability in age.

2. Previous work

Using dating methods such as soil-profile development, maximum lichen diameter, and granite weathering characteristics, Benedict (1973, 1981, 1985) suggested that glacial advances in the Front Range occurred after recession of latest Pleistocene Pinedale glaciers, including Ptarmigan and multiple Satanta Peak advances.

Benedict (1981) estimated that outwash from the younger of two Satanta Peak moraines in Fourth of July valley (located 1 km southeast of Arapaho Pass) was trenched by the North Fork of Middle Boulder Creek prior to ~10.4 cal ka (Table 1, Fig. 1).[2] Benedict (1985) also dated basal peat and woody materials overlying Satanta Peak outwash deposits in Arapaho Creek valley (located 0.8 km southeast of Arapaho Pass, Fig. 1). The ages range from 11.4 to 10.1 cal ka (Table 1). Based on ages of these organic sediments, and a conjecture that Pinedale ice had retreated to its cirque by 12 ^{14}C ka, Benedict (1985) concluded that the Satanta Peak advance may have begun as early as 12 ^{14}C ka (13.8 cal ka) and ended by ~10 ^{14}C ka (11.5 cal ka).

Work by Madole (1986) supports Benedict's (1981, 1985) conjecture that Pinedale ice had retreated to its source area(s) by 12 ^{14}C ka. Pinedale glaciers probably disappeared from the Front Range sometime after 13.68 ^{14}C (16.9 cal ka) and before 11.76 ^{14}C (13.5 cal ka). The older limit is based on a date for the deglaciation of an area located 1 km west of Buffalo Pass (Madole, 1986, Fig. 1), which was near the divide mantled by a Pinedale icecap. The younger limit is based on a lake-sediment age from a site located on the Continental Divide ~1.5 km south of Buffalo Pass (Madole, 1980). La Poudre Pass (Fig. 1), a site on the ice divide of a transection glacier that flowed north into the Cache la Poudre River valley and south into the Colorado River valley, was completely ice free by ~10.0 ^{14}C (11.5 cal ka) (Madole, 1980).

Basal sediments from Front Range cirques also have yielded ages indicating the disappearance of Pinedale glaciers prior to 12.0 ^{14}C ka (13.8 cal ka); e.g., Davis et al.

[2]When available we reference only the calibrated humin-based ^{14}C ages listed in Table 1. And when referencing the date, we use the midpoint of the range.

(1992) reported an age of 12.28±0.34 ^{14}C (14.3 cal ka) from basal sediment in Blue Lake (Table 1), a high-elevation (3450 m) body of water located at the head of the South St. Vrain Creek drainage basin (Fig. 1).

Benedict (personal communication, 2006) cored a fen in a trough that separates two end moraines that impound Caribou Lake. Locust (*melanopolus* sp.) mandibles from proglacial lake sediments beneath 2 m of peat yielded an age of 11.70±0.60 ^{14}C ka (~13.6 cal ka); Table 1; Dethier et al. (2003, p. 100). These data, together with minimum ages for the late Satanta Peak advance as old as 11.5 cal ka (Benedict, 1981, 1985), suggest that the second Satanta Peak advance may have occurred during the Younger Dryas event (12.94–11.64 ka) and may be correlative to the Titcomb Lakes advance (13.8–11.4 ^{10}Be ka) in west-central Wyoming (Gosse et al., 1995). The early Satanta Peak advance occurred prior to the Younger Dryas event. The early Satanta Peak glacier receded at approximately the same time as the recession of the Sierran Recess Peak glacier (11.19±0.07 ^{14}C ka; ~13.1 cal ka; Clark and Gillespie, 1997).

In the upper part of the Front Range Glacier Creek drainage basin, a broad-crested end moraine is situated upvalley of Sky Pond (Fig. 1, 3320 m), and a second end moraine is situated within Sky Pond. Menounos and Reasoner (1997) correlated a layer of clastic sediment at a depth of 1.8–2.6 m, in a sediment core from Sky Pond, to a Satanta Peak advance which they associated with the upvalley moraine. The interval of clastic sediment is bracketed by ages of 10.41±0.09 and 9.97±0.08 ^{14}C ka (12.3–11.3 cal ka, Table 1), a time interval that overlaps the latter part of the Younger Dryas. They obtained an age of 12.04±0.06 ^{14}C ka (13.9 cal ka, Table 1) at a depth of 3.3 m on Gyttja immediately above a diamict. They suggested that the diamict is a till that was deposited during deglaciation of the Sky Pond Basin and that the 12.04±0.06 ^{14}C ka age probably represents a minimum limiting age for the moraine in Sky Pond. Alternatively, the diamict may represent an early Satanta Peak advance or it may be a pile of rubble transported to the lake by mass movement.

3. Sites

In this paper, we examine surface-exposure ages of boulders from three sites in north-central Colorado: (1) Isabelle Glacier (Figs. 1 and 2), (2) Butler Gulch (Figs. 1, 3, 4) and (3) Chicago Lakes (Figs. 1 and 5). The bedrock and all boulders sampled at these sites consist of Middle Proterozoic granite and quartz monzonite. Figs. 3 and 5 show surficial geology of the Butler Gulch and the Chicago Lakes sites. Sample coordinates, sample elevations, boulder heights, and sample thicknesses are listed in Table 2.

The moraine at Isabelle Glacier has been assumed to date to the Little Ice Age (LIA) and provides a means of estimating cosmogenic inheritance of a young moraine deposited near its cirque source area. In general, the timing of the LIA in the Northern Hemisphere is arguable, with glacier advance, decreased air temperature, and meridional atmospheric intensity being used to define the beginning and end of the LIA. Neither of these parameters indicates a synchronous change of climate across the globe or Northern Hemisphere (Holzhauser, 1997; Kreutz et al., 1997; Jones et al., 1998; Mann et al., 1999; Crowley and Lowery, 2000; Grove, 2001a, b; Briffa and Osborn, 2002). Not only do the different proxies of the LIA yield different LIA histories, high-resolution tree-ring records from several areas around the globe do not even indicate the presence of a LIA signal (Bradley and Jones, 1992). For our purposes the lichen-based age range (AD 1650–1850) calculated for glaciers of the LIA Arapaho Peak advance by Benedict (1973) will be used to estimate the timing of the LIA boulders found at Isabelle Glacier.

The boulders at the Isabelle Glacier site are large, angular,[3] free of lichen, and lie within a few hundred meters of the cirque headwall (Fig. 2). At the Butler Gulch site, a series of nested arcuate moraines extend about 1 km from the cirque headwall. Boulders at this site tend to be blocky and are relatively small (<1 m tall) (Fig. 6). Most boulders from Chicago Lakes are relatively large >1 m tall), tend to be subrounded (Fig. 7), and the distal edge of moraines extend about 2 km downvalley from the cirque headwall. Boulders at this site are distributed along the side and downvalley of Lower Chicago Lake (Fig. 5).

Moraines like those in Butler Gulch and the Chicago Lakes areas have long been recognized in the northern Front Range and the adjacent Never Summer Range, and have been given various names (Ives, 1938; Ray, 1940, Eschman, 1955; Richmond, 1960; Madole, 1963, 1972; Benedict, 1973, 1981, 1985; Davis, 1987, 1988; Davis and Osborn, 1987). Although remnants of these moraines are widespread, they are difficult to distinguish in some valley heads, particularly those that are deep, narrow, and fall abruptly a short distance away from cirques. In such places, till tends to be scattered over slopes that also are littered with bouldery mass-movement deposits of various ages. In contrast, moraines are easily recognized in places where valley floors fronting cirques are relatively broad and slope at low angles for at least 2–3 km. The latter describes the terrain at the head of North Boulder Creek (Madole, 1972; Benedict, 1973, 1981, 1985; Davis, 1987, 1988; Davis and Osborn, 1987) and in several drainage basins farther south, such as those of Clear Creek (which includes the Butler Gulch and Chicago Lakes areas) and the North Fork of the South Platte River.

4. Methods

4.1. Sample collection

Samples were collected with hammer and chisel from the tops of boulders as close to the center of the upper surface as possible. The location and elevation of each boulder was

[3] The terms angular, subangular, and rounded are used in this paper in the sense of Pettijohn (1949, pp. 58–59).

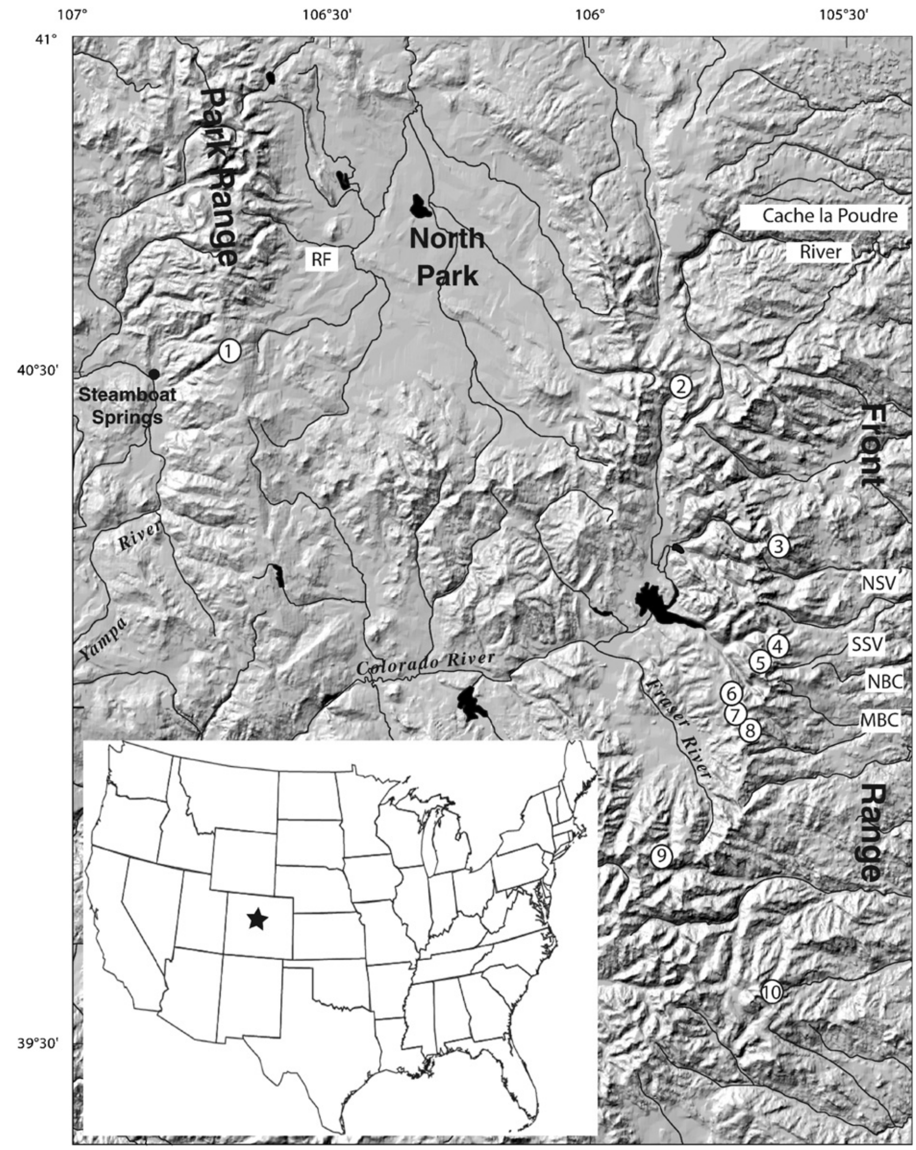

Fig. 1. Map showing sites mentioned in text. Sites are (1) Buffalo Pass, (2) La Poudre Pass, (3) Sky Pond, (4) Blue Lake, (5) Isabelle Glacier, (6) Caribou Lake, (7) Arapaho Pass, (8) Fourth of July valley, (9) Butler Gulch, and (10) Chicago Lakes. NSV = North St. Vrain River, SSV = South St. Vrain River, NBC = North Boulder Creek, MBC = Middle Boulder Creek. Star in US inset indicates site location.

measured using a hand-held global positioning system (GPS). Elevation was further refined using sample locations plotted on topographic maps; the estimated horizontal accuracy was 3–8 m and the estimated vertical accuracy was better than 10 m. Given the difficulty of finding large unfractured boulders on moraine ridge crests at the Butler Gulch site, some fractured boulders were sampled at ridge crest and swale locations. All but two boulders sampled at the three sites came from end moraines. The two boulders (BGRG03-1, 2) came from a rock glacier on the south side of the Butler Gulch valley, 1.5–2.0 km east of the outermost arcuate early Holocene moraine.

4.2. ^{10}Be determinations

Laboratory and procedures for ^{10}Be age determinations are the same as those in Benson et al. (2005). The chemical and isotopic ratios used in the calculation of surface-exposure ages are given in Table 3. A present-day sea-level high-latitude (SLHL) production value of 6.2 ± 0.1 at/g yr^{-1},[4] calculated using the scaling model of Lifton et al. (2005), a variable atmosphere (Stone, 2000), the 1945 Definitive Geomagnetic Reference Field (DGRF), and samples from the calibration sites discussed in Nishiizumi et al. (1989), Bierman et al. (1996), Gosse and Klein (1996), Kubik and Ivy-Ochs (2004), and Abramowski (2004) were applied in calculating the surface-exposure ages. The EXCEL worksheet included in Lifton et al. (2005) was

[4]The 0.1 at/g yr^{-1}. 1$-\sigma$ value for the ^{10}Be production rate is small in that it is a function of the number of iterations done in the spreadsheet; i.e., $1\sigma = fcn\sqrt{n}$, where n is the number of iterations made in calculating the ^{10}Be age. If one averages over the individual calculations of production rate, the 0.1-σ value equals ~0.3 at/g yr^{-1}.

Fig. 2. Isabelle Glacier and moraine. Boulder in foreground with circular lid on surface is 1 m tall and 2 m wide.

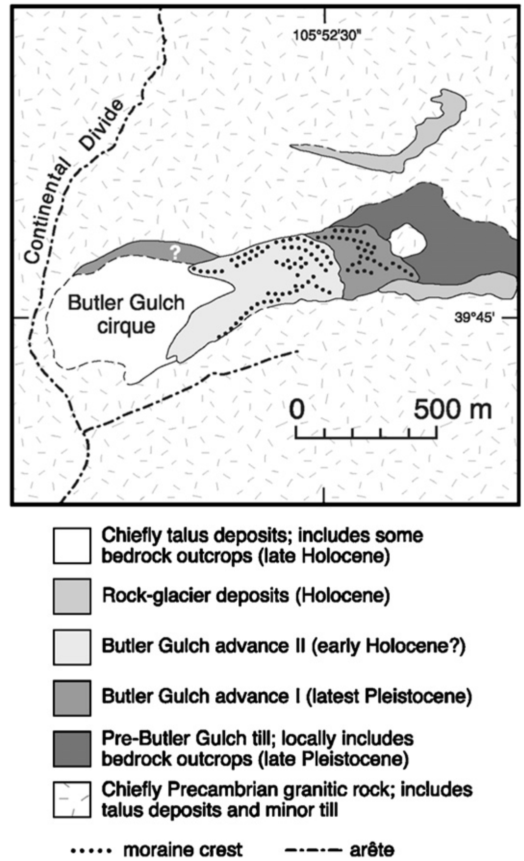

Fig. 3. Surficial geologic map of Butler Gulch area. Figure was not corrected for radial distortion associated with transfer of topography from aerial photos.

used to calculate ^{10}Be ages of the boulders. The EXCEL-based program accounts for geomagnetic effects (intensity of the geomagnetic field and position of the geomagnetic dipole axis) on time-integrated ^{10}Be production and includes atmospheric corrections after Stone (2000). The 6.2 ± 0.1 at/g yr^{-1} value is valid for SLHL under present-day geomagnetic conditions (intensity and orientation). The procedure for estimating ^{10}Be ages used by Lifton et al. (2005) differs from that used by other workers in the field that do not consider geomagnetic effects and that obtain a mean production rate by integrating its value over several thousand years. In the latter studies, Stone's (2000) production value of 5.1 ± 0.3 at/g yr^{-1} is commonly used to obtain ^{10}Be ages of quartz samples. This value represents an integration of cosmogenic production over 13 to >24 cal ka. A surface-exposure age calculated using a constant 5.1 at/g yr^{-1} value will differ from a calculation using a magnetically scaled variable production rate, having a present-day value of 6.1 at/g yr^{-1} for rock surfaces that have been exposed sometime during the past 13 cal ka. However, an integrated mean value of the magnetically scaled production rate will equal 5.1 at/g yr^{-1} for rock surfaces exposed prior to 13 cal ka.

Calculations for the effects of shielding and sample thickness on in situ cosmogenic production were done outside the EXCEL worksheet and a rock density of 2.70 was used in the model calculations.

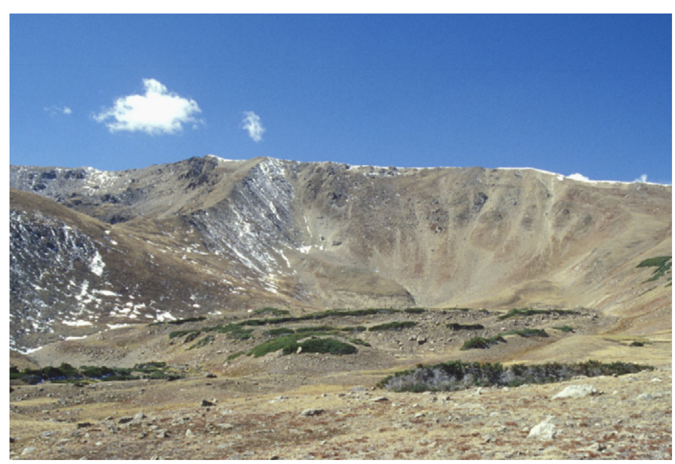

Fig. 4. End moraines at the Butler Gulch site. The height of the headwall above the cirque floor is ~240 m. The Butler Gulch moraine has an average width of ~275 m. Moraine crests ranges 3–9 m above the intervening swales.

4.3. Topographic shielding.

Topographic shielding at the three sites was evaluated by means of topographic maps. The inclination and horizon-

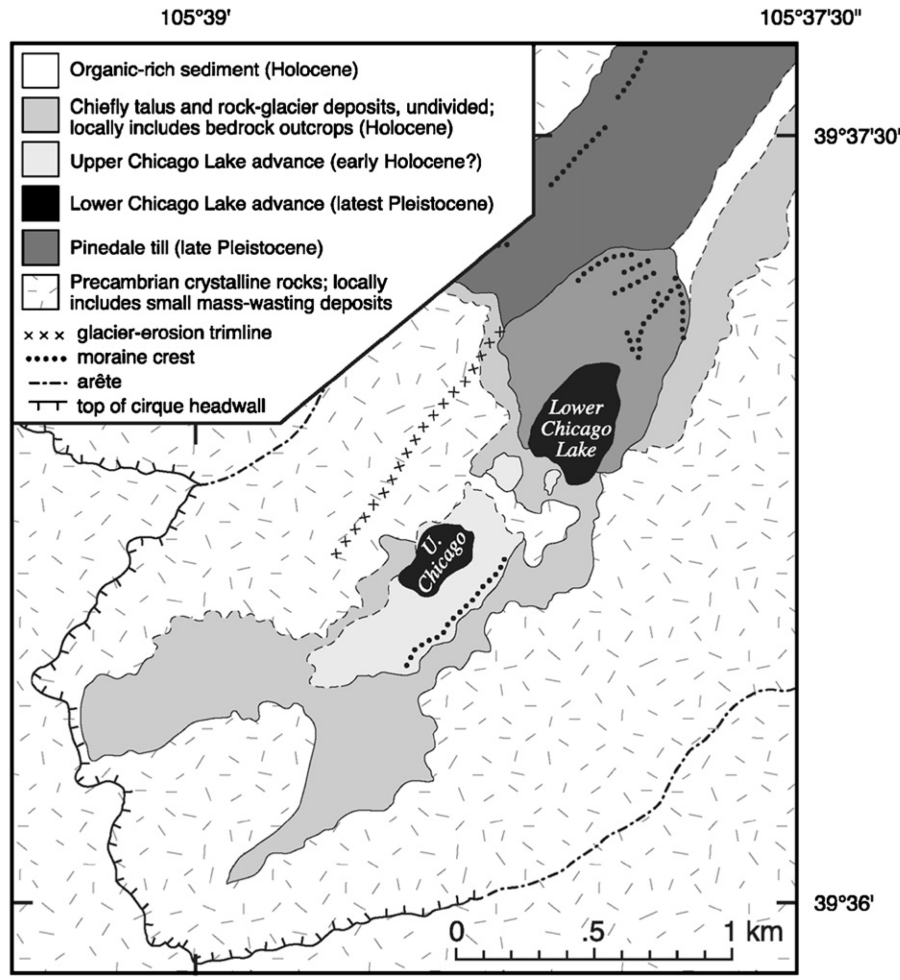

Fig. 5. Surficial geologic map of Chicago Lakes area. Figure was not corrected for radial distortion associated with transfer of topography from aerial photos.

tal distance between the sample sites and nearby summits were measured along radii spaced 30° apart. Topographic shielding was calculated from equations in Dunne et al. (1999). A mean shielding value of 0.982 was obtained for Butler Gulch and Chicago Lakes moraines and a mean value of 0.951 was obtained for the Isabelle Glacier moraine.

4.4. Snow and sediment shielding

Snow and sediment shielding can diminish the measured surface-exposure age by decreasing the amount of cosmogenic radiation that reaches the rock surface; e.g., snow accumulating on end-moraine boulders tends to produce surface-exposure ages that are less than the true exposure interval. However, snow and ice accumulating in a cirque area tend to minimize cosmogenic exposure of the underlying bedrock, thus decreasing inheritance in boulders eventually transported to end moraines.

Three of the four youngest samples at Butler Gulch (BG03-2, -3, -4; Table 1) were located on the distal side of the arcuate early Holocene moraine that extends farthest from the cirque headwall, and the other young sample (BG03-6; Table 1) was located farther downvalley in what may be an even older early Holocene or latest Pleistocene moraine. BG03-3 and -6 are located on the sides of topographic mounds and BG03-4 lies near a topographic low. Therefore, these boulders may have been exhumed by the erosion and BG03-4 may have been covered with snow and ice from time to time. In addition, BG03-2, -3, and -4 are horizontally fractured and may have lost surface material via frost action or spalling over time. Boulder BG03-2 has a surface with a saddle-like shape that may indicate loss or material. For these reasons, the surface-exposure ages of the four Butler Gulch samples (BG03-2, -3, -4, -6) may represent minimum estimates of moraine deposition.

5. Results

5.1. The effect of surface-erosion rates on cosmogenic ages of the Butler Gulch and Chicago Lakes moraines

The surface-exposure ages of north-central Colorado end-moraine boulders are shown in Table 2 without correction for rock-surface erosion rates. Studies of

Table 2
^{10}Be ages of end-moraine boulders from the Isabelle Glacier, Butler Gulch and Chicago Lakes areas

Sample no.	UTM coordinates (NAD27)	Elevation (m)	Height above ground (cm)	Sample thickness (cm)	Cosmogenic age (ka)	1−σ age error (ka)
Isabelle Glacier						
I03–1	13 T 445251 4434782	3663	125	2.0	1.40	0.20
I03–2	13 T 445235 4434769	3663	240	3.0	0.70	0.10
I03–3	13 T 445116 4434712	3663	170	3.0	0.50	0.00
I03-4	13 T 445121 4434706	3663	90	3.0	0.40	0.00
I03-5	13 T 445265 4434769	3663	55	5.0	3.00	0.20
I03-6	13 T 445292 4434787	3663	50	5.0	0.50	0.10
Butler Gulch						
BG03-1	13S 424901 4400407	3663	56	5.0	8.90	0.40
BG03-2	13S 424991 4400417	3664	0–90	4.0	6.80	0.30
BG03-3	13S 425088 4400342	3657	40	2.0	5.00	0.30
BG03-4	13S 425074 4400312	3655	30	5.0	6.10	0.30
BG03-5	13S 425117 4400282	3659	35	3.0	9.40	0.40
BG03-6	13S 425234 4400367	3615	50	3.0	6.60	0.40
BG03-7	13S 424913 4400409	3666	85	2.5	10.20	0.30
BG03-8	13S 424924 4400410	3673	40–70	2.0	9.10	0.40
BG03-9	13S 424948 4400409	3671	30–35	4.0	10.20	0.50
BG03-10	13S 424986 4400346	3662	35–40	5.0	9.50	0.40
BGRG03-1	13S 425908 4400386	3658	30–40	6.0	8.30	0.30
BGRG03-2	13S 426077 4400461	3543	45	4.0	10.20	0.50
Chicago Lakes						
CL03-1	13S 445863 4385533	3476	81	3.0	12.80	0.60
CL03-2	13S 445865 4385509	3476	128	4.0	12.90	0.60
CL03-3	13S 445803 4385434	3500	20–57	4.0	17.50	0.60
CL03-4	13S 445778 4385505	3488	51	3.0	11.00	0.60
CL03-5	13S 445804 4385537	3487	0–85	2.0	11.80	0.60
CL03-6	13S 445752 4835572	3496	71	4.0	12.30	0.70
CL03-7	13S 445719 4385552	3503	125	3.0	12.50	0.60
CL03-8	13S 445644 4385595	3517	0–56	5.0	12.20	0.50
CL03-9	13S 445621 4385463	3512	142	4.0	12.00	0.60

Fig. 6. Butler Gulch boulder BG03-4 with 45-cm-long hammer on top. Note vertical and horizontal fractures in boulder as well as subangular shapes of boulders in picture.

Fig. 7. Subrounded boulders at Chicago Lakes area. Boulder in center foreground has a height of 90 cm.

Table 3
Be-isotope data

Lab no.	Sample	$^{10}Be/^{9}Be$ [1E−12]	Error (%)	Carrier (mg)	Weight (g)	^{10}Be at/g [1E+4]
Isabelle Glacier						
ZB2769	I03-1	0.203	11.1	0.350	47.15	10.05
ZB2770	I03-2	0.149	11.7	0.350	55.26	6.31
ZB2771	I03-3	0.064	9.6	0.350	35.35	4.24
ZB2772	I03-4	0.095	9.6	0.350	55.34	4.01
ZB2773	I03-5	0.269	7.1	0.360	33.95	19.06
ZB2774	I03-6	0.100	15.4	0.350	58.67	4.00
ZB2775	Blank	0.027	20.6			
Butler Gulch						
ZB2762	BG03-1	0.937	4.0	0.350	35.19	62.27
ZB2763	BG03-2	0.683	3.3	0.360	35.31	46.53
ZB2764	BG03-3	0.484	4.5	0.360	35.11	33.16
ZB2765	BG03-4	0.588	4.1	0.360	35.21	40.14
ZB2766	BG03-5	0.983	3.5	0.360	35.25	67.08
ZB2767	BG03-6	0.660	4.9	0.350	35.21	43.85
ZB2768	Blank	0.030	15.2	0.350		
ZB2962	BG03-7	0.888	3.0	0.350	28.27	73.43
ZB2963	BG03-8	0.709	4.4	0.350	25.29	65.60
ZB2964	BG03-9	0.737	4.1	0.350	23.70	72.68
ZB2965	BG03-10	0.938	3.8	0.350	32.65	67.19
ZB2966	BGRG-03-1	0.547	3.3	0.350	22.20	57.59
ZB2967	BGRG-03-2	1.157	5.1	0.350	40.21	67.31
ZB2968	Blank	0.030	13.9	0.350		
Chicago Lakes						
ZB3183	CL03-1	1.467	4.7	0.350	40.83	84.02
ZB3184	CL03-2	0.646	5.0	0.360	18.53	83.83
ZB3185	CL03-3	0.986	3.4	0.350	19.08	120.80
ZB3186	CL03-4	0.730	5.6	0.350	24.06	70.91
ZB3187	CL03-5	0.924	5.4	0.350	27.87	77.52
ZB3188	CL03-6	0.753	5.5	0.344	21.56	80.20
ZB3189	CL03-7	1.071	4.7	0.346	29.79	83.13
ZB3190	CL03-8	0.912	4.2	0.344	26.02	80.53
ZB3191	CL03-9	1.062	10.1	0.340	30.57	79.00
ZB3192	Blank	0.032	17.7			

The results in this table were normalized to a secondary standard (S555) with a nominal value of $^{10}Be/Be = 95.5 \times 10^{-12}$ which itself was normalized to BEST433 (Hoffman et al., 1987). The uncertainty of the standard ($\pm 2.5\%$) is not included in the quoted uncertainty of the samples.

granodiorites in north-central Colorado (Benedict, 1993) have shown that boulder surfaces erode at rates $\leqslant 1$ mm/ka. Applying a 1-mm weathering rate to boulders from north-central Colorado changes their exposure ages only by ~0.1%. Thus, surface weathering (with the exception of spalling) of these end-moraine boulders should not have altered their surface-exposure ages by a substantial amount.

5.2. The potential effect of inheritance on cosmogenic ages of the Isabelle Glacier, Butler Gulch, and Chicago Lakes moraines

In a study of the Pinedale moraines near the downvalley end of Fremont Lake, Wyoming, Gosse et al. (2003) noted that boulder ages decrease from the front of the moraine to the back, and suggested that the first boulders to arrive at the moraine may have had an inherited component because the glacier first eroded regolith that had been exposed to cosmic radiation before glaciation. With respect to our three sites, boulder transport distances (a few hundred meters to a couple of kilometers) are much less than those of boulders in Pinedale end moraines. Thus, inheritance from bedrock cirque walls may be an important issue in the study of the small Front Range glaciers.

Inheritance is obvious when the age of one boulder is much older than the ages of other boulders on the same moraine (e.g., CL03-3 in Table 2). Inheritance, however, is not always easily discerned and there is no simple or completely reliable method of detecting its existence or assessing its magnitude. Indeed, inheritance must be assumed to be present when valley-wall erosion rates are not sufficient to remove it (see Discussion below). If inheritance exists, erosion during glacial transport must remove ~2 m from a boulder to effectively eliminate it (see Fig. 7; Benson et al., 2005).

End-moraine boulders from the Isabelle Glacier and Butler Gulch sites have relatively wide ranges in surface-exposure age (Table 2, Fig. 8) which should not be the case, given the probable duration (a few to several hundred years) of the climatic events leading to the glacial advances. For example, the ages of Isabelle Glacier end-moraine boulders range from 0.4 to 3.0 ^{10}Be ka. The three youngest surface-exposure ages of boulders from the Isabelle Glacier moraine have a relatively narrow range (AD 1500–1600), which is slightly older than the lichen-based age range calculated for glaciers of the LIA Arapaho Peak advance (AD 1650–1850) (Benedict, 1973). If, for example, we accept AD 1650–1850 as the approximate interval of deposition of boulders on the LIA Isabelle Glacial moraine, the inheritance values of the boulders range from essentially 0 to ~3 ^{10}Be ka (Table 2, Fig. 8A). Thus, the surface-exposure ages of boulders from Isabelle Glacier demonstrate that inheritance exists in some boulders derived from glacial systems with short transport distances. The lack of inheritance exhibited by boulders I03-3, -4, and -6 probably indicates that their present-day upper surfaces were originally within the cirque bedrock wall and were shielded by >0.5 m of rock (Table 2).

Under what conditions would boulders derived from valley or cirque walls have little or no inheritance? To answer this question we calculated cosmogenic inheritance values at 0, 1, and 2 m inward from the surface of the Butler Gulch cirque wall (elevation = 3730 m) after 1000 yr of exposure. The slope of the cirque wall (30°) and topographic shielding were considered in the calculation. Because geomagnetic variability was not an issue in this calculation and also because the Lifton et al. (2005) model does not calculate ^{10}Be production with depth, ^{10}Be production as a function of depth was calculated using a model developed by Bill Phillips, University of Edinburgh (Phillips, personal communication, 2003). A ^{10}Be production rate of 5.1 at/g yr^{-1} was used in the calculations which

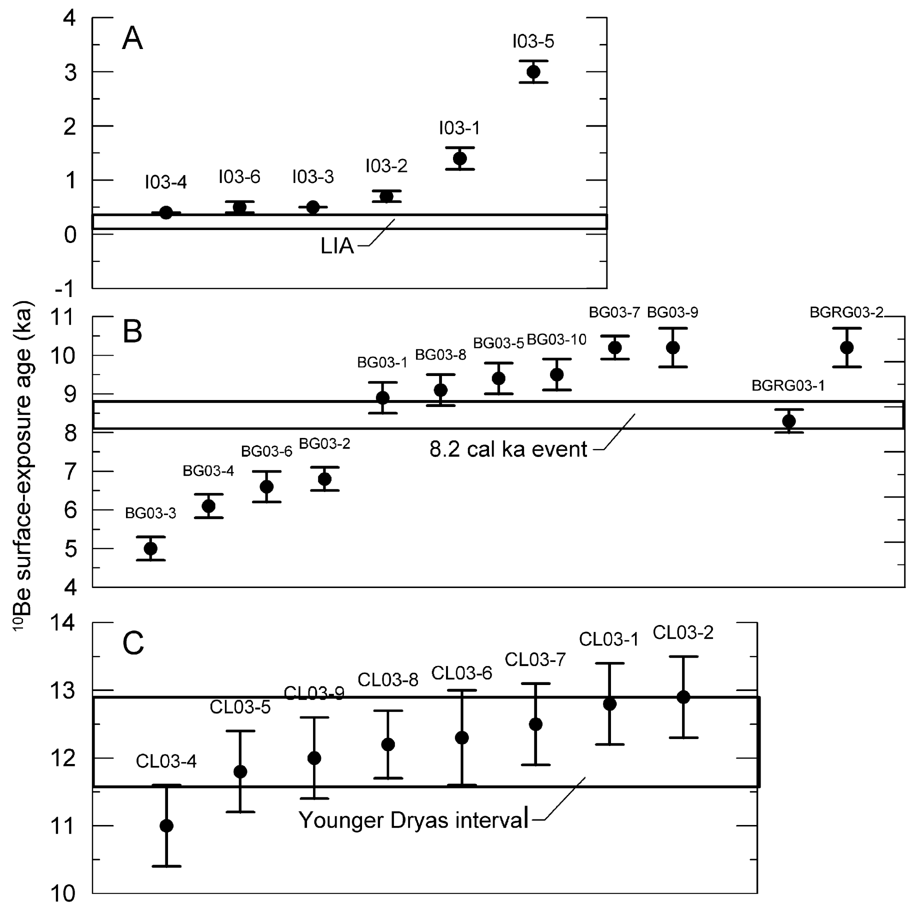

Fig. 8. Surface-exposure ages of end-moraine boulders from (a) Isabelle Glacier, (b) Butler Gulch, and (c) Chicago Lakes. Black circles indicate surface-exposure ages calculated using present-day ^{10}Be production value of 6.2 at/g yr^{-1} at SLHL (Pigati, personal communication) and the EXCEL worksheet from Lifton et al. (2005) which corrects for geomagnetic effects (Pigati and Lifton, 2004). The horizontal rectangles indicate times of specific climate events that may have resulted in the advance of glaciers in the Northern Hemisphere. In this diagram the LIA is assumed to have occurred in the Front Range between AD 1650 and 1850.

show that mean erosion rates must exceed 5 mm/yr in order for a rock surface exposed to the atmosphere to acquire <100 yr of inheritance in 1000 yr (Table 4).

Caine (2001) has shown that granitic alpine cliffs in the upper Green Lakes area of north-central Colorado erode at rates much lower than this value. For the past 25 yr, rockfall from the granodiorite valley walls corresponded to a cliff retreat rate of only 0.02 mm/yr and, during the past ~12 ka, the retreat rate averaged only ~0.25 mm/yr. The latter value is similar to cosmogenic nuclide-derived erosion rates (0.07–0.11 mm/yr) measured in glaciated granitic catchments in the Haslach Basin, Germany (Morel et al., 2003). Triassic limestones in the Bavarian Alps underwent moderately high back-weathering rates (0.15–0.73 mm/yr) during the late-glacial period (Sass and Wollny, 2001); however, these rates, although greater than those undergone by granitic rocks, are still much smaller than the 5 mm/yr rate. Thus, most bulk erosion rates for granitic bedrock are not sufficient to limit inheritance to <100 yr, and measured rates, ranging from ~0.1–0.2 mm/yr, yield a surface-inheritance value of ~900 yr and 1-m-deep inheritance value of ~160 yr (Table 4).

Table 4
The effect of erosion rate on cosmogenic inheritance

Erosion rate (mm/yr)	Cosmogenic inheritance (yr) after 1000 yr		
	0 m	1 m	2 m
10.0	57	10	2
5.0	113	19	3
1.0	469	80	14
0.1	901	157	27

^{10}Be production rate at SLHL = 5.1 at/gSiO$_2$ yr^{-1}.
^{10}Be production rate at 3730 m corrected for slope and distant shielding = 62.4 at/g yr^{-1}.
Distances are relative to exposed rock surface.

Only rockfall events in which several meters of wall rock frequently (>5m/1000 yr) topple to the basin floor produce material that lacks substantial inheritance. Even those events produce some boulders that have appreciable inheritance by virtue of their original position on or near the valley- or cirque-wall surface. The frequency of such boulder production events is not known for our sites

but is not expected to be high, given the observations of Caine (2001).

5.3. Surface-exposure ages of the Butler Gulch and Chicago Lakes moraines

The subangular boulders composing the Butler Gulch end moraines were derived via rockfall from the cirque wall and were transported only a few hundred meters downvalley (Fig. 4). The end-moraines are not sharp crested but they do reach heights of 3–9 m above intervening swales. The Butler Gulch boulders exhibit a wide range of surface-exposure ages, which suggests that processes such as cosmogenic inheritance, shielding, and spalling may have occurred, otherwise the Butler Gulch Glacier would appear to have been active for 5.2 ^{10}Be ka when probably it was active for, only at most, a few hundred years.

We expect that the two rock-glacier boulders (BGR03-1 and -2) would have lost little or no material before deposition. Their ^{10}Be ages, which, therefore, include inheritance, span the ^{10}Be age range of the six oldest Butler Gulch end-moraine boulders (Table 2, Fig. 8B), suggesting that at least some of the older Butler Gulch end-moraine boulders have an inherited cosmogenic component.

We suggest the hypothesis that the Butler Gulch moraine may have formed during the 8.2 cal ka event (Fig. 8B). The 8.2 cal ka event was caused by outburst floods from Lake Agassiz and Lake Ojibway that actually occurred between 8.75 and 8.2 cal ka (Barber et al., 1999; Ellison et al., 2006). If this hypothesis is correct, then four of the Butler Gulch boulders have one-sigma values younger than the 8.2 cal ka event and four of the boulders have one-sigma values greater than the 8.2 cal ka event. This suggests that the boulders with ages >8.75 ^{10}Be ka have inheritance values ranging from ~0.75 to ~1.5 ^{10}Be ka and that the four boulders with ages younger than 8.2 cal ka may have experienced surface spalling or cosmogenic shielding by sediment and(or) snow (Fig. 8B). The latter conjecture is consistent with the fractured nature of some of the boulders and their position on the landscape (see Section 4.4 and Fig. 6). If inheritance is not a factor in the values of the six oldest Butler Gulch boulders, and the four youngest ages were affected by shielding or spalling, the Butler Gulch moraine would appear to have been deposited between 8.90 ± 0.5 and 10.2 ± 0.6 ^{10}Be ka (Fig. 8, Table 2). Because all of the six one-sigma ages fail to overlap (e.g., BG03-1does not overlap BG03-7), the distribution of ages indicate that the glacier responsible for the formation of the Butler Gulch moraine would have had to exist for ~1.3 ^{10}Be ka. We know of no study that suggests the existence of such a long-lived early Holocene climate event that would have affected the Front Range.

Eight of nine Chicago Lakes boulders have ages that fall within the Younger Dryas interval (Alley et al., 1993). End-moraine boulders from the Chicago Lakes area are located about 2 km downvalley from the Chicago Lakes cirque headwall; thus, the boulder transport distance was considerably greater than at Butler Gulch, suggesting that some subglacial erosion and rounding of boulders may have occurred. However, the Chicago Lakes boulders are subrounded (Fig. 7) in contradistinction to subangular end-moraine boulders deposited by latest Pleistocene advances of the same magnitude that occurred elsewhere (e.g., Gosse et al., 1995; Clark and Gillespie, 1997), suggesting that a different process led to their rounding.

We speculate that most of the rounding of Chicago Lakes boulders occurred as the result of spheroidal weathering on high valley walls that extend above the level reached by Pleistocene glaciers and that this process occurred before mass movement that delivered the boulders to the valley floor.

Another possibility is that intense weathering of valley-wall bedrock above the limit of Pinedale glaciation in the Chicago Lakes valley preconditioned boulders to rounding during glacial transport. The brief interval (about 1–2 ka) between recession of the Pinedale glacier and advance of Younger Dryas glaciers may account for the minimal amount of inheritance believed to be associated with Titcomb and Chicago Lakes end-moraine boulders. In addition, bedrock in glaciated valleys may have remained covered with snow and ice much of the time between the onset of the Pinedale glaciation and the termination of the Younger Dryas glaciation further reducing the cosmogenic exposure of the bedrock.

6. Discussion

Previous studies in the Front Range have indicated a Satanta Peak glacial advance that may have occurred during the Younger Dryas interval (Benedict, 1973, 1981, 1985; Menounos and Reasoner, 1997). Evidence for glaciation before 13.5 cal ka in Arapaho Creek valley and before 13.8 cal ka at Sky Pond can be argued to represent the existence of an early Satanta Peak glacial advance which occurred slightly before the recession (13.1 cal ka) of the Sierran Recess Peak Glacier (Clark and Gillespie, 1997).

End-moraine boulders at the Isabelle Glacier site were derived from the headwall and were transported a few hundred meters across the cirque from the north facing part of the headwall. End-moraine boulders at the Butler Gulch site mostly were derived from cirque-wall rockfall and were transported only a few hundred meters downvalley, and end-moraine boulders at the Chicago Lakes site were mostly derived from rockfall that may have experienced spheroidal weathering on valley wall(s). Surface-exposure ages of Butler Gulch end-moraine boulders are neither precise nor accurate enough to assign definite times of moraine formation. However, we offer the hypothesis that the Butler Gulch advance occurred during the 8.2 cal ka event and that cosmogenic inheritance, rock-surface spalling, and sediment/snow shielding accounts for the scatter of ages outside the time interval occupied by the

8.2 cal ka event. This implies that inheritance values up to ~1.5 ^{10}Be ka exist in Butler Gulch boulders (Fig. 8B) which is consistent with inheritance values measured on boulders from the LIA Isabelle Glacier moraine.

The Chicago Lakes Glacier appears to have advanced during the Younger Dryas interval/late Satanta Peak event. Therefore, we consider it equivalent in time to the Titcomb Lakes advance.

A question remains as to why surface-exposure ages of Titcomb Lakes and Kromer end moraines appear largely unaffected by cosmogenic inheritance (Gosse et al., 1995; Kerschner et al., 2006; i.e., why was cosmogenic inheritance minimized at these sites when it had a substantial effect at the Butler Gulch site?

Small cirque glaciers exist today in both the Titcomb basin, western Wyoming and in the Kromer valley, Austria (Gosse et al., 1995; Kerschner et al., 2006). Glaciers may have existed in these areas immediately prior to the 8.2 cal ka and Younger Dryas intervals. If so, winter snow and perennial ice in the cirques could have shielded the underlying bedrock from cosmic radiation, minimizing the accumulation of cosmogenic inheritance. On the other hand, most of the end-moraine boulders in the Butler Gulch area may have been derived from rocks falling from cirque walls that were exposed to perennial cosmogenic radiation. Cirque Glaciers do not exist at the Butler Gulch site today and they may not have existed in this area immediately prior to the early Holocene glacial advance. Thus, snow and ice may not have shielded bedrock derived from cirque walls that were transported downvalley during the Butler Gulch Glacier advances.

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