LATE PLEISTOCENE GLACIATION AND DEGLACIATION IN THE CRESTONE PEAKS
AREA, COLORADO SANGRE DE CRISTO MOUNTAINS, USA – CHRONOLOGY AND
PALEOCLIMATE

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1 Abstract

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3 Cosmogenic ^{10}Be surface-exposure dating and numerical glacier modeling are used
4 to reconstruct glacial chronology and climate in the Colorado Sangre de Cristo
5 Mountains during the local last glacial maximum (LLGM) and the subsequent
6 deglaciation. Twenty-two surface-exposure ages on moraine boulders and polished-
7 bedrock outcrops in the Willow Creek valley and ten in two adjacent valleys indicate
8 that glaciers were at or near their maxima from $\sim 21\text{ka}$ until $17\text{-}16\text{ ka}$, and then
9 retreated rapidly, nearly deglaciating the Willow Creek valley entirely by $\sim 14\text{ ka}$.
10 Coupled energy/mass-balance and flow modeling of two of the glaciers indicates
11 that, if changing ice extent was driven only by temperature and insolation changes,
12 temperature depressions of 5.0 and 5.1°C from modern conditions, with an
13 uncertainty of approximately $\frac{+1.5^\circ}{-1.0}^\circ\text{C}$, would have sustained the glaciers in mass-
14 balance equilibrium at their LLGM extents. Doubling or halving of modern
15 precipitation during the LLGM would have been associated with $2.7\text{-}3.0^\circ\text{C}$ and 6.9-
16 7.0°C temperature depression respectively. Approximately half of the subsequent
17 LLGM-to-modern climate change was accomplished by $\sim 14\text{ ka}$. If the rapid main
18 phase of deglaciation between about 16 ka and 14 ka was driven solely by
19 temperature and insolation changes, it would have been associated with a
20 temperature rise of about 2.5°C , at a mean rate of approximately 1.1°C/ky . This new
21 chronology of the last glaciation is generally consistent with others developed
22 recently in the Colorado Rocky Mountains. The numerical modeling,

however, suggests a lesser LLGM temperature depression from modern conditions than have most previous studies in Colorado.

1. Introduction

Records of alpine glaciation have long been viewed as sensitive indicators of past climate change. Until recently, attempts to glean quantitative paleoclimatic information from evidence of past alpine-glacier extent have involved primarily the assessment of paleo snowlines (or paleo equilibrium lines) and their implications for climate changes (e.g., Porter, 1964, 1975; Andrews et al., 1970; and many subsequent studies). Dating of former alpine glaciation relied on a combination of radiocarbon dating of fortuitously preserved organic matter associated with glacial deposits, and relative-dating methods, based primarily on topographic evolution of depositional features, weathering, and soil development (e.g., Burke and Birkeland, 1979). Over the last two decades, two developments have revolutionized the use of paleoglaciology to assess past climate change. First, surface-exposure dating using cosmogenic radionuclides has made it possible to date glacial deposits and erosional features directly, commonly with higher precision and accuracy than was previously possible (Ivy-Ochs and Briner, 2014). Second, advances in numerical modeling have made possible more physically based investigations of the relationship between climate change and past glacier extent. Recent work by Ward et al. (2009) and Dühnforth and Anderson (2011) in the San Juan Mountains of southwestern Colorado and the Front Range of northern Colorado used a combination of

cosmogenic exposure dating of moraine boulders and upvalley polished bedrock and numerical glacier modeling to document the timing of the last glacial maximum and the pacing of deglaciation, and magnitudes of equilibrium-line change and climate change associated with the local last glacial maximum (LLGM) and the subsequent deglaciation. In this study, we take a similar approach in the Sangre de Cristo Mountains in south-central Colorado. We present thirty-two new ^{10}Be cosmogenic-radionuclide exposure ages on moraine boulders and glacially polished bedrock outcrops that constrain the timing of the LLGM and subsequent deglaciation. These are the first numerical chronological constraints on glaciation in the range. We use a more climatically explicit modeling approach than that used in the earlier studies, to assess the precipitation and temperature change combinations that could have sustained glaciers at their LLGM extents, and the timing, magnitudes, and rates of climate change associated with the subsequent deglaciation. In addition, to facilitate comparison of the timing of the LLGM and of deglaciation in the Sangre de Cristo to earlier work in the Colorado Rocky Mountains, we recalculate a large number of previously reported ^{10}Be ages using newer, more accurate, production rates and scaling.

2. Study area

The northern portion of the Colorado Sangre de Cristo Mountains is a narrow (~15 km wide), nearly linear, NNW-SSE-trending range in south-central Colorado (Fig. 1). The range is bounded on the west by the San Luis Valley, the two separated by an

active Neogene-Quaternary fault system. The Wet Mountain Valley bounds the range on the east, across both older, latest Mesozoic to Paleogene, and Neogene-Quaternary structures (Lindsey et al., 1983). Peaks along the range crest approach or exceed 4000 m. Blanca Peak, at 4372 m, is the highest point in the range. Valley floors on the two sides of the range are at about 2300-2500 m, with broad bajadas abutting both range fronts. The current study area, the "Crestone Peaks area" of the range (Fig. 2), contains at least six summits above 4240 m. Local relief in the Crestone Peaks area commonly approaches 1 km over horizontal distances of 2 km or less.

Bedrock geology of the northern portion of the range is dominated by folded and thrust-faulted upper Paleozoic sedimentary rocks, with more limited outcrops of Proterozoic crystalline and lower Paleozoic sedimentary rocks (Lindsey et al., 1984, 1985, 1986). The Crestone Peaks area is dominated by the Crestone Conglomerate Member of the Pennsylvanian-Permian Sangre de Cristo Formation, which consists of well-cemented arkosic conglomerate, sandstone, and minor siltstone. The resultant very hard rock produces steep cliffs and preserves abundant glacial polish and striae.

Valleys draining both sides of the range show evidence of Pleistocene glaciation including cirques, U-shaped valleys, and abundant glacially sculpted and polished landforms. On the east side of the range, where glaciers generally reached the mountain front, latero-terminal moraine complexes are typically well preserved. On

the west side late Pleistocene glaciers generally did not reach the mountain front and moraine complexes are typically less well preserved. In the Crestone Peaks area, however, latero-terminal moraines are well preserved on both sides of the range. Three of the valleys containing these moraines, the Willow and South Crestone Creek valleys on the west side of the range and the South Colony Creek valley in the east side, are the subject of this study (Fig. 2). Late Pleistocene glaciers in these valleys descended steeply from cirque basins at about 3600-3800 m to termini at 2950-3100m on the west side of the range and about 2750m on the east side. During their last glacial maximum stands, Willow Creek glacier had a length of 6.2 km and an area of about 8.3 km², South Crestone Creek glacier a length of 4.9 km and an area of about 5.1 km², South Colony Creek glacier a length of 10.3km and an area of about 15 km². There are no present-day glaciers in the Colorado Sangre de Cristo although snowfields remain through most summers in a few protected high-altitude locations in all three valleys.

The Colorado Sangre de Cristo experiences a continental climate, with large diurnal and seasonal temperature ranges and relatively little precipitation. Mean annual precipitation ranges from about 350 mm and 500mm at the western and eastern mountain fronts respectively, to about 1000 mm along higher portions of the range crest (<http://www.prism.oregonstate.edu/>). Refsnider et al. (2009) reviewed the climate of the range using data from 18 stations, including COOP (National Weather Service Cooperative Network) meteorological stations and SNOTEL (Snowpack telemetry) stations. The remainder of this paragraph is drawn from their discussion.

Both summer and winter temperatures are largely controlled by altitude, with summer (June-August) lapse rate more than twice as steep as winter (October-April) lapse rate ($-8.7^{\circ}\text{C}/\text{km}$ vs. $-4.2^{\circ}\text{C}/\text{km}$, respectively). Precipitation generally varies with altitude, with mean annual precipitation decreasing by about $0.46\text{ m}/\text{km}$ on the west side of the range and $0.57\text{ m}/\text{km}$ on the east, but precipitation distribution and seasonality are complex, with multiple moisture sources affecting different altitudes and the two sides of the range differently. Stations at altitudes above about 3000 m on both sides of the range experience two subequal precipitation peaks in late winter/early spring (March-April) and in summer (July-August). At lower elevations on both sides of the range the late-winter/early-spring precipitation peak is weaker and the summer precipitation peak is dominant. During the winter months, precipitation is strongly (positively) correlated with altitude, reflecting orographic control on precipitation from cyclonic storms. During the summer, when precipitation results primarily from monsoon-related convection, the altitude/precipitation relationship is much weaker. Overall, winter precipitation is somewhat higher at a given altitude on the east side of the range than on the west side, an observation that Refsnider et al. (2009) attribute to (1) the presence of the topographic barrier of the massive San Juan Mountains to the west, which blocks westerly cyclonic moisture flow to the Sangre de Cristo and (2) the occurrence of southeasterly sourced moisture, particularly during the late winter and spring, associated with cyclonic storms that pass across northern Arizona and New Mexico.

3. Previous work on glaciation of the Crestone Peaks area

The first description of glacial deposits in the Crestone Peaks area of the Sangre de Cristo dates from the early 20th century (Siebenthal, 1907). Systematic mapping in the area by Scott and Taylor (1974), McCalpin (1982), and Lindsey et al. (1985, 1986) identified deposits from two glacial advances. On the basis of weathering and soil characteristics those workers correlated the advances to the Pinedale Glaciation (Last Glacial Maximum) and Bull Lake Glaciation (penultimate glaciation) in the Wind River Range of Wyoming. Prior to the current study, no numerical ages were available for glacial deposits anywhere in the Colorado Sangre de Cristo.

In earlier work, several of the current authors and their collaborators attempted to draw paleoclimatic inferences from the extent of Pleistocene glaciers in the Colorado Sangre de Cristo, using a variety of methods. Refsnider et al. (2009) determined LLGM ELAs throughout the range, using accumulation area ratio (AAR) methods. In the absence of modern glaciers, however, they were unable to determine the magnitude of ELA depression. Using a combination of modern climate data and LLGM ELAs from throughout the Colorado portion of the range they concluded that LLGM moisture sources and precipitation patterns in the range were similar to modern patterns and, assuming no change in total precipitation or precipitation seasonality, that temperatures would need to have been 6–7.5°C colder than modern to sustain LLGM glaciers. Brugger et al. (2009) applied both ELA-based methods and a degree-day mass-balance model to deduce climate from

reconstructed LLGM glaciers in the Blanca massif ~40 km south-southeast of the present study area (Fig. 1), concluding that a 7.0–8.0°C temperature depression would have been necessary to sustain LLGM glaciers there without an accompanying change in precipitation.

4. Methods

This study focuses on Willow and South Crestone Creek valleys on the west side of the Sangre de Cristo range, where cosmogenic ^{10}Be surface-exposure dating was used to develop a detailed chronology of the LLGM and subsequent deglaciation, and numerical glacier modeling was used to assess the climate during the LLGM and deglaciation. The chronology of the last glaciation also includes a number of surface-exposure ages from moraines in the South Colony Creek valley on the east side.

4.1 Cosmogenic ^{10}Be surface-exposure dating

4.1.1 Sampling methods

Samples were collected for cosmogenic ^{10}Be surface-exposure dating from moraine boulders in all three drainages (Fig. 3a). Multiple closely spaced latero-terminal moraines are present in each valley, and on the basis of their well-preserved morphology and lack of evidence of deep weathering, appear to have been deposited during the LLGM. In Willow and South Colony Creek valleys, boulders

were sampled from both the outermost and the innermost moraines of the apparent LLGM complex. In South Crestone Creek valley, only the outermost moraine was sampled. Samples were taken from large boulders, firmly rooted in moraine material, on low-slope portions of moraine crests. Most samples were from boulders standing at least 1m above the moraine surface, although a few were collected on smaller boulders, standing as little as 50cm above the surface. Sampling was limited to near-horizontal portions of boulder tops, and where possible, to surfaces with preserved glacial polish. In Willow Creek valley, almost all moraine boulder samples were taken from areas of, or immediately adjacent to, preserved polish. In the South Crestone and South Colony drainages, however, moraine boulders were more weathered, with little, if any, polish preserved, necessitating sampling from areas without polish. All sampled boulders were composed of the well-cemented Crestone Conglomerate Member of the Sangre de Cristo Formation.

Glacially polished bedrock outcrops, also of the Crestone Conglomerate, were sampled upvalley from the LLGM moraine complex in the Willow Creek drainage (Fig. 3b). Samples were collected from the crests of non-vegetated, generally wind-swept, whalebacks and small roches moutonnées. Most samples were taken from features that stood at least several meters above surrounding areas of the valley floor, although in a few cases, particularly in the uppermost portions of the valley, samples were taken from the crest of small features with as little as 50 cm of relief. Greater local relief likely minimized potential snow shielding of samples, but may have increased the possibility of inherited preglacial ^{10}Be inventories. In areas of

lesser local relief the opposite was likely the case – greater potential for snow or soil shielding, but less potential for inheritance.

4.1.2 Sample processing

Physical and chemical processing of samples was completed at the CosmogenicNuclide Preparation Lab at the State University of New York (SUNY) Geneseo following methods of Laabs et al. (2013), which are modified from Kohl and Nishiizumi (1992), Ditchburn and Whitehead (1994), Bierman et al. (2002), and Munroe et al. (2006). Quartz grains were separated from crushed and milled samples using hand magnets, froth flotation (Clifton et al., 2005), and chemical etching in dilute acids. Samples were spiked with 300-350 µg of a commercially made beryllium carrier and dissolved in hydrofluoric acid. Procedural and carrier blanks were prepared using the same mass of beryllium carrier added to samples. The beryllium fraction of each sample was chemically isolated and prepared into targets for $^{10}\text{Be}/^9\text{Be}$ measurement by accelerator mass spectrometry (AMS) at the Purdue University Rare Isotope Measurement Laboratory (Muzikar et al., 2003).

The $^{10}\text{Be}/^9\text{Be}$ of procedural blanks ranged from 5.6×10^{-15} to 7.0×10^{-15} and all sample $^{10}\text{Be}/^9\text{Be}$ ratios were 3.41×10^{-13} or greater, indicating that more than 98% of measured ^{10}Be in all samples was produced in situ. Blank corrections were applied to the $^{10}\text{Be}/^9\text{Be}$ of all samples prior to computing ^{10}Be concentrations.

4.1.3 Calculation of ^{10}Be surface-exposure ages

Quantification of the sea-level, high-latitude spallogenic production rate of in situ cosmogenic ^{10}Be has improved significantly in recent years due to the increasing number of sites where production has been calibrated (e.g., Borchers et al., 2016). Additionally, improvements in sample-preparation methods and AMS measurement have likely afforded more reliable calibrations of *in situ* production. Such improvements suggest that the globally averaged spallogenic reference production rate of Balco et al. (2008), which has been used to compute cosmogenic exposure ages in numerous previous applications of ^{10}Be surface-exposure dating in the Rocky Mountains, is too high by as much as 14% (Heyman, 2014), thereby yielding exposure ages that are too young. A recently published calibration of ^{10}Be production by Lifton et al. (2015) reports the in situ ^{10}Be inventory of multiple independently dated surfaces near 2000 m asl at Promontory Point in the Lake Bonneville basin of northern Utah, ~750 km northwest of the present study area. The age, latitude, and altitude of these surfaces are similar to the sample sites in the Sangre de Cristo Mountains; therefore, the Lifton et al. (2015) calibration constitutes an appropriate production model for computing ^{10}Be surface-exposure ages in this study.

In addition to this regional production calibration, a relatively new production-rate scaling model of Lifton et al. (2014) provides a physically based alternative to the CRONUS-Earth models. This model, termed the “LSD scaling scheme” (for Lifton-

Sato-Dunai), represents the most robust consideration of the effects of geomagnetic latitude, atmospheric depth, and time on the spallogenic and muonic production of in situ ^{10}Be . As discussed by Phillips et al. (2016), this model has the advantage of having a stronger physical basis for computing the production of ^{10}Be . The model was used with a reference production rate of $4.0 \text{ atoms g}^{-1} \text{ yr}^{-1}$ in a recent compilation of recalculated ^{10}Be surface-exposure ages of moraines and glacially polished bedrock in the western U.S. (Shakun et al., 2015).

We report ages computed using both the Promontory Point calibration (Lifton et al., 2015) with the time-dependent "Lm" scaling model implemented in the CRONUS-Earth online exposure age calculator version 2.3 (Balco et al., 2008), and the LSD scaling model (Lifton et al., 2014) as used by Shakun et al. (2015). The two models yield statistically overlapping ages. For simplicity, only exposure ages computed with the Promontory Point/Lm ("PPLm") model are discussed in the text. All individual exposure ages are reported with 1σ external uncertainty (measurement and production rate error) and assume constant exposure at each sample site; i.e., zero erosion and/or shielding of sample surfaces. The majority of samples were from boulders and outcrops with preserved glacial polish, indicating that very little or no postglacial weathering had occurred, and those samples without polish and with evidence of weathering showed no systematic bias towards younger ages.

Analysis of data from the South Colony SNOTEL station (<http://wcc.sc.egov.usda.gov/nwcc/site?sitenum=773>) suggests that at wind-protected sites, snow shielding might result in underestimation of true ages by as

much as 10% in the highest-elevation portions of the study area. However, as noted in section 4.1.1, in selecting upvalley sample sites, we favored promontories, which are likely windswept, reducing potential snow-shielding error.

For moraines with three or more exposure ages, the mean $\pm 1 \sigma$ of the ages is reported as the best estimate of the moraine age. Because cosmogenic exposure ages of moraine boulders can display considerable variability in cases where the boulders have experienced differing exposure histories (due to surface erosion, degradation of the moraine crest, or variable shielding by snow or sediment), the reduced chi-squared statistic (χ^2_r) is reported as an indicator of the degree of variability among individual exposure ages (c.f., Rood et al., 2011).

4.2 Paleoglacier modeling

To assess changes from modern temperature and precipitation (defined in our modeling as 1971-2000 AD mean values) that could have sustained steady-state glaciers at their mapped late Pleistocene extents in the Willow and South Crestone Creek drainages, we employed a numerical energy/mass-balance and flow model developed by Plummer and Phillips (2003). The modeling couples a two-dimensional energy/mass-balance model with a two-dimensional model of ice flow in the horizontal plane to compute the extent and thickness of a glacier at mass-balance equilibrium with input climate variables. This approach first determines the mass balance across a digital elevation model of a glacial valley using a set of

monthly energy-balance calculations, and then uses the resultant annual mass-balance grid as a distributed source term for the flow model. The flow model computes the area and thickness of ice based on a finite-difference solution to linearized equations for ice flow by deformation and sliding. Simulations were run to assess the year-round temperature depression that would have been associated with a given change in precipitation—modeled as a spatially and seasonally uniform percent change in monthly precipitation—over a range of precipitation values of 50% to 200% of modern. We did not alter precipitation seasonality from its modern pattern, nor did we change temperature lapse rates or the ratios of precipitation at different elevations from modern conditions. We also used the model to estimate the amount and rate of temperature and/or precipitation change necessary to deglacialate the Willow Creek valley following the LLGM, making use of moraine-boulder and polished-bedrock CRN ages to establish the timing of deglaciation.

Full descriptions of the modeling procedure and earlier modifications are in Plummer and Phillips (2003), Laabs et al. (2006), Refsnider et al. (2008), and Leonard et al. (2014). Here, we describe the overall procedure and discuss modifications made in this study. All modeling was done at 30 m resolution.

4.2.1 Energy- and mass-balance model

The energy/mass balance model computes an annual-net-balance grid of an area of interest, including water-equivalent snow accumulation in areas of positive mass

balance and water-equivalent potential ice melt in areas of potentially negative mass balance. Primary inputs to the model are digital topography and modern monthly climate data, including values and altitudinal gradients of mean temperature and precipitation, standard deviation of daily mean temperature, relative humidity, wind speed (varying with altitude), and cloud cover. Temperature and precipitation gradients were determined from linear regressions of 30 arc-second PRISM (Parameter-elevation Regressions on Independent Slopes Model—<http://www.prism.oregonstate.edu/>) grids of 1971-2000 AD climate means against altitude, cropped to an area of ~98 km² including and surrounding the modeled domains, extending from the range divide to below the Pleistocene termini of the two modeled glaciers. Other climate input came from sources listed in Table 1. Shortwave-radiation calculations in the model account for orbital variations using ages from our CRN-based chronology (reported in section 5.1). Basic model parameters are given in Table 2. Monthly energy and mass flux values are calculated at the ground/snow/ice surface, including longwave radiation, shortwave radiation, sensible heat, latent heat, conductive, and advective fluxes, snow accumulation, and ablation due to melt and sublimation. The calculations are described in detail in Plummer and Phillips (2003).

The model accounts for the redistribution of snow by avalanching but not by wind. The routine for redistributing snow by avalanching has not been previously reported and is described here. The goal of the routine is to redistribute snow across a sloping surface as would occur by avalanching. The calculations effectively

move snow from steep slopes onto down-gradient slopes of lower angle and then disperse snow in those accumulation areas to represent the effect of gradual energy dissipation. The routine includes several adjustable parameters that affect the amount of snow removed and the dispersion behavior. As used in this study, avalanches occur in the model when snow water-equivalent depth reaches 0.1 m on slopes of 30° or greater. Avalanche events can occur up to three times a month and each event removes 30% of the snowpack. Modeled avalanches shift snowdown slope to slopes of less than 30°, at which point the snow is deposited according to a Gaussian smoothing function that distributes most of the mass within a two-pixel (60 m) radial distance. Because each downslope transport event carries snow to the first down-gradient cell with slope less than a critical value, successive passes use different critical slope angles to further disperse the accumulated snow.

4.2.2 Flow model

The ice-flow model combines the net annual mass-balance grid with expressions of ice flux by deformation and sliding to compute the steady-state ice area and thickness, using computational procedures described in Plummer and Phillips (2003). The continuity equation for ice flow,

$$\frac{\partial H}{\partial t} - \left(\frac{\partial q}{\partial x} + \frac{\partial q}{\partial y} \right) - b = 0 \quad (\text{Eq. 1})$$

where H is the ice thickness, t is time, q is the vertically integrated ice flux in the horizontal (xy) plane, and b is the net mass-balance term provided by the energy- and mass-balance calculations, is solved by finite differences. Fluxes in the continuity equation are based on velocities associated with internal deformation, u_d , and sliding, u_s :

$$u = u_d + u_s = \frac{2A}{n+2} \tau^n H + B \tau^m \quad (\text{Eq. 2})$$

where u is total velocity, A and B are flow- and sliding-law coefficients respectively, n and m are flow- and sliding-law exponents, respectively, and τ is the basal shear stress. Following Jarosch et al. (2013), we incorporate a flux-limiting scheme into the model. The specified ice-flow parameters are constant and uniform in the model domain.

The computational procedure described here assumes that the shallow-ice approximation (Hutter, 1983) is appropriate for the modeled paleoglaciers. Although this approach may be less accurate than a full-Stokes procedure at modeling flow processes in steep glacial valleys (Le Meur et al., 2004), comparison studies have demonstrated that the two approaches yield similar ice extents and thicknesses (Le Meur and Vincent, 2003; Leysinger Vieli and Gudmundsson, 2007). Because our goal is to reconstruct steady-state glacier geometry, for which the mass balance is a dominant constraint, the shallow-ice approximation should provide a sufficiently accurate model of glacier response for inferring past climate conditions.

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391 Following Laabs et al. (2006) and Leonard et al. (2014), flow modeling was initially
392 conducted using two different sets of deformation and sliding parameters; (1) a
393 “stiff-ice” set— $n = 3$, $m = 2$, $A = 5 \times 10^{-8} \text{ yr}^{-1} \text{ kPa}^{-3}$, $B = 7.5 \times 10^{-4} \text{ myr}^{-1} \text{ kPa}^{-2}$
394 (Plummer and Phillips, 2003; Laabs et al., 2006; Refsnider et al., 2008) and (2) a
395 “soft-ice” set— $n = 3$, $m = 2$, $A = 2.5 \times 10^{-7} \text{ yr}^{-1} \text{ kPa}^{-3}$, $B = 3.75 \times 10^{-3} \text{ myr}^{-1} \text{ kPa}^{-2}$
396 (Laabs et al., 2006, Leonard et al., 2014). These values of the A and B coefficients are
397 the values used in the model calculations. They are not directly comparable to the
398 values cited in some previous publications, including Plummer and Phillips (2003),
399 Laabs et al. (2006), and Refsnider et al. (2008), which reflected an earlier approach
400 apportioning glacier motion between internal deformation and basal sliding. In the
401 Sangre de Cristo, the “soft-ice” parameters produced a closer fit to the preserved
402 moraines and upvalley geomorphic indicators of ice thickness than did the “stiff-ice”
403 parameters, and consequently were used in final calculations. Adjustments to these
404 parameters impact overall ice distribution only to the extent that changes in ice
405 thickness alter the pattern of flow. The distribution of ice thickness and flow pattern
406 changed relatively little over the modeled range of the parameters. The use of “soft-
407 ice” rather than “stiff-ice” parameters resulted in a slight increase in the amount of
408 temperature depression from modern conditions necessary to produce a glacier of a
409 specific extent. The magnitude of this increase varied somewhat from glacier to
410 glacier and with the amount of precipitation change from modern used in the
411 simulation. For models run with no change in precipitation the use of “soft-ice”
412 rather than “stiff-ice” parameters increased the amount of necessary temperature

depression by 0.1-0.2°C. Although the “soft-ice” parameters appear to provide a better fit to paleoglacier geometries, we consider the impact of ice rheology assumptions in our error analysis.

4.2.3 Model validation

There are no modern glaciers or mapped permanent snowfields in the study area against which to test model results. However, examination of aerial photography indicates that typically a few small snowfields persist through the ablation season at high altitudes in both Willow and South Crestone Creek drainages. We compared the extent of modeled snowfields generated using modern (1971-2000 AD) climate input and modern orbital parameters to the areas of snowfields in five available sets of late summer/early fall (August 26-October 12) air photos from the same time interval (<http://earthexplorer.usgs.gov/>). When run with no change from modern temperature or precipitation, the model produced areas of net annual snow accumulations significantly larger than the observed snowfields. A modeled temperature increase of about 1.2 °C resulted in snowfields of approximately the same extent as the observed snowfields (Fig. 4). We discuss the implications of this discrepancy between modeled and observed snowfields for our paleoglacier modeling results and their associated uncertainties in section 6.2. The modeled snowfields are located somewhat downslope from the observed snowfields, suggesting that snow accumulates and remains on slightly steeper slopes (>30°) than assumed in our modeling. While this difference is important in the location of

the small, avalanche-fed snowfields, it is not likely to be significant in modeling the much larger Pleistocene glaciers.

5. Results

5.1 Glacier Chronology

Twenty-one boulders from LLGM moraines were analyzed for cosmogenic ^{10}Be surface-exposure ages in the three valleys—eleven in Willow Creek valley, three in South Crestone Creek valley, and seven in South Colony Creek valley—to assess timing of the LLGM. In addition, eleven samples from glacially polished bedrock surfaces in the upper Willow Creek valley were analyzed to constrain the timing and rate of post-LLGM deglaciation. ^{10}Be sample data and ages are shown in Table 3.

Three of the four moraines from which we obtained at least three ages display a less-than-expected dispersion of ages about the mean, indicated by a χ^2_r of less than unity, and by the narrow range of 1σ relative to external uncertainty of individual exposure ages. The alignment of exposure ages from these moraines validates the assumption that differences in exposure history among sampled erratic boulders are negligible.

5.1.1 Local Last Glacial Maximum

5.1.1.1 Willow Creek

The LLGM latero-terminal moraine complex at Willow Creek includes several moraine crests that splay from the outermost LLGM moraine downvalley from the point at which the glacier emerged from the steep mountain front (Fig.5a). A distinct and continuous innermost splay impounds a postglacial meadow and terminates about 350 m upvalley from outermost moraine. Between these two, several intermediate moraine crests splay from the outermost crest and are better preserved on the right-lateral (north) side of the valley than on the left-lateral (south) side. Dated boulders on the outermost right-lateral moraine clustered into two groups spatially and in terms of exposure age (Figs.5a and 6). Three boulders on the most distal portion, close to the terminus and downvalley from any splays, yielded ages ranging from 17.9 ± 0.9 ka to 18.5 ± 0.8 ka, with a mean age of 18.2 ± 0.3 ka. Further upvalley along the crest of this moraine, in the region where multiple crests splay off it, four sampled boulders yielded uniformly younger ages ranging from 15.7 ± 0.8 ka to 16.9 ± 0.8 ka, with a mean of 16.1 ± 0.5 ka. Two boulders on the outer left-lateral moraine across the valley yielded similar ages (15.9 ± 0.8 ka and 16.7 ± 0.9 ka), as did two boulders on the inner right-lateral moraine (15.7 ± 0.7 ka and 16.7 ± 0.8 ka). Taken together, the six ages on the proximal portion of the outer lateral moraine on the two sides of the valley yield a mean age of 16.2 ± 0.5 ka.

5.1.1.2 South Crestone Creek

The latero-terminal moraine complex of the LLGM South Crestone Creek paleoglacier consists of multiple moraine crests spread over a downvalley distance of about 450 m. We sampled three boulders from the terminal portion of the outermost crest (Figs.5a and 6), and obtained ages ranging from 17.4 ± 0.8 ka to 20.7 ± 1.1 ka, with a mean of 19.6 ± 1.6 ka.

5.1.1.3 South Colony Creek

We sampled boulders from the South Colony Creek moraine complex in an area where large inner and outer moraine crests were previously interpreted by Lindsay et al. (1986) as having been deposited during the maximum stands of the Pinedale (LLGM) and the Bull Lake Glaciations, respectively (Figs.5b and 6). All of the samples yielded exposure ages from the last glaciation. Four boulders on the inner left-lateral moraine yielded ages ranging from 18.1 ± 0.8 ka to 19.3 ± 0.9 ka, with a mean age of 18.7 ± 0.5 ka. Two on the outer left-lateral moraine yielded ages of 17.7 ± 0.9 ka and 21.5 ± 1.0 ka. The right-lateral moraine, mapped as Bull Lake age by Lindsay et al. (1986), yielded a single age of 18.0 ± 0.9 ka, which overlaps with ages on both left-lateral moraines.

5.1.2 Post-LLGM recession

In the Willow Creek valley, eleven samples were dated from polished bedrock surfaces upvalley from the LLGM moraine complex, at five locations between about

35% (~2.2 km) and almost 90% (~5.6 km) of the distance upvalley from the LLGM terminus to the cirque headwall (Figs. 5c and 7). Two or more samples, generally separated by a minimum of several tens of meters, were dated at each location to provide a check on individual ages. With the exception of a single outcrop (discussed in the following paragraph) that shows strong evidence of an inherited ^{10}Be inventory, paired ages differ by no more than 1.4 ky and all overlap within uncertainty (Table 3).

The most downvalley three bedrock samples, all from a large mid-valley whaleback ~2.2 km upvalley from the outermost LLGM moraine, yielded ages (26.1 ± 1.2 ka, 32.8 ± 1.8 ka, 32.9 ± 2.1 ka) significantly older than any of the boulder ages from the terminal moraine complex in either this valley or the two adjacent valleys. These older ages strongly suggest presence of an inherited ^{10}Be inventory accumulated during a prior interval of exposure and preserved because of limited erosion during the last glaciation. Excluding this outcrop there is a general decrease in sample age from the LLGM moraines to the most upvalley samples (Fig. 7). There is, however, a reversal of expected ages between the third most-upvalley samples (14.3 ± 0.7 ka and 15.6 ± 0.8 ka; ~3.5 km upvalley from the LLGM terminus) and the next pair of samples farther upvalley (16.3 ± 0.8 ka and 17.1 ± 0.8 ka; ~4.5 km upvalley from the LLGM terminus). The latter site has ages indistinguishable from the moraine-boulder ages far down valley, suggesting limited ^{10}Be inheritance at that site as well.

5.2 Glacier Modeling

Coupled mass/energy-balance and flow models were run for the LLGM extents of the South Crestone and Willow Creek paleoglaciers using insolation values for 21 and 18 ka, respectively, reflecting the approximate timing of the maximum extent of each glacier (see discussion in section 6.1 below). Insolation values for 14 ka were used to model deglaciation of the uppermost CRN sample sites in the Willow Creek drainage. The direct effect of changing insolation is quite small. For example, simulations run on South Crestone Creek paleoglacier with 18 ka insolation require slightly more temperature depression than those run with 21 ka insolation, but the difference is about 0.06°C. Models were run to steady-state conditions at each ice extent on the assumption that glacier response times were short in comparison to the timescale of climate changes we hope to model. Approximate response times were calculated for the modeled glaciers equilibrated with modern precipitation amounts, from modeled ice thicknesses and ablation using the equation:

$$t_r = \frac{H_{max}}{\dot{a}_0} \quad (\text{Eq. 3})$$

where t_r is response time, H_{max} is maximum ice thickness and $\dot{a}_0$ is net ablation rate at the terminus. We used maximum ice thickness (Cuffey and Paterson, 2010) rather than “characteristic” ice thickness (Jóhannesson, 1989) because it is more easily determined and because it yields a more conservative result (i.e., longer response time). Calculated response times for the modeled LLGM Willow Creek glacier (H_{max} =

164 m; $\dot{a}_0 = 4.6$ m/yr) and South CrestoneCreek glacier($H_{max}= 119$ m; $\dot{a}_0 = 3.0$ m/yr) are 36 and 40 years respectively. Response time for the much smaller ~ 14 ka modeled glacier in Willow Creek valley ($H_{max}= 67$ m; $\dot{a}_0 = 1.7$ m/yr)is 39 years.This approach to determining response times is strictly appropriate only for small mass-balanceperturbations. We assessed glacier response to larger perturbations by imposing a step temperature increase of 2.5°C , with no change in precipitation, on the modeled LLGM Willow Creek glacier, causing the ice margin to retreat far upvalley.The change in glacier area from the LLGM steady-state to a new steady-state area equilibrated with the imposed higher temperature was 98% complete in 82 model years. While there is considerable uncertainty in these calculations, these results indicate that the response time of the glaciers is much shorter than the millennial-scale climate changes of interest here, and thus that use of steady-state modeling is appropriate.

Using “soft-ice” flow and sliding parameters described in section 4.2.2, with no change from modern (1971-2000 AD) precipitation amounts and seasonality, model simulations indicate that a temperature depression from modern conditions of about 5.0°C would have been necessary to sustain a glacier in mass-balance and flow equilibrium at the LLGM extent of Willow Creek glacier (Fig.8a). A nearly identical 5.1°C depression would have been necessary to sustain the LLGM South Crestone Creek glacier (Fig.8b). A halving or doubling of year-round precipitation for Willow Creek glacier would have been associated with LLGM temperature depressions of 7.0°C and 2.7°C respectively. For South Crestone Creek glacier the

necessary temperature depressions would have been 6.9°C and 3.0°C. Fig.9a plots necessary LLGM temperature depressions for four precipitation scenarios (0.5-, 1-, 1.5-, and 2-times modern precipitation) considered for each of the glaciers. Following Laabs et al. (2006), 2nd-order polynomial best-fit lines are used to predict other temperature- and precipitation-change combinations that could have sustained each glacier at its LLGM extent. With no change from modern precipitation, the modeled LLGM equilibrium-line altitude would have been ~3580 m in the Willow Creek drainage and ~3575 m in the South Crestone Creek drainage. Refsnider et al. (2009) previously estimated LLGM equilibrium-line altitudes for the two valleys as 3707 m and 3556 m respectively, using accumulation-area ratio methods.

The model was also run to determine temperature and/or precipitation changes from the modern that could have sustained the Willow Creek glacier with a terminus immediately upvalley of the two most upvalley CRN polished-bedrock sample sites (with ¹⁰Be exposure ages of approximately 14.0ka; Figs.5c and 7). Model simulations indicate that with no change from modern precipitation, a temperature depression of about 2.5°C would have been necessary to sustain a glacier at this extent (Fig.10). Temperature-depression values necessary to sustain the 14.0ka glacier over a range of precipitation from 50% to 200% of modern are shown in Fig.9b along with the 18.2 ka (LLGM) temperature/precipitation combinations from the same valley. With no change from modern precipitation, the modeled 14.0ka equilibrium-line altitude in the upper Willow Creek drainage would

be ~3800 m. This relatively low altitude may reflect strong topographic shading and avalanche accumulation on the small remnant glacier, more than regional climate.

6. Discussion

6.1 Chronology

Cosmogenic exposure ages of moraines from the three valleys are plotted in Figs. 5 and 6. The mean of individual exposure ages ($\pm 1\sigma$) is considered the best estimate of the time when ice abandoned each moraine in cases where the χ^2_r statistic is close to or less than unity, which is observed for the moraines in Willow and South Colony valleys. In South Crestone valley, exposure ages are more dispersed and therefore the oldest exposure ages may best represent the time when this moraine was abandoned (cf., Applegate et al., 2012).

We use a combination of moraine abandonment ages and upvalley polished-bedrock exposure ages to infer a deglaciation chronology for the Willow Creek valley.

Previous studies in the Rocky Mountains have used polished-bedrock ages in this way (Gosse et al., 1995; Guido et al., 2007; Ward et al. 2009; Dühnforth and Anderson, 2011; Young et al., 2011), although there are some potential problems with the approach. Several of the studies (Guido et al., 2007; Ward et al. 2009; Dühnforth and Anderson, 2011) identified the presence of inherited pre LLGM ^{10}Be inventories in some samples of glacially scoured bedrock. We encountered the same

problem at two of our upvalley sample sites. The deglaciation chronology and modeling discussed below is based largely on the mean exposure age of 16.2 ± 0.5 ka for the LLGM moraine in Willow Creek valley and the furthest upvalley pair of bedrock exposure ages, which show no evidence of inheritance. Ward et al. (2009) have also pointed out that readvances during an overall period of recession may complicate interpretation of deglaciation history from bedrock ages. Because there are no mapped moraines upvalley of the terminal moraines in Willow Creek valley that might indicate a pause or readvance of the ice margin during overall deglaciation, we must assume some pattern of ice retreat between the time when the terminal moraine was abandoned (16.2 ka) and the time when ice retreated upvalley of the sites yielding the youngest bedrock exposure ages (14.0 ka). We assume that, although the ice margin likely fluctuated during overall retreat, the upvalley surface was not uncovered by ice until 14.0 ka and remained ice-free thereafter. Computing a mean rate of ice retreat and inferring attendant changes in climate based on this assumption excludes the possible effects of multiple intervals of ice cover/uncover of the surface or erosion during a (possible) later interval of ice cover. The short duration of this time interval minimizes the possibility of (1) a great depth of bedrock erosion by glacial scouring during any post-14.0 ka readvance and (2) the impact of multiple intervals of ice cover on the in situ ^{10}Be inventory.

The oldest boulder ages from South Crestone and South Colony valleys indicate that glaciers were at or near their LLGM extents at about 21 ka. While there is no direct

evidence of the extent of the Willow Creek glacier at that time, it was probably also close to its LLGM position, although it appears to have attained its maximum stand as many as 2000-2500 years later. A strong clustering of exposure ages suggests that the glaciers in Willow and South Colony valleys began to retreat from their maximum or near-maximum positions between 19 and 18 ka. The six ages on the proximal section of the outermost lateral moraines at Willow Creek (mean = 16.2 ± 0.5 ka), however, indicate that the glacier either persisted at, or readvanced to, close to its maximum extent and then retreated again from that position shortly after 16.2 ka. Upvalley polished-bedrock ages (Figs. 5c and 7) show a general decrease in age from the young cluster of outermost moraine ages to the farthest upvalley pair of ages and suggest that after 16.2 ka the Willow Creek glacier receded rapidly and by about 14 ka the valley floor was almost completely deglaciated. Evidence of an inherited in situ ^{10}Be inventory at some sample sites and the possibility of snow shielding prevent more detailed interpretation of the pattern and timing of deglaciation from bedrock exposure ages.

In order to compare this Sangre de Cristo chronology with other late Pleistocene glacial chronologies in Colorado, we recalculated previously reported ages with the new Promontory Point production rate (Lifton et al., 2015) and L_m scaling model (Balco et al., 2008) used to calculate the “PPLm” ages reported in the current study. The recalculated ages are presented below, along with the originally published ages given in *italics* to facilitate comparison to the earlier publications.

666 The Sangre de Cristo chronology is very similar to that documented in the Sawatch
 667 Range, approximately 125 km to the northwest, and generally similar to that in the
 668 Front Range slightly more than 200 km to the north (Fig. 1). Young et al. (2011)
 669 dated moraine boulders and polished-bedrock outcrops in three adjacent valleys in
 670 the central Sawatch Range. They report scattered exposure ages on boulders on the
 671 outermost LLGM moraines of two of the glaciers, in the Pine Creek and Clear Creek
 672 valleys, with age clusters of 23.6 ± 1.4 ka, 20.5 ± 0.2 ka, and 16.9 ± 0.4 ka (*original ages*
 673 22.4 ± 1.4 ka, 19.1 ± 0.2 ka, and 15.8 ± 0.4 ka). More recent work by Schweinsberg et al.
 674 (2016; computed using the same PPLm production scheme but with version 2.2 of
 675 the CRONUS-Earth calculator) in neighboring Lake Creek valley, tightly constrains
 676 the age of the outermost two LLGM moraines there to 21.8 ± 0.3 ka and 21.7 ± 0.4
 677 ka. The youngest moraine-boulder ages and upvalley polished-bedrock ages from the
 678 three valleys indicate rapid recession from near-LLGM positions beginning between
 679 17 and 16 ka (*original ages 16 ka and 15 ka*), with first upvalley ages suggesting
 680 significant ice retreat had occurred by the 16.7 ± 0.4 ka to 15.3 ± 0.3 ka (*original ages*
 681 15.6 ± 0.4 to 14.1 ± 0.3 ka; Young et al., 2011). The timing of full deglaciation in the
 682 Sawatch Range was not established by Young et al. (2011), but their furthest
 683 upvalley age indicates that the Lake Creek valley was about 80% deglaciated by
 684 14.3 ± 0.3 ka (*original age 13.2 $\pm$ 0.3 ka*). Young et al. (2011) argue that the timing of
 685 this main phase of deglaciation is consistent throughout the western United States,
 686 an argument supported by the Sangre de Cristo exposure ages reported here. In the
 687 southern Sawatch Range, Brugger (2007) dated five boulders from LLGM moraines
 688 of the Taylor River glacier, yielding a mean LLGM ^{10}Be age of 20.8 ± 1.9 ka (*original*

689 *mean age 19.5±1.8 ka*). Brugger (2007) also reported ^{36}Cl ages for three of the
 690 boulders, but in view of uncertainties in ^{36}Cl production rates, we do not include
 691 those ages in this discussion. Brugger (2007) attributed two younger ^{10}Be ages with
 692 a mean of 16.2 ± 0.6 (*original mean age 14.8±0.8 ka*) on the ice-proximal slope of the
 693 LLGM moraine in nearby Texas Creek to surface instability during a long interval of
 694 ice stagnation following the maximum stand.
 695
 696 Based on a combination of ^{10}Be , ^{26}Al , and ^{36}Cl surface-exposure ages on moraine
 697 boulders, Ward et al. (2009) concluded that Middle Boulder Creek glacier in the
 698 Colorado Front Range began to retreat from its LLGM moraines at about 20 ka, an
 699 age which might be closer to 21 ka based on a recalculated ^{10}Be age of 20.8 ± 2.9 ka
 700 (*original age 18.2±2.9 ka*) and recalculated ^{26}Al ages of 20.3 ± 1.3 ka and 22.5 ± 1.4 ka
 701 (*original ages 19.0±1.8 and 22.7±2.0*; Schildgen, 2000). Again, for reasons mentioned
 702 above, we do not include two ^{36}Cl ages of Benson et al. (2004, 2005) in this
 703 discussion. A recalculated initial upvalley polished-bedrock ^{10}Be age of 19.8 ± 2.6 ka
 704 (*original age 17.7±2.7 ka*) indicates that by that time the ice margin had receded
 705 significantly (~5 km or 30-35% of the glacier's total length) from its LLGM position,
 706 suggesting significant deglaciation may have begun somewhat earlier there than in
 707 the Sawatch Range (Young et al., 2011) and Sangre de Cristo (this study). Six
 708 clustered bedrock samples from the upper portion of the Middle Boulder Creek
 709 valley reported by Ward et al. (2009) yield recalculated ^{10}Be ages ranging from
 710 14.2 ± 0.7 ka to 14.8 ± 0.7 ka (*original ages 13.2±1.2 ka to 13.8±1.3 ka*). These
 711 ages indicate that the valley was 85-90% deglaciated by 15-14 ka, very close to the

712 timing of nearly complete deglaciation in the Sawatch and Sangre de Cristo. In
713 adjacent North Boulder Creek, Dühnforth and Anderson (2011) report ages of 23.6–
714 18.3 ka on LLGM moraine boulders, with major deglaciation completed by about 12
715 ka. Recalculated ages range from 24.9 ± 1.5 ka to 19.5 ± 1.1 ka for the moraine
716 boulders, suggesting initiation of deglaciation by about 19.5 ka, also earlier than in
717 the Sangre de Cristo and Sawatch ranges. The recalculated ages, combined with
718 Dühnforth and Anderson's (2011) Fig. 5, indicate that the valley was about 65%
719 deglaciated by about 15.5 ka and about 80% deglaciated by about 13 ka.

720
721 Available chronologies of the LLGM and the deglacial interval in the eastern
722 Colorado ranges (Sangre de Cristo, Sawatch, Front), then, are broadly consistent
723 with one another, with one possible difference. In all three ranges, LLGM extent was
724 reached by about 21 ka, and near-LLGM extent persisted for several thousand years.
725 In the Sangre de Cristo and Sawatch ranges, major recession from LLGM moraine
726 complexes did not begin until 17–16 ka. It may have begun about 2000 years earlier
727 in the Front Range. In all three ranges, extensive deglaciation had occurred by about
728 14 ka, although small glaciers likely persisted in currently glacier-free drainages for
729 some time after 14 ka.

731 6.2 Glacier Modeling Results and Paleoclimate

732
733 Our numerical-model simulations indicate that Willow Creek and South Crestone
734 Creek glaciers could have existed in steady state at their LLGM geometries with a

temperature depression of 5.0-5.1°C from modern with no change from modern precipitation amount, seasonality, or altitudinal gradient, or with any other of the suite of temperature/precipitation combinations indicated in Fig.9a. To use these model results to infer climate during the LLGM, however, we must consider several possible sources of error.

First, our modeling has treated LLGM climate reconstruction as a two-variable problem—only year-round temperature and spatially (altitudinally) and temporally (month-to-month) uniform changes in precipitation were modeled. Possible changes in temperature and precipitation seasonality were not considered, nor were changes in the temperature lapse rate or the ratio of precipitation between high and low elevations. Changes in cloud cover, wind speed, and relative humidity were not considered. In this study we did not undertake a systematic analysis of the effects of plausible changes in these variables. Sensitivity analyses by Plummer and Phillips (2003) concluded that the last three climatic variables mentioned above result in a pooled uncertainty of approximately $\pm 0.5^{\circ}\text{C}$ in model results.

A second potential source of error is related to the fact that our model simulation of semi-permanent snowfields under modern climate produced larger-than-observed snowfields, with the observed snowfields best simulated when the model was run with a temperature increase of 1.2°C from modern conditions (Fig. 4). This suggests that model parameters used to determine various energy fluxes may be incorrect and that the model might also underestimate the temperature depression necessary

to establish steady-state LLGM glacier conditions by about 1.2°C. However, the difficulty in assessing avalanche and wind redistribution of snow and the importance of those factors in determining the location and extent of the very small modern semi-permanent snowfields make us reluctant to adjust inferred temperature-depression values for the LLGM to account for this mismatch. The fact that the record of observed snowfields is limited to only five years of late summer photographs suggests additional caution. Instead of applying a blanket correction to our paleoglacier model results, we pool this uncertainty with other potential sources of error to arrive at overall error estimates for our climate reconstructions.

Third, Anderson et al. (2014) have argued that it is inappropriate to consider the outermost LLGM moraines of a glacier to be indicative of long-term LLGM climate, arguing that glacier extent is a function of both long-term climate and interannual variability of temperature and precipitation. Using a linear glacier model (Roe and O'Neal, 2009) to assess the effect of interannual variability on glacier extent through a modeled 4000-year LLGM interval of constant climate, Anderson et al. (2014) found that for LLGM valley glaciers in the Colorado Front Range, modeled mean glacier lengths, inferred to reflect mean LLGM climate, were 10-15% less than modeled maximum glacier lengths. On this basis they argued that the use of outermost LLGM moraines as indicators of mean LLGM climate would result in overestimation of temperature depression and/or precipitation enhancement.

In a companion study being prepared for publication by several of the current author the issues raised by Anderson et al. (2014) are examined in detail. Here, we summarize the results of that study and their implications for our modeling results in the Sangre de Cristo. The approach of the companion study is similar to that of Anderson et al. (2014), with three main differences. First, it uses a revised model of glacier response to interannual forcing (Roe and Baker, 2014), which includes a better representation of glacier response than the model (Roe and O'Neal, 2009) used by Anderson et al. (2014). Second, it considers the impact of interannual variability both with an assumed constant long-term climate (Anderson et al., 2014) and with a climate that includes millennial-scale, low-amplitude temperature and precipitation oscillations. Third, it assesses how much change in long-term temperature would be necessary to produce a change in maximum glacier-length excursion equivalent to that produced by interannual variability. This calculation allows a quantification of the error that interannual variability imparts to paleoclimate estimates. The companion study simulated glacier length changes over a 7500-year-duration LLGM (Clark et al., 2009), using the mean model parameters of Anderson et al. (2014) coupled with paleoglacier geometries and modern climate input from the current study area. This modeling indicates that interannual variability is likely to result in an overestimation of LLGM temperature depression of 0.30-0.36 °C for the Sangre de Cristo paleoglaciers under the assumption of a constant long-term climate. The overestimation is half of that amount or less if small millennial-scale oscillations in temperature ($\pm 0.5^{\circ}\text{C}$) and precipitation ($\pm 5\%$) are introduced. Pooling uncertainties in model input values and character of LLGM-

interval climate, the companion study finds that the effect of interannual variability on the length of the two paleoglaciers could result in anything from a slight *underestimation* of associated temperature depression to a nearly 0.6°C *overestimation*. Given this range it is difficult to estimate how much error in the amount of LLGM temperature depression or precipitation change from the modern is likely to result from the use of the outermost moraine to characterize LLGM climate. In our error assessment, we take a conservative approach, assuming a potential overestimation of 0.6°C, the maximum value from pooled errors of our sensitivity analysis.

The second and third sources of error discussed above (model overestimation of the extent of snowfields under modern conditions and possible misrepresentation of the magnitude of climate change due to use of the outermost moraines to characterize mean LLGM climate) result in potential errors in opposite directions. Because the magnitude of each error source is uncertain, we have pooled them with our best assessment of the potential error related to other sources discussed above, including those related to uncertainty in estimates of paleo cloud cover, wind, and humidity, and uncertainty related to glacier rheology. Summation of these quantifiable errors in quadrature results in potential uncertainties of +1.3°C and -0.8°C (indicating that our results may *underestimate* temperature depression by up to 1.3°C, or *overestimate* it by up to 0.8°C). Additional, more difficult to quantify, error sources likely raise overall uncertainty. Consequently, we adjust our overall potential error estimate to approximately $\pm 1.5^{\circ}\text{C}$.

6.2.1 LLGM climate

Our model results suggest that the LLGM glaciers in the Crestone Peaks area were sustained with a lesser magnitude climate deviation from modern than suggested by most previous paleoglaciological studies in Colorado. Based on analysis of modern conditions at late Pleistocene glacier equilibrium lines in the Colorado Sangre de Cristo, Refsnider et al. (2009) concluded that, assuming no change from modern precipitation, a temperature depression of about 6°C would have been necessary to sustain glaciers on the western side of the range. This is slightly greater than the 5.0-5.1 $^{+1.5}_{-1.0}$ °C cooling indicated for the same side of the range in the current study. Refsnider et al. (2009) found that absent change in precipitation, a greater temperature depression, of about 7.5°C, would have been necessary to sustain LLGM glaciers on the eastern side. This intrarange difference may reflect changing precipitation patterns rather than actual differences in temperature depression. Utilizing both ELA analysis and degree-day modeling, Brugger et al. (2009) concluded that, absent change in precipitation, sustaining the LLGM glaciers in the Blanca massif, ~40 km to the south-southeast in the Sangre de Cristo (Fig. 1), would have required a temperature depression of 7.0–8.0°C, again with somewhat more temperature depression necessary of the eastern side of the range than on the west.

Some paleoglacier-based climate studies in other ranges in Colorado have also indicated a greater LLGM temperature depression than do our model results for

Willow and South Crestone Creek paleoglaciers. An early, state-wide, ELA-based assessment by Leonard (1989), which did not include the Sangre de Cristo Mountains, indicated that on average an 8.5°C LLGM temperature depression would have been necessary, absent any change in precipitation. More recent work by Brugger (2010) examining 23 paleoglaciers in the southern Sawatch and Elk ranges (Fig. 1), using both an ELA-based approach and a degree-day glacier model, indicated that a temperature depression of $6.9 \pm 0.6^\circ\text{C}$ would have been necessary to sustain LLGM glaciers there with no precipitation change. By contrast, in the Colorado Front Range (Fig. 1), Dühnforth and Anderson (2011) using a combination of degree-day modeling and ELA analysis, estimated that the LLGM North Boulder Creek glacier could have been sustained with a 4.5–6.0°C temperature depression with little or no change in precipitation, consistent with results reported here.

At present it is unclear whether the range of LLGM temperature-depression values from paleoglacier studies completed to date in the Colorado Rocky Mountains reflects actual spatial differences in temperature or precipitation change, or if it more strongly reflects differences in glaciological methods. It is also possible that we have somewhat underestimated the LLGM temperature depression in the Sangre de Cristo in this study, in view of the larger positive potential error in our temperature depression estimate, due primarily to model overprediction of modern snowfield extent (sections 4.2.3 and 6.2). Work-in-progress applying the current modeling approach more broadly across Colorado and adjacent states should yield insight into these questions.

872

873 In the above discussion we present our results in terms of coupled changes in
874 temperature and precipitation, as our modeling does not allow us to separate the
875 effects of these two mass-balance forcings. Very few non-glacial proxy records in the
876 Colorado mountains extend significantly back into the LLGM interval identified in
877 this study (~21-16 ka) to help us resolve this issue. To our knowledge, only an early
878 pollen study in the northern Front Range (Fig. 1) by Legg and Baker (1980)
879 provided quantitative estimates of LLGM climate conditions—suggesting a
880 depression of July mean daily maximum temperature of about 5 °C and an increase
881 of about 15 cm (~20%) in mean annual precipitation compared to present values.
882 Regional proxy records indicate that during this interval effective precipitation was
883 higher than present in the Great Basin and the U.S. Southwest and lower than
884 present in the Northwest and northern Rockies (Thompson et al., 1993; Asmerom et
885 al., 2010; Oster et al., 2015). In a regional analysis of proxy records in the western
886 United States, Colorado falls near the boundary between drier-than-present regimes
887 to the north and wetter-than-present regimes to the south and west (Oster et al.,
888 2015). A possible approach to solving the “two-variable problem” is to use output
889 from numerical climate models applied to the LGM to provide an initial estimate of
890 temperature depression, and then to solve for precipitation using the glacier model
891 results shown in Fig. 9a. This approach is premised on the
892 observation that temperature output from climate models is generally more accurate
893 than precipitation output (Flato et al., 2013). An ensemble mean of nine PMIP3
894 (Paleoclimate Modelling Intercomparison Project Phase III) LGM (21 ka) climate

models suggests that at that time, temperature on the western slope of the Sangre de Cristo near the field area was depressed by $\sim 8.8 \pm 2.7^\circ\text{C}$ compared to the modern (Oster et al., 2015, Supplementary Table S-9). Using our glacier-model output, a $>8^\circ\text{C}$ LLGM temperature depression would have been associated with less than 50% modern precipitation.

6.2.2 Climate change during deglaciation

Nearly complete deglaciation of Willow Creek valley occurred between about 18.2 ka and 14.0 ka. At about 16.2 ka, the glacier was almost as extensive as at 18.2 ka (Fig. 5a), and the temperature and/or precipitation changes from modern conditions necessary to sustain the 16.2 ka glacier would have been essentially the same as those shown in Fig. 9a for the 18.2 ka LLGM glacier. Thus, virtually all of the LLGM-to-14.0 ka climate-change forcing must have taken place after about 16.2 ka. If precipitation remained at modern levels throughout the deglaciation interval, about half of the total post-LLGM temperature rise would have occurred by 14.0 ka. Temperature would have increased by about 2.5°C in the approximately 2200 years following 16.2 ka, at a mean rate of about $1.1^\circ\text{C}/1000\text{ years}$. The amount and rate of temperature change implied by these differing ice extents is, of course, a function of the assumed lack of change in precipitation. If, for example, precipitation was higher than modern during the LLGM and lower than modern during deglaciation, the amount and rate of

temperature increase from 16.2 ka to 14.0ka would have been less than that just suggested.

A small number of non-glacial proxy records in Colorado span the time interval of deglaciation of the Willow Creek valley, including a lacustrine sediment and isotope record from San Luis Lake at ~2350 m asl on the floor of the San Luis Valley (Yuan et al., 2013), about 40 km south-southwest of the present study area, and pollen records from the Gunnison Basin (Fall, 1997; Briles et al. 2012) 100-150 km to the northwest (Fig. 1). The lacustrine record indicates that moisture conditions in the San Luis Valley fluctuated significantly through the Sangre de Cristo deglaciation interval, with relatively dry conditions from 16.5 ka to 15.7 ka, wetter conditions from 15.7 ka to 14.9 ka, and a return to dry conditions after 14.9 ka, persisting until ~12.7 ka. The pollen record (Briles et al., 2012) also suggests a transition to drier conditions after about 14.7 ka. These studies do not, however, provide quantitative estimates of precipitation change that could be used to limit the range of possible deglaciation temperature/precipitation combinations shown in figure 9b.

Our inference from model results that approximately half of the post-LLGM climate change took place during the interval of rapid deglaciation ending about 14ka, the remainder after that time, is similar to the conclusions of Dühnforth and Anderson (2011) in the Front Range. They found that at the North Boulder Creek glacier 350-500 m of ELA rise took place between the LLGM and ~13ka (*original age ~12 ka*), another 250-300m of rise occurred thereafter.

7. Conclusions

Moraine-boulder and polished-bedrock ^{10}Be surface-exposure ages in three valleys in the Crestone Peaks area of the Colorado Sangre de Cristo Mountains indicate that glaciers were at or close to their LLGM extent between about 21 ka and 16.2ka, and subsequently retreated rapidly to nearly deglaciade at least the Willow Creek valley by about 14.0 ka. Assuming that precipitation rates and seasonality were the same as modern, maximum LLGM glacier extents would have been associated with temperatures 5.0-5.1°C colder than modern, with an uncertainty of about $\frac{+1.5}{-1.0}$ °C. A doubling or halving of precipitation compared to modern would indicate LLGM temperature depressions of 2.7-3.0°C and 6.9–7.0°C respectively. About half of the LLGM-to-modern climate change occurred during rapid ice recession between approximately 16.2ka and 14.0 ka. If recession was driven exclusively by climate warming and insolation changes, this would imply a mean rate of temperature increase of about 1.1°C/1000 years during its major phase.

The new chronology for the Sangre de Cristo Mountains is consistent with other recently developed chronologies in the Colorado Rocky Mountains (Ward et al., 2009; Young et al., 2011; Dühnforth and Anderson, 2011; Schweinsberg et al., 2016), although deglaciation from the LLGM in the Front Range may have begun somewhat earlier than in either the Sawatch Range or the Sangre de Cristo. The paleoclimate estimates developed here are consistent with those of some recent work in Colorado

963 (Dühnforth and Anderson, 2011), but indicate smaller temperature depressions
964 than do most other studies (Leonard, 1989; Refsnider et al., 2009; Brugger, 2010).
965 The extent to which this reflects geographic differences in the magnitude and
966 character of climate change, possibly changes in regional precipitation pattern in
967 addition to temperature differences, and the extent to which it is an artifact of the
968 different approaches used to evaluate paleoclimate from glacier extent and their
969 uncertainties, is not known. Continued development of both the approaches taken in
970 this study and other paleoglaciological methods will help to answer these questions.

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984

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Figure captions

1 – Digital terrain model of Colorado showing the location of the Sangre de Cristo Mountains and other ranges and study sites mentioned in the text. Crestone Peaks study area is indicated by small box. GB—Gunnison Basin; SLL—San Luis Lake.

2 –Digital terrain model of the Crestone Peaks area of the Sangre de Cristo, identifying the valleys studied. Heavy dotted lines indicate positions of outermost local last glacial maximum (LLGM) moraines in each valley. Lighter dashed lines identify ^{10}Be sample areas shown in figure 5.

3 – Sampling sites for ^{10}Be surface-exposure dating.a) Moraine boulder sample WIL12-06 on the outer left-lateral moraine of the LLGM complex in the Willow Creek valley. Sample yielded an age of 16.7 ± 0.9 ka. b) Polished-bedrock sample sites WIL11-06 AND WIL11-07, approximately 3.3 km upvalley from the LLGM moraine complex in Willow Creek valley. Samples yielded ages of 14.3 ± 0.7 and 15.6 ± 0.8 respectively. Ages were calculated using the PPLm production model (see section 4.1.3).

4 – Observed and modeled snowfields in Willow Creek cirque. Red outlines indicate areas of late-summer snowfields in three or more of five available sets of air photos taken during the 1971-2000 AD interval. Black outlines indicate areas of late summer snowfields in at least two of the sets. The mean area of late-summer

snowfields in the five sets falls between those of the red and black outlines. Blue indicates areas of modeled annual net snow accumulation with no change from 1971-2000 AD precipitation, but with temperature increased 1.2 °C from 1971-2000 AD values. Shading indicates water-equivalent depth (WED) of the modeled snowfields, with darker blues indicating greater depth. Maximum modeled WED is 1.42 m. Light blue pixels along ridge crests indicate areas of modeled shallow snowfields (mean WED ~0.1 m) that do not avalanche to the snowfields below in the model, as calculated slopes along many ridge crests are less than the 30-degree threshold for avalanching. In fact, slopes along the ridges are steep enough that no snow remains on them through the summer season. Insert indicates position of orthophoto (yellow dashed line) in the Willow Creek valley. Orthophoto source is <https://gdg.sc.egov.usda.gov/>.

5 – Locations of ¹⁰Be ages. a) Willow Creek (bottom) and South Crestone Creek (top) valley moraine-boulder ages. Sample clusters on Willow Creek moraines discussed in text (section 5.1.1.1) are identified by color. Red indicates samples from distal portion of outermost right-lateral moraine; blue indicates samples from proximal portion of the same moraine; green indicates samples from the outermost left-lateral moraine; purple indicates samples from the inner right-lateral moraine. b) South Colony Creek valley moraine-boulder ages. c) Willow Creek valley polished-bedrock ages. All ages were calculated using the PPLm production model (see section 4.1.3).

6 – Plot of moraine-boulder ^{10}Be ages from the three study-area valleys. Ages from the Willow Creek moraines are color coded as in figure 5a. Ages were calculated using the PPLm production model (see section 4.1.3). Detailed sample information is given in table 3. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

7 – Plot of Willow Creek valley moraine-boulder and upvalley polished-bedrock ^{10}Be ages. Moraine-boulder ages are color coded as in figure 5a. Ages were calculated using the PPLm production model (see section 4.1.3). Detailed sample information is given in table 3. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

8 – Modeled local last glacial maximum (LLGM) glaciers in a) Willow Creek and b) South Crestone Creek drainages. Both models were run with “soft-ice” flow and sliding parameters and no change from modern precipitation. Willow Creek model was run with a year-round 5.0 °C temperature depression from modern, South Crestone Creek with a 5.1 °C temperature depression. Ice surface contour interval = 50 m. Shading indicates relative ice thickness—thicker ice is darker. Maximum modeled ice depth is 164 m in Willow Creek drainage, 119 m in South Crestone Creek drainage. Dashed line at terminus indicates position of dated outermost LLGM moraine in each valley.

9 – Combinations of temperature and precipitation change from modern (1971-2000 AD) conditions that could have sustained paleoglaciers in mass-balance equilibrium at their mapped extents. a) LLGM Willow Creek (red, solid line) and South Crestone Creek (blue, dashed line) paleoglaciers. b) Willow Creek paleoglacier at its LLGM extent (~18.2–red, solid line) and its extent at the time of nearly complete deglaciation of the valley (~14.0 ka–green, dashed line). A curve for the ~16.2ka extent of the Willow Creek glacier would be essentially the same as the ~18.2 ka curve. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

10 – Modeled Willow Creek glacier at ~14.0 ka, with a temperature depression of 2.5 °C from modern (1971-2000 AD) conditions, no change from modern precipitation, and “soft-ice” flow and sliding parameters. Shading indicates relative ice thickness—thicker ice is darker. Maximum modeled ice depth is 67 m. Dashed line downvalley indicates position of dated outermost LLGM moraine. The two most upvalley CRN sample localities that constrain deglaciation of that position in the valley to ~14.0 ka are indicated by red filled circles. Ice-surface contour interval = 50m. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Table 1

Table 1. Modern climate data used in glacier modeling

CLIMATE PARAMETER	DATA SOURCE	SITE	LATITUDE LONGITUDE	ALTITUDE (m)	DATES OF RECORD (AD)
Monthly mean temperature vs. altitude	PRISM 30-arc sec	Throughout basins	37 °57.8 to 38° 1.2' N 105° 34.7' to 105° 40.7' W	2519 - 4116	1971-2000
Monthly mean precipitation vs. altitude	PRISM 30-arc sec	Throughout basins	37 °57.8 to 38° 1.2' N 105° 34.7' to 105° 40.7' W	2519 - 4116	1971-2000
Monthly standard deviation of daily mean temperature, monthly mean percent of days with measureable precipitation ¹	SNOTEL (Snow Telemetry)	South Colony	37° 58' N 105° 32' W	3192	1992-2011
Monthly mean relative humidity, mean monthly wind speed	RAWS (Remote Automated	Black Mountain	37° 51.7' N 105° 17.1'	2738	2003-2011

Weather Station)

¹ Used as approximation of monthly mean cloud cover.

Table 2. Parameters used in glacier modeling.

MASS/ENERGY-BALANCE MODEL PARAMETERS

Snow albedo	0.8
Ice albedo	0.4
Snow/ice emissivity	0.99
Basin emissivity	0.94
Wind k^1	.001
Ground heat flux	0.1 Wm^{-2}

FLOW MODEL PARAMETERS

Deformation-law coefficient (A)	
“Soft-ice” simulation	$2.5 \times 10^{-7} \text{ yr}^{-1} \text{ kPa}^{-3}$
“Stiff-ice” simulation	$5 \times 10^{-8} \text{ yr}^{-1} \text{ kPa}^{-3}$
Deformation-law exponent (n)	3
Sliding-law coefficient (B)	
“Soft-ice” simulation	$3.75 \times 10^{-3} \text{ myr}^{-1} \text{ kPa}^{-2}$
“Stiff-ice” simulation	$7.5 \times 10^{-4} \text{ myr}^{-1} \text{ kPa}^{-2}$
Sliding-law exponent (m)	2

¹ Proportional change in wind speed per meter of altitude

Table 3. Cosmogenic ¹⁰Be data and surface-exposure ages of glacial features in the Sangre de Cristo Range

Moraine and sample ID	Latitude (°N)	Longitude (°W)	Elevation (m asl)	Sample thickness (cm)	Topographic shielding factor	[¹⁰ Be] (atoms · g SiO ₂ ⁻¹) ^a	1σ	LSD spall. scaling factor ^b	LSD muon scaling factor ^b	LSD exposure age (ka) ^c	1σ external (kyr) ^d	Promontory Point-Lm exposure age (ka) ^e	1σ external (kyr) ^d
Willow Creek Inner Right-Lateral Moraine													
WIL11-16	37.9888	-105.6486	3018	2.0	0.9872	5.20E+05	1.20E+04	8.3605	1.9266	15.9	0.7	15.7	0.7
WIL11-17	37.9891	-105.6478	3016	2.0	0.9897	5.56E+05	1.54E+04	8.3469	1.9256	16.9	0.8	16.7	0.8
Willow Creek Outer Left-Lateral Moraine													
WIL12-01	37.9852	-105.6481	3024	2.0	0.9961	5.34E+05	1.73E+04	8.3589	1.9283	16.1	0.8	15.9	0.8
WIL12-06	37.9849	-105.6503	2970	2.0	0.9952	5.42E+05	2.23E+04	8.0995	1.9058	16.9	0.9	16.7	0.9
Willow Creek Outer Right-Lateral Moraine, proximal portion													
WIL11-01	37.9887	-105.6502	3023	2.0	0.9900	5.33E+05	1.19E+04	8.3907	1.9289	16.2	0.7	16.0	0.7
WIL11-02	37.9887	-105.6502	3023	1.0	0.9903	5.32E+05	1.88E+04	8.3496	1.9278	16.0	0.8	15.8	0.8
WIL11-04	37.9890	-105.6495	3036	2.0	0.9918	5.30E+05	1.73E+04	8.4588	1.9344	15.9	0.8	15.7	0.8
WIL11-14	37.9890	-105.6495	3036	2.0	0.9927	5.66E+05	1.86E+04	8.3916	1.9291	17.1	0.8	16.9	0.8
										mean = 16.3 ± 0.5 (n=4 , χ ² R ^f = 0.47)	mean = 16.1 ± 0.5 (n=4 , χ ² R= 0.47)		
Willow Creek Outer Right-Lateral Moraine, distal portion													
WIL11-03	37.9879	-105.6519	2996	1.0	0.9926	5.98E+05	2.15E+04	8.2616	1.9177	18.2	0.9	17.9	0.9
WIL14-01	37.9852	-105.6534	2962	2.5	0.9951	5.92E+05	1.99E+04	8.1467	1.9047	18.5	0.9	18.3	0.9
WIL14-02	37.9854	-105.6534	2963	2.5	0.9937	6.00E+05	1.32E+04	8.1642	1.9054	18.7	0.8	18.5	0.8
										mean = 18.5 ± 0.3 (n=3 , χ ² R= 0.087)	mean = 18.2 ± 0.3 (n=3 , χ ² R= 0.13)		
South Colony Creek Inner Left-Lateral Moraine													
SCO-01	37.9882	-105.4926	2941	7.0	0.9875	5.69E+05	1.92E+04	7.9962	1.8952	18.8	0.9	18.6	0.9
SCO-02	37.9887	-105.4918	2932	3.5	0.9892	6.08E+05	1.88E+04	7.9764	1.892	19.5	0.9	19.3	0.9
SCO11-01	37.9875	-105.4984	3024	3.0	0.9948	6.04E+05	1.44E+04	8.4692	1.9313	18.3	0.8	18.1	0.8
SCO11-02	37.9876	-105.4975	3024	3.0	0.9945	6.29E+05	1.97E+04	8.5188	1.9325	18.9	0.9	18.8	0.9
										mean = 18.9 ± 0.5 (n=4 , χ ² R= 0.34)	mean = 18.7 ± 0.5 (n=4 , χ ² R=0.34)		
South Colony Creek Right-Lateral Moraine													
SCO11-6B	37.9821	-105.4960	3043	2.0	0.9964	6.15E+05	2.31E+04	8.5213	1.9384	18.2	0.9	18.0	0.9
South Colony Creek Outer Left-Lateral Moraine													
SCO11-07	37.9899	-105.4953	3000	2.0	0.9946	5.88E+05	2.29E+04	8.327	1.9207	17.9	0.9	17.7	0.9
SCO11-08	37.9874	-105.4996	3000	2.0	0.9952	7.20E+05	1.93E+04	8.5827	1.9273	21.3	1.0	21.5	1.0
South Crestone Creek Moraine													
SCR-01	38.0003	-105.6497	3161	2.5	0.944	6.00E+05	1.78E+04	9.1688	1.9897	17.5	0.8	17.4	0.8
SCR-02	38.0004	-105.6495	3167	5.0	0.944	6.99E+05	2.64E+04	9.3473	1.9957	20.2	1.1	20.5	1.1
SCR-03	38.0007	-105.6485	3181	4.5	0.9249	7.01E+05	2.71E+04	9.4483	2.0025	20.4	1.1	20.7	1.1
										mean = 19.6 ± 1.6 (n=3, χ ² R=3.3)	mean = 19.6 ± 1.9 (n=3, χ ² R= 4.5)		
Willow Creek Glacially Scoured Bedrock													
WIL11-09	37.9853	-105.5959	3754	3.0	0.9525	6.76E+05	2.70E+04	13.0024	2.2722	13.9	0.8	14.1	0.8
WIL11-10	37.9854	-105.5948	3761	4.0	0.9561	6.70E+05	2.66E+04	13.0576	2.2759	13.8	0.8	14.0	0.8
WIL11-05A	37.9930	-105.6042	3559	1.0	0.9625	7.62E+05	2.43E+04	11.7019	2.1763	16.9	0.8	17.1	0.8
WIL11-05B	37.9930	-105.6042	3559	2.0	0.961	7.20E+05	2.13E+04	11.6789	2.1758	16.2	0.8	16.3	0.8
WIL11-06	37.9926	-105.6151	3517	1.0	0.9416	6.05E+05	1.79E+04	11.3012	2.1534	14.2	0.7	14.3	0.7
WIL11-07	37.9923	-105.6148	3519	1.0	0.943	6.66E+05	2.15E+04	11.3615	2.1554	15.5	0.8	15.6	0.8
WIL11-08	37.9924	-105.6222	3463	3.5	0.9565	6.67E+05	1.81E+04	11.0218	2.129	16.2	0.7	16.2	0.7
WIL11-12	37.9918	-105.6210	3461	2.0	0.9347	7.00E+05	4.05E+04	11.0313	2.1285	17.1	1.2	17.2	1.2
WIL12-03	37.9909	-105.6293	3162	2.0	0.9323	1.16E+06	4.81E+04	9.8856	2.008	31.7	1.8	32.7	1.8
WIL12-04	37.9907	-105.6294	3146	2.0	0.9263	8.98E+05	2.74E+04	9.5381	1.9943	25.6	1.2	26.1	1.2
WIL12-04B	37.9907	-105.6294	3146	2.0	0.9263	1.15E+06	6.06E+04	9.7911	2.0009	31.9	2.1	32.9	2.1

Note: all samples have a density of 2.7 g/cm³ and did not display evidence of surface erosion. All ¹⁰Be measurements were completed by accelerator mass spectrometer (AMS) at PRIME Lab (Purdue University). All samples were measured against AMS standard coded as "KNSTD07" in the CRONUS Earth online exposure age calculator and were measured against Beryllium AMS standard Be-01-5-3 (Nishiizumi et al., 2007). Mean exposure ages of moraines are computed from all samples with 1σ uncertainty. Ages in italics are considered outliers.

^aBlank-corrected concentration.

^bComputed from the scaling model of Lifton et al. (2014). "LSD" = Lifton-Sato-Dunai.

^cComputed using the LSD scaling factors given in two preceding columns.

^dUncertainty of AMS measurement and production rate.

^eComputed from the calibrated production rate at Promontory Point (Lifton et al., 2015) with "Lm" scaling by the CRONUS online exposure-age calculator, version 2.3 (Balco et al., 2008).

^fFor all moraines, this value is computed from the surface exposure ages $\pm$ external uncertainty.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10