

Pinedale glacial history of the upper Arkansas River valley: New moraine chronologies, modeling results, and geologic mapping

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ABSTRACT

This field-trip guide outlines the glacial history of the upper Arkansas River valley, Colorado, and builds on a previous GSA field trip to the area in 2010. The following will be presented: (1) new cosmogenic ^{10}Be exposure ages of moraine boulders from the Pinedale and Bull Lake glaciations (Marine Isotope Stages 2 and 6, respectively) located adjacent to the Twin Lakes Reservoir, (2) numerical modeling of glaciers during the Pinedale glaciation in major tributaries draining into the upper Arkansas River, (3) discharge estimates for glacial-lake outburst floods in the upper Arkansas River valley, and (4) ^{10}Be ages on flood boulders deposited downvalley from the moraine sequences. This research was stimulated by a new geologic map of the Granite 7.5' quadrangle, in which the mapping of surficial deposits was revised

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based in part on the interpretation of newly acquired LiDAR data and field investigations. The new ^{10}Be ages of the Pinedale terminal moraine at Twin Lakes average 21.8 ± 0.7 ka ($n = 14$), which adds to nearby Pinedale terminal moraine ages of 23.6 ± 1.4 ka ($n = 5$), 20.5 ± 0.2 ka ($n = 3$), and 16.6 ± 1.0 ka ($n = 7$), and downvalley outburst flood terraces that date to 20.9 ± 0.9 ka ($n = 4$) and 19.0 ± 0.6 ka ($n = 4$). This growing chronology leads to improved understanding of the controls and timing of glaciation in the western United States, the modeling of glacial-lake outburst flooding, and the reconstruction of paleotemperature through glacier modeling.

INTRODUCTION

This field trip to the upper Arkansas River valley of central Colorado (Fig. 1) builds on a previous GSA field trip (McCalpin et al., 2010), in an area with a long history of bedrock mapping by renowned geologists of the U.S. Geological Survey beginning in the 1870s (e.g., Hayden, 1874; Emmons, 1886). This trip focuses primarily on Quaternary deposits, which were first described by Capps (1909). Much of the research presented here is new since the 2010 field trip (McCalpin et al., 2010), and thus we will visit both new and previous locations of interest.

The focus of this field trip is a new chronology of a moraine complex in the upper Arkansas River valley developed from cosmogenic ^{10}Be exposure dating. The high sensitivity of alpine glaciers to climate variability (Oerlemans, 2005) makes moraines among the best proxies for reconstructing spatial and temporal patterns of climate change. Since the advent of cosmogenic nuclide exposure dating, distinct patterns in the timing of mountain glacier advance and retreat across the western United States have emerged (e.g., Gosse et al., 1995; Licciardi et al., 2001, 2004; Brugger, 2007; Munroe et al., 2006; Licciardi and Pierce, 2008; Refsnider et al., 2008; Laabs et al., 2007, 2009, 2011, 2013; Briner, 2009; Ward et al., 2009; Young et al., 2011). However, directly dated glacial features are still relatively sparse, and improving the knowledge of climate during the Last Glacial Maximum (LGM; ~ 25 – 19 ka) requires a more comprehensive record of glaciation spanning the Rocky Mountains. Generating chronologies of Pinedale glaciation (the Pinedale glaciation is the name for Marine Isotope Stage [MIS] 2 deposits in the Rocky Mountains; ~ 30 – 10 ka) is necessary for understanding the timing of both glacial culminations and subsequent deglaciation. This information allows for identification of mechanisms that drive glacier change, such as regional versus global forcing.

Highlights of this field trip include: (1) new cosmogenic ^{10}Be exposure ages of moraine boulders from the Pinedale and Bull Lake glaciations (the Bull Lake glaciation is the name for MIS 6 deposits in the Rocky Mountains; ~ 190 – 130 ka) located directly southeast of the Twin Lakes Reservoir (Fig. 2), (2) numerical modeling of glacier ice during the Pinedale glaciation in major tributaries that terminated in the upper Arkansas River valley, (3) discharge estimates for glacial-lake outburst floods in the upper Arkansas River valley downstream of Clear and Pine Creek valleys, and (4) ^{10}Be exposure ages on flood boulders deposited

downvalley from the moraine sequences. Recent research in this area was stimulated by a new geologic map of the Granite 7.5' quadrangle covering parts of Lake and Chaffee Counties, Colorado, in which the mapping of surficial deposits was revised based in part on the interpretation of newly acquired LiDAR data and field investigations (Shroba et al., 2014).

The general itinerary of the trip is a 414 km (257-mile [mi]) loop that starts and ends at the Colorado Convention Center, 700 14th Street, Denver, Colorado 80202. This one-day field trip begins at 8:00 a.m. and returns to Denver at $\sim 11:00$ p.m. (Fig. 1).

Upper Arkansas River Valley Geomorphology and Previous Work

The geologic map of the Granite 7.5' quadrangle (Fig. 3) contains one of the most complete sequences of till and outwash in central Colorado, particularly in areas along and near the Twin Lakes and Clear Creek Reservoirs, and the Arkansas River. The upper Arkansas River valley encompasses the headwaters for the Arkansas River, which drains the Mosquito Range to the east and the Sawatch Range to the west. Surficial deposits in this region and their associated landforms have been the subject of reconnaissance studies and detailed investigations since the early 1870s (e.g., Hayden, 1874; Capps and Leffingwell, 1904; Davis, 1905; Westgate, 1905; Capps, 1909; Ray, 1940; Richmond and Tweto, 1965; Tweto and Case, 1972; Nelson and Shroba, 1998; Briner, 2009; Young et al., 2011; Ruleman et al., 2013; Shroba et al., 2014).

Bedrock in this region is composed primarily of Precambrian plutonic and metamorphic basement rocks (mostly gneisses and granites), as well as Late Cretaceous and Paleogene plutonic rocks, and Neogene basin-fill sediments (Tweto et al., 1976, 1978; Scott et al., 1978). During the pre-Bull Lake, Bull Lake, and Pinedale glaciations, Sawatch Range glaciers extended eastward toward the upper Arkansas River valley. Glaciers that flowed down Lake, Clear, and Pine Creek valleys extended beyond the range front and onto the floor of the upper Arkansas River valley, depositing abundant till and outwash gravel that formed large lateral and end moraines and extensive outwash terraces (Nelson and Shroba, 1998; Shroba et al., 2014). At their Pinedale maximum extent, it is thought that glaciers emanating from Clear and Pine Creek valleys dammed the Arkansas River (Scott, 1984; Nelson and Shroba, 1998; Lee, 2010). Clear and Pine Creek valleys each host a pair of sharp, single-crested lateral moraines that

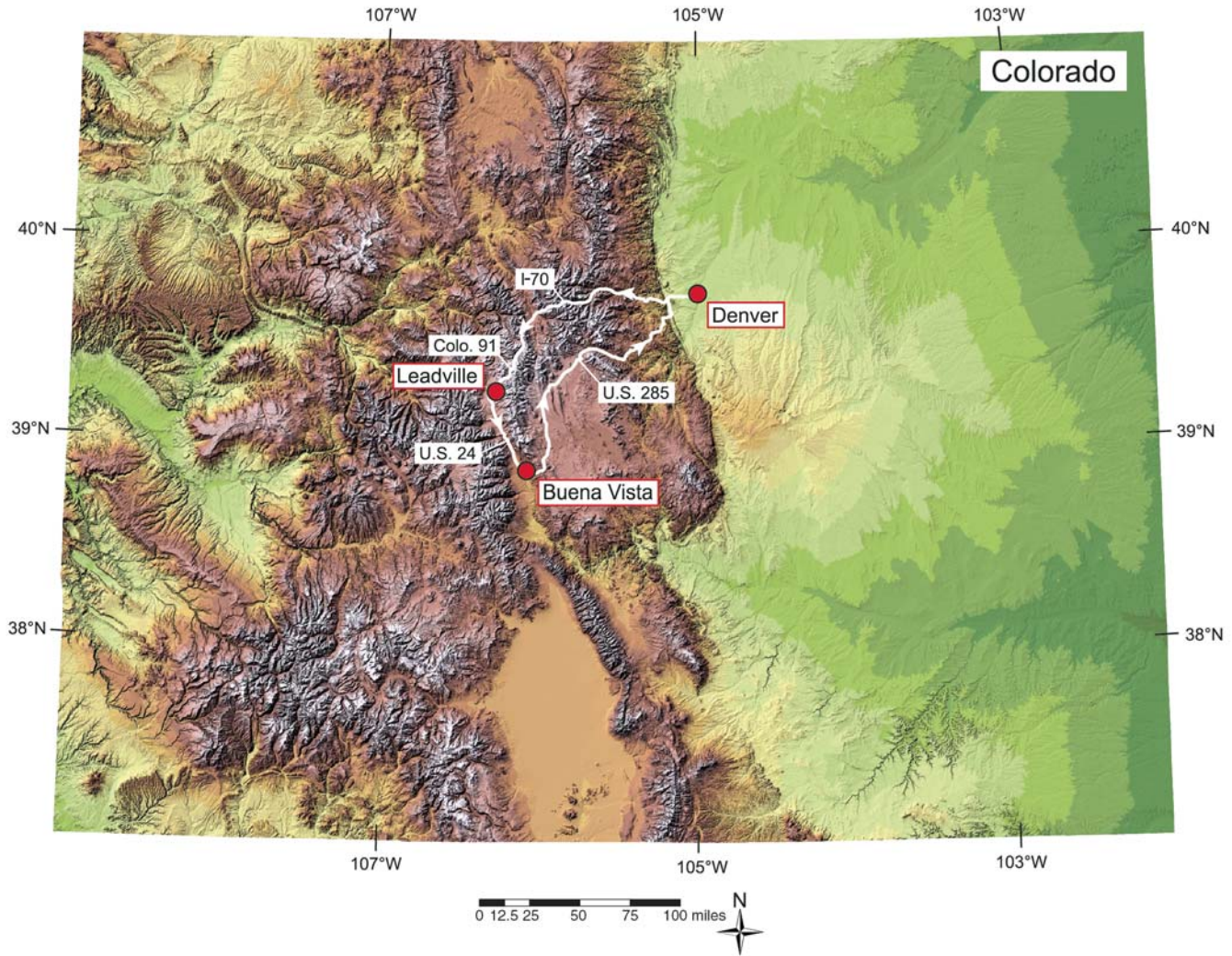


Figure 1. Route of field trip (white line with arrows) superimposed on an elevation relief map of Colorado. This one-day field trip heads west on Interstate Highway 70 (I-70) to Copper Mountain, then southwest on Colorado Highway 91 (Colo. 91) to Leadville, Colorado (100 mi). All of the field-trip stops are located off U.S. Highway 24 (U.S. 24) between Leadville and Buena Vista, Colorado (55 km [34 mi]). The route back to Denver follows U.S. Highway 285 (U.S. 285; 198 km [123 mi]).

are Pinedale in age, and are easily distinguishable from more extensive Bull Lake moraines, which are characterized by a more subdued surface morphology and a higher degree of surface weathering of exposed boulders (Nelson and Shroba, 1998; Briner, 2009; Young et al., 2011). In contrast to Clear Creek and Pine Creek valleys, Lake Creek valley contains numerous Pinedale lateral and end moraines (Nelson and Shroba, 1984; Shroba et al., 2014) located directly inboard of more extensive Bull Lake moraines. Glacier retreat from the terminal moraines in Clear Creek and Pine Creek valleys likely led to a series of glacial-lake outburst floods that flowed south and deposited boulder-rich terrace gravels downstream of Pine Creek valley. These terrace gravels rest ~15 and ~6 m above the present course of the Arkansas River (Scott, 1984; Lee, 2010).

Recent work presented as part of the 2010 GSA-sponsored field trip to the upper Arkansas River valley (McCalpin et al., 2010) included ^{10}Be exposure ages of boulders deposited during the Pinedale and Bull Lake glaciations in the Clear Creek and Pine Creek valleys (Fig. 4; Schildgen, 2000; Briner, 2009; Young et al., 2011). These ^{10}Be exposure ages are recalculated here using a production rate of 4.00 ± 0.15 atoms/g/yr (sea-level high latitude) and time-dependent ('Lm') scaling, which is based on a recently published calibration of in situ ^{10}Be production at Promontory Point in northern Utah (Lifton et al., 2015).

Bull Lake moraines are preserved in Lake Creek, Clear Creek, and Pine Creek valleys. Prior to the new work presented here, earlier attempts to date the Bull Lake glaciation had yielded ^{10}Be ages ($n = 10$) from the left lateral Bull Lake moraine (Dry

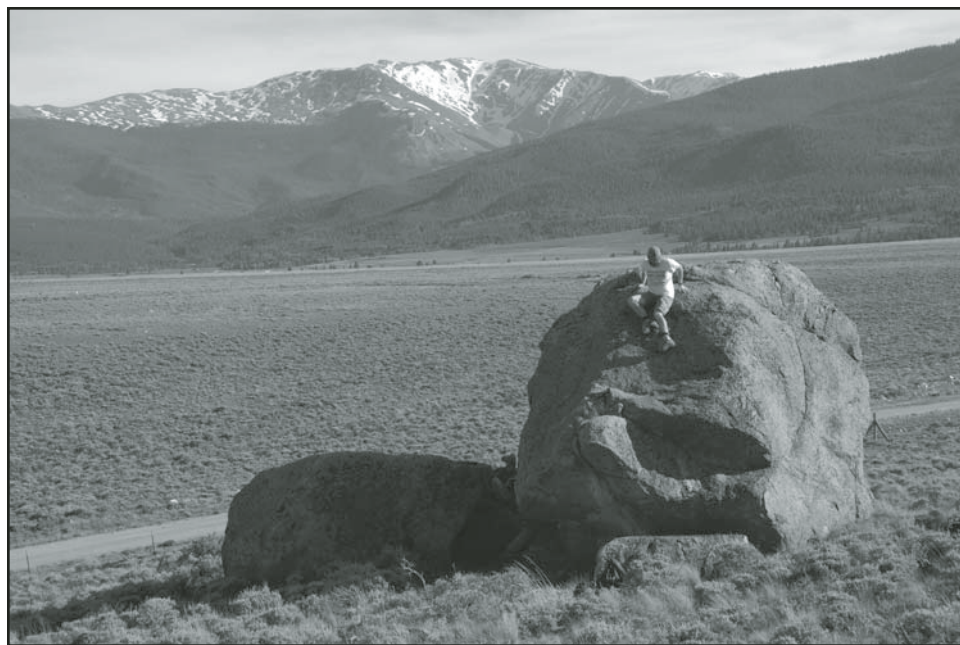


Figure 2. Sampling a Bull Lake moraine boulder in the Twin Lakes moraine sequence.

Creek Gulch) in Pine Creek valley that range from ca. 74.7 ka to 4.0 ka, all likely anomalously young (Briner, 2009). While there is some evidence for a glacier advance between MIS 6 and MIS 2 glaciation in the Rocky Mountains (e.g., Colman and Pierce, 1986), most records across this region indicate the presence of only MIS 6 and MIS 2 age deposits (e.g., Pierce, 2004; Licciardi and Pierce, 2008). The lack of glacier deposits between MIS 6 and MIS 2 is also thought to be true for the Arkansas River valley (Nelson and Shroba, 1998). Thus, it is likely that the ^{10}Be ages from Dry Creek Gulch reflect postdepositional processes (e.g., exhumation) rather than the age of moraine deposition (Briner, 2009). Although till of Bull Lake age is preserved in Clear Creek valley (Shroba et al., 2014), there is no numerical age control on these deposits.

Existing ^{10}Be ages indicate asynchronous Pinedale maxima at 23.6 ± 1.4 ka, 20.5 ± 0.2 ka, 19.0 ± 0.2 ka, and 16.6 ± 1.0 ka for valley glaciers in the upper Arkansas River valley (Fig. 4; Young et al., 2011). ^{10}Be ages of glacially polished bedrock surfaces located upvalley of Pinedale terminal moraines in valleys of the Lake, Clear, and Pine Creeks were also obtained, and indicate that retreat of all three paleoglaciers initiated synchronously in the upper Arkansas River valley between ca. 17–16 ka. ^{10}Be ages from Clear Creek valley suggest that ice initially retreated from its Pinedale maximum extent at 20.5 ± 0.2 ka, which is correlative with the age of the upper flood gravel (20.9 ± 0.9 ka; Fig. 4), implying that the Clear Creek glacier acted as the main ice dam impounding the Arkansas River (Young et al., 2011). It is somewhat enigmatic that the age of the lower terrace (19.0 ± 0.2 ka) is not expressed in the moraine records, given that deposition of the lower flood gravel requires glaciers to have been at or near their Pinedale maxima in order to dam the Arkansas River. Based on

the close correspondence between Clear Creek moraine abandonment and the upper flood terrace ^{10}Be ages, Young et al. (2011) postulate that the Clear Creek valley glacier was near its Pinedale maximum ~ 19.0 ka (Fig. 4).

Eighteen new ^{10}Be exposure ages presented in this report (and computed using the same production rate and scaling described above) are from large crystalline boulders on moraine crests enclosing the Twin Lakes Reservoir in Lake Creek valley. These moraines are composed of glacial debris recently mapped as Pinedale and Bull Lake by Shroba et al. (2014). Till of Pinedale age was mapped as two separate units: (1) Pinedale “younger” till (Qtpy) forms large, sharp-crested moraines that grade to slightly lower outwash gravel that is directly inboard of or upvalley of low-relief, subdued, and hummocky Pinedale “older” moraines; and (2) Pinedale “older” till (Qtpo) forms low moraines that grade to higher outwash gravel that is slightly outboard of or downvalley of large, sharp-crested younger moraines mapped as Pinedale “younger” till (Qtpy; Shroba et al., 2014). Till deposited during the Bull Lake glaciation in the Granite quadrangle is not subdivided into older and younger units. Thus, the primary objective of our recent work was to apply ^{10}Be dating to develop a detailed chronology for the three mapped moraine groups and their associated tills on the south side of the Twin Lakes Reservoir (Bull Lake, Pinedale “older,” and Pinedale “younger”). Our results provide direct age control for multiple glacier maxima including two map units (Pinedale “older” and Bull Lake) previously mapped in less detail by Shroba (1977) and Nelson and Shroba (1998) in this area, and address whether the ages of Bull Lake and Pinedale moraines mapped by Shroba et al. (2014) near Lake Creek are in accord with ages reported in adjacent valleys in the Arkansas River basin and across the western United States.

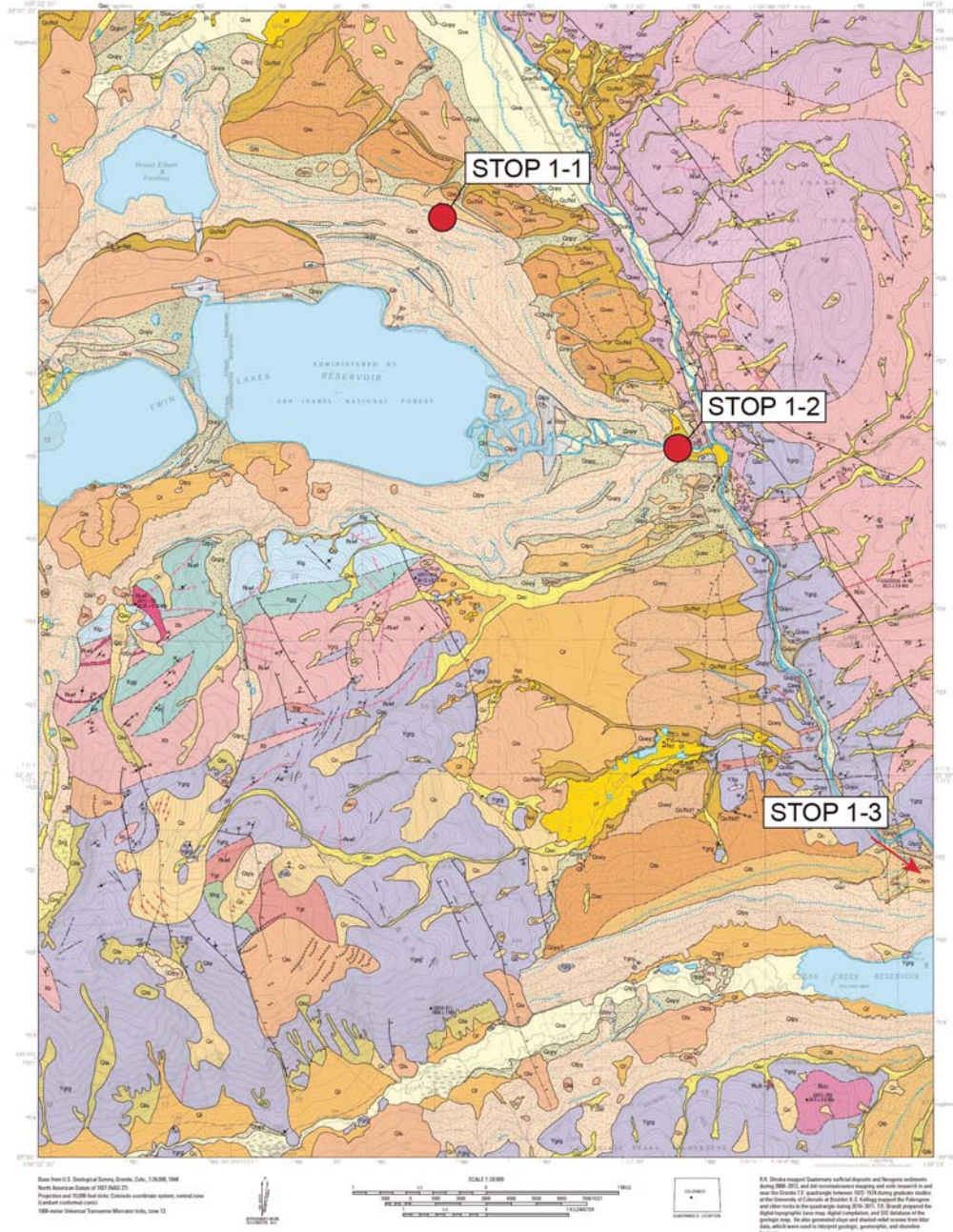


Figure 3. Geologic map of the Granite 7.5' quadrangle, Lake and Chaffee Counties, Colorado, by Shroba et al. (2014). Stop locations 1-1 and 1-2 are shown (red circles). Stops 1-3 through 1-6 are located outside the map boundaries.

Cosmogenic Nuclide Exposure Dating

Cosmogenic nuclide exposure dating (Gosse and Phillips, 2001; herein exposure dating) has emerged as the premier technique for directly dating moraines. The method has benefited from recent advances in our knowledge of nuclide production rates and the ability to obtain high analytical precision for low-

level isotope concentrations in geologically young surfaces. This method relies on the in situ production of cosmogenic nuclides (^3He , ^{10}Be , ^{14}C , ^{21}Ne , ^{26}Al , ^{36}Cl) in exposed rock surfaces. Cosmogenic isotopes are produced by collisions between cosmic-ray particles and molecules in Earth's atmosphere and lithosphere, with target nuclei within certain minerals in exposed surfaces on Earth (Gosse and Phillips, 2001; Balco,

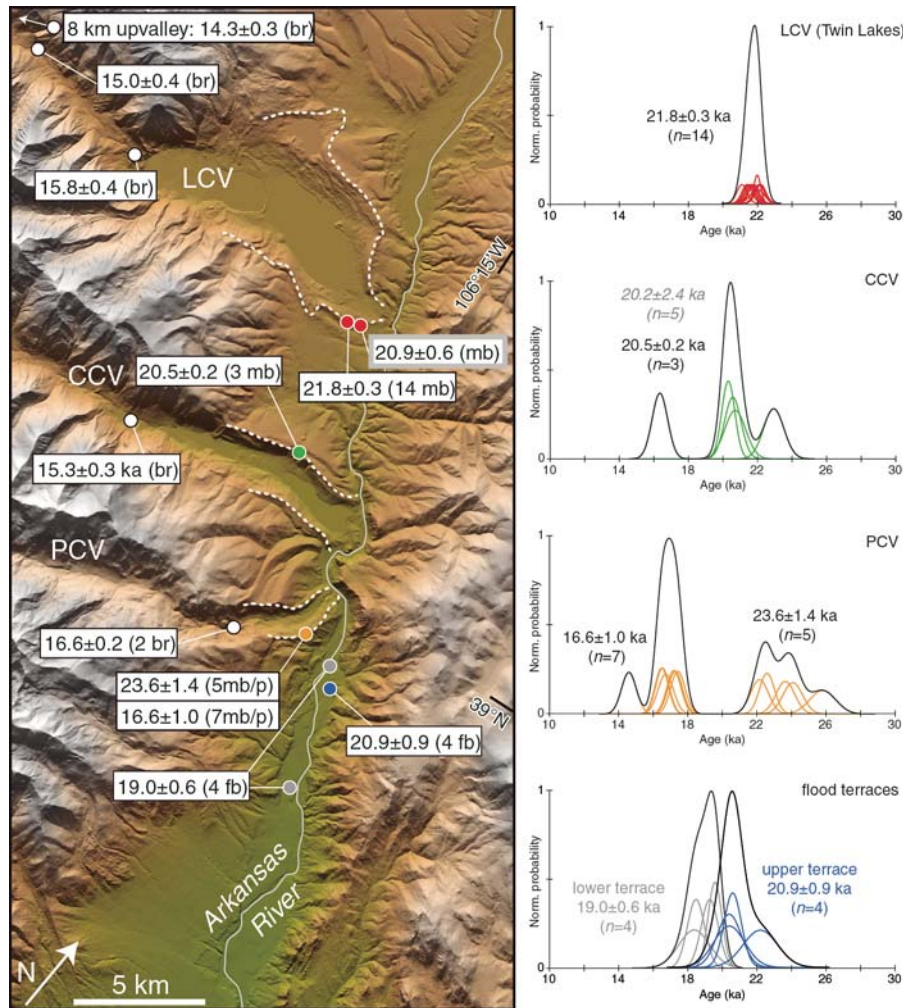


Figure 4. Left panel: Upper Arkansas valley study area showing Pinedale ^{10}Be ages (ka) with 1σ uncertainty. Dashed lines reflect the downvalley limit of till of Pinedale glaciation. ^{10}Be ages from Lake Creek valley (LCV) are new, except for one age (box with gray outline; Schildgen, 2000), while ^{10}Be ages from Pine Creek valley (PCV) and Clear Creek valley (CCV) were previously published (Briner, 2009; Young et al., 2011). br—bedrock; mb—moraine boulder; mb/p—moraine boulders and pebbles; fb—terrace boulder (gray = lower terrace, blue = upper terrace). Right panel: Probability density functions ('camel plots'; Balco, 2011) for ^{10}Be ages from large boulders on Pinedale terminal moraines and flood deposits. All ^{10}Be ages including those from previous work are recalculated using the production rate calibrated at Promontory Point (Lifton et al., 2015) and 'Lm' scaling.

2011). The concentration of in situ cosmogenic nuclides within the uppermost surface of Earth's crust (~ 3 m) is directly related to the duration that the surface has been exposed to cosmic ray bombardment. Thus, the longer a surface has been exposed to cosmic rays, the greater the abundance of cosmogenic nuclides that surface will contain. By measuring the concentration of a particular in situ cosmogenic nuclide in a rock surface, combined with knowledge of that nuclide-specific production rate, it is possible to determine the time elapsed since that surface was initially exposed (i.e., exposure age).

Although several in situ cosmogenic nuclides are produced within Earth's crust, ^{10}Be is the most commonly measured nuclide because (1) quartz-bearing lithologies are widespread and cosmogenic ^{10}Be is primarily produced in quartz, (2) chemical procedures for extracting ^{10}Be from quartz mineral separates are well established (Kohl and Nishiizumi, 1992; Corbett et al., 2016), (3) it is possible to achieve high-level precision and accuracy for accelerator mass spectrometry analysis of ^{10}Be samples, and (4) the production rate of ^{10}Be in quartz is well constrained. As a result of these recent advances, ^{10}Be exposure dating has

become one of the best resolved and widely used applications of the method, and is well suited for sites in central Colorado due to the abundance of glacially polished, crystalline, quartz-rich moraine boulders and glacially sculpted bedrock. In particular, ^{10}Be measurements have been extensively used to establish late Pleistocene moraine chronologies in Colorado and across the western United States (e.g., Licciardi et al., 2001, 2004; Pierce, 2004; Licciardi and Pierce, 2008; Munroe et al., 2006; Brugger, 2007; Laabs et al., 2007, 2009, 2011, 2013; Briner, 2009; Ward et al., 2009; Young et al., 2011).

Despite the wide use of ^{10}Be dating, age uncertainties can be introduced by incomplete knowledge of isotope production rates and scaling models, as well as impacts associated with a host of geological processes. The reliability of ^{10}Be exposure dating in the western United States is strengthened by the recent development of a new locally calibrated production rate, thereby improving the accuracy of the dating method (Lifton et al., 2015; Borchers et al., 2016). We calculate all ^{10}Be exposure ages for the upper Arkansas River valley using the Promontory Point production rate of $4.00 \pm$

0.15 atoms/g/yr (sea-level high latitude) and ‘Lm’ scaling (Lifton et al., 2015).

Geological uncertainties may arise from unrecognized prior exposure (resulting in isotopic inheritance), rock surface erosion, post-depositional rotation or exhumation of boulders, and burial by snow or loess. Of these factors, isotopic inheritance would result in apparent ages that are too old, whereas the other three phenomena would result in apparent ages that are too young. Sampling protocols used to date moraines and bedrock in this and previous work conducted in the upper Arkansas River valley were designed to minimize potential geological uncertainties. Moraine boulders and erratics selected for sampling are typically large, stable, and well preserved, and many of them exhibit original glacial smoothing or striæ on their upper surfaces. Similarly, sampled bedrock surfaces were carefully targeted to yield exposure ages that accurately constrain the timing of deglaciation.

Within the upper Arkansas River valley, we build on previous work conducted by Schildgen (2000), Briner (2009), and Young et al. (2011; Fig. 4) by dating three distinct moraine groups including the Pinedale “younger,” Pinedale “older,” and Bull Lake positions enclosing Twin Lakes. We obtained seven ^{10}Be exposure ages from the Pinedale “younger” moraine, seven ^{10}Be exposure ages from the Pinedale “older” moraine, and four ^{10}Be exposure ages from the Bull Lake moraine. At the time of this writing, one ^{10}Be exposure age from the recessional moraine that bisects the Twin Lakes Reservoir and four ^{10}Be ages from cirque lips in upper Lake Creek valley are pending.

ROAD LOG

Driving directions from the Colorado Convention Center, Denver, Colorado to Stop 1-1. Depart the Colorado Convention Center (700 14th Street, Denver, Colorado 80202) and head southeast on 14th Street toward California Street. Turn right onto Glenarm Place and after 0.3 km (0.2 mi) turn right onto Colfax Ave. Continue ~1.3 km (~0.8 mi) and then use the right lane to merge southward onto Interstate Highway 25 (I-25) via the ramp to Colorado Springs. After merging southbound with I-25, take Exit 209B for U.S. Highway 6 (U.S. 6; 6th Avenue) toward Lakewood. Continue west on U.S. 6 for 15 km (9 mi) and then take the exit onto I-70 (west) toward Grand Junction. Continue west on I-70 for 105 km (65.5 mi). Take Exit 195 for Colorado Highway 91 (Colo. 91) toward Copper Mountain and Leadville and drive for 35.5 km (22.1 mi). Continue south through Leadville onto U.S. Highway 24 (U.S. 24; Poplar Street) for another 26.2 km (16.3 mi) and into the field trip area (Fig. 5). Turn right onto Colorado Highway 82 (Colo. 82), which cuts through the Twin Lakes end moraine. Continue on Colo. 82 for ~2.9 km (~1.8 mi) and then turn right onto County Road 10 (CR 10). Follow CR 10 up the hill for ~1.6 km (~1 mi) until you reach an intersection and the paved road ends. Turn left onto Pan Ark Drive (gravel road) and continue for ~1.1 km (~0.7 mi) (road turns into Parry Peak Drive). Park at the top of the hill where the road makes a sharp left.

STOP 1-1. Overview of Lake Creek Valley and the Twin Lakes Reservoir

This stop provides an overview of the Twin Lakes Reservoir (Fig. 6). The viewpoint is atop the Pinedale “younger” end moraine, with a view of the dated Pinedale and Bull Lake moraines to the south, the Arkansas River to the southeast, and Lake Creek valley to the west. This stop will serve as an introduction to the Twin Lakes field site, the purpose of our recent work here, a review of the prior research conducted in the region (i.e., Shroba, 1977; Nelson and Shroba, 1998; Schildgen, 2000; Briner, 2009; Young et al., 2011), and the new geological map of the area (Shroba et al., 2014).

Lake Creek valley drains the north-south-trending Sawatch Range and is subparallel to adjacent valleys to the south (Clear Creek and Pine Creek valleys) and north (Halfmoon Creek and Rock Creek valleys), each of which contained sizeable glaciers during Quaternary glaciations (Ruleman et al., 2013). The Lake Creek glacier extended just shy of the present course of the Arkansas River during the Pinedale glaciation (Fig. 6; Shroba, 1977; Nelson and Shroba, 1998; Shroba et al., 2014). Bull Lake deposits were initially identified based on age-related soil properties

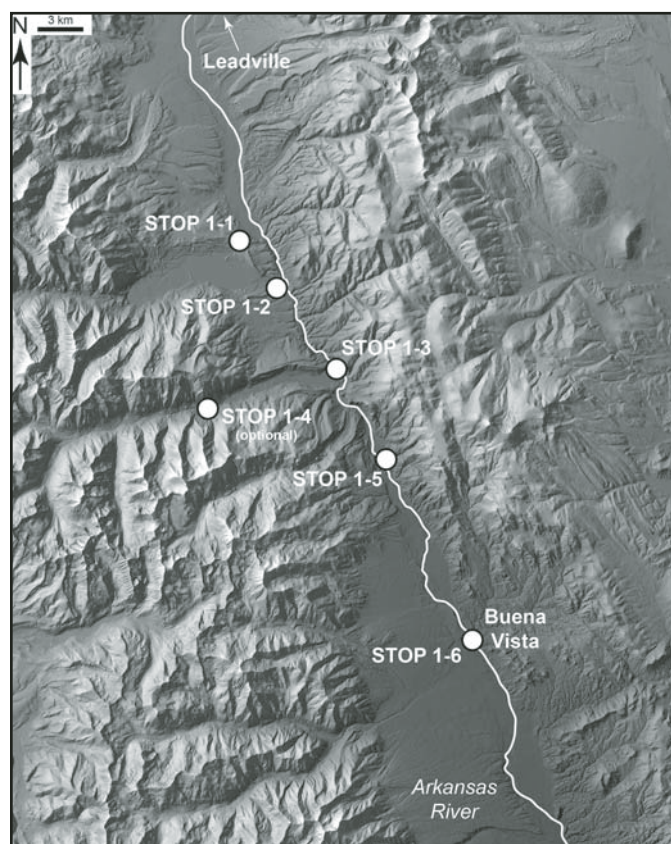


Figure 5. Stop locations in the upper Arkansas River valley, Colorado. All stops are located between the towns of Leadville and Buena Vista, Colorado.

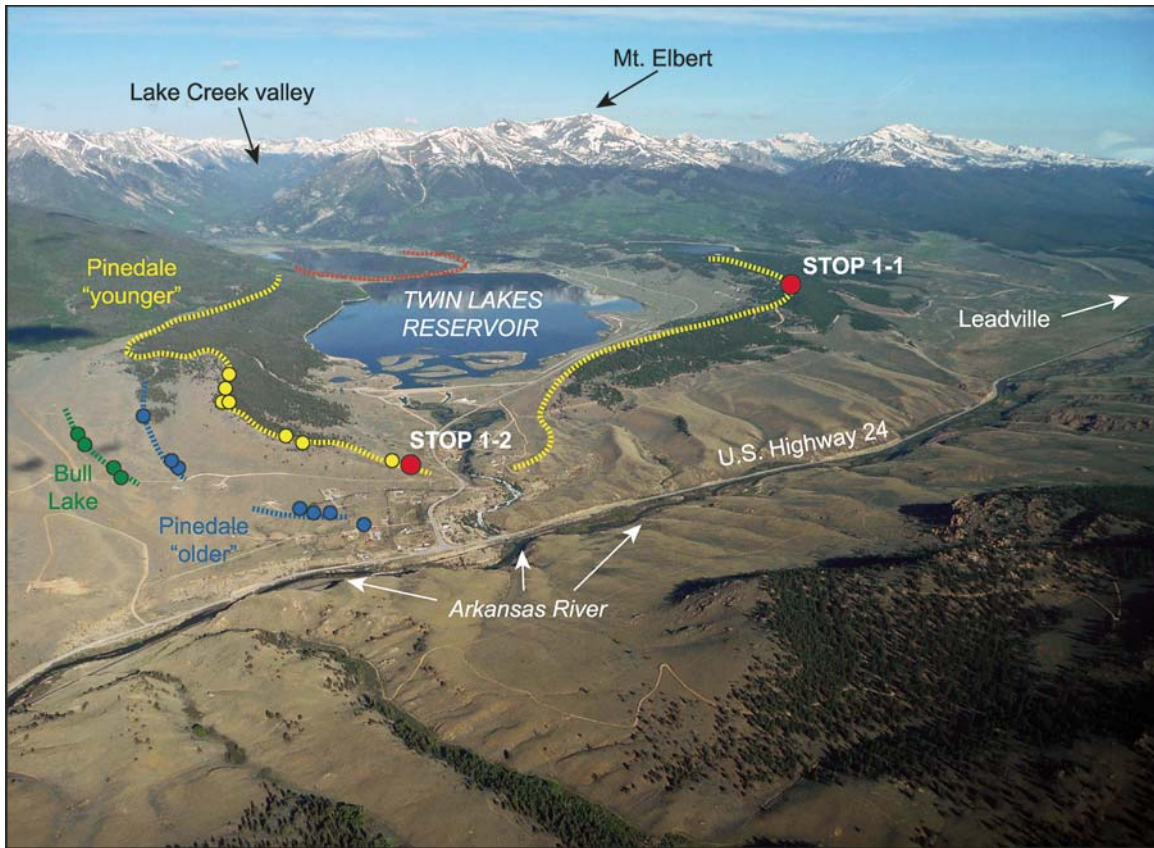


Figure 6. Twin Lakes Reservoir and adjacent area looking to the west. Yellow, blue, and green dashed lines demarcate crests of the Pinedale “younger” (Qtpy), Pinedale “older” (Qtpo), and Bull Lake (Qtb) moraines, respectively. Also shown in the same colors are the approximate locations of dated boulders on each moraine crest, the modern course of the Arkansas River, and stop locations 1-1 and 1-2.

(Shroba, 1977; Nelson and Shroba, 1998; Shroba et al., 2014), although deposits of Bull Lake age are relatively scarce as they were typically overridden and buried by the subsequent Pinedale advance(s). Bull Lake remnants are present on the north (left-lateral) and south (right-lateral) side of Lake Creek in the upper Arkansas valley, and are characterized by subdued surface morphology (Lee, 2010; Shroba et al., 2014). Lee (2010) noted a general lack of boulders on the Bull Lake left-lateral moraine; however, a few extremely large boulders exist on the right-lateral, and were targeted for ^{10}Be dating.

Lake Creek valley and adjacent areas preserve one of the most extensive suites of moraines in the upper Arkansas River valley, including evidence for multiple Pinedale advances (Figs. 7 and 8). Although Lake Creek glacier may have advanced earlier during the Pinedale glaciation, the Pinedale “older” moraine is the most extensive end moraine associated with the Pinedale glaciation (blue lines, Figs. 6 and 7). Directly inboard of the Pinedale “older” moraine are prominent, sharp-crested lateral moraines that are also Pinedale in age (yellow lines, Figs. 6 and 7, Pinedale “younger”; Lee, 2010; Shroba et al., 2014). A recessional moraine divides the Twin Lakes Reservoir into two

separate lakes and is located ~5 km upvalley from the prominent end moraines complex (orange lines; Figs. 6 and 7).

Prior to this work, the only numerical age control for moraines in Lake Creek valley came from two ^{10}Be ages of 21.0 ± 0.6 ka and 20.9 ± 0.6 ka for boulders on the Bull Lake and Pinedale right-lateral moraines, respectively (Schildgen, 2000). The ^{10}Be age of the Bull Lake boulder is considered erroneously young, possibly due to post-depositional processes that resulted in an age indistinguishable from those on the adjacent younger Pinedale moraines (Schildgen, 2000). Young et al. (2011) constrained the timing of ice retreat in Lake Creek valley with ^{10}Be ages of 15.8 ± 0.4 ka, 15.0 ± 0.4 ka, and 14.3 ± 0.3 ka from glacially sculpted bedrock at locations ~10, ~16 and ~25 km upvalley of the terminal moraine complex, respectively (Fig. 4). To establish a robust chronology for the moraines, we obtained seven ^{10}Be exposure ages from the Twin Lakes Pinedale “younger” moraine, seven ^{10}Be exposure ages from the Pinedale “older” moraine, four ^{10}Be exposure ages from the Bull Lake moraine, and one exposure age (pending) from the recessional moraine that bisects the Twin Lakes Reservoir (Fig. 7).

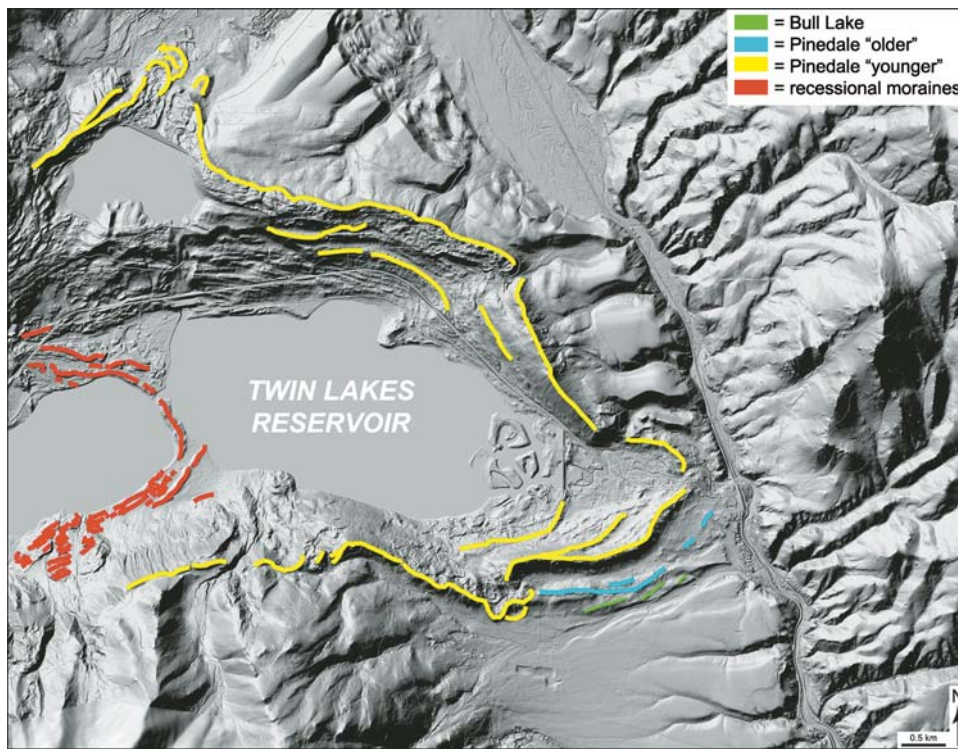


Figure 7. Twin Lakes study area. Colored lines demarcate prominent moraine crests drawn from the LiDAR base image. Green, blue, yellow, and orange are Bull Lake, Pinedale “older,” Pinedale “younger,” and recessional Pinedale moraines, respectively, mapped by Shroba et al. (2014).

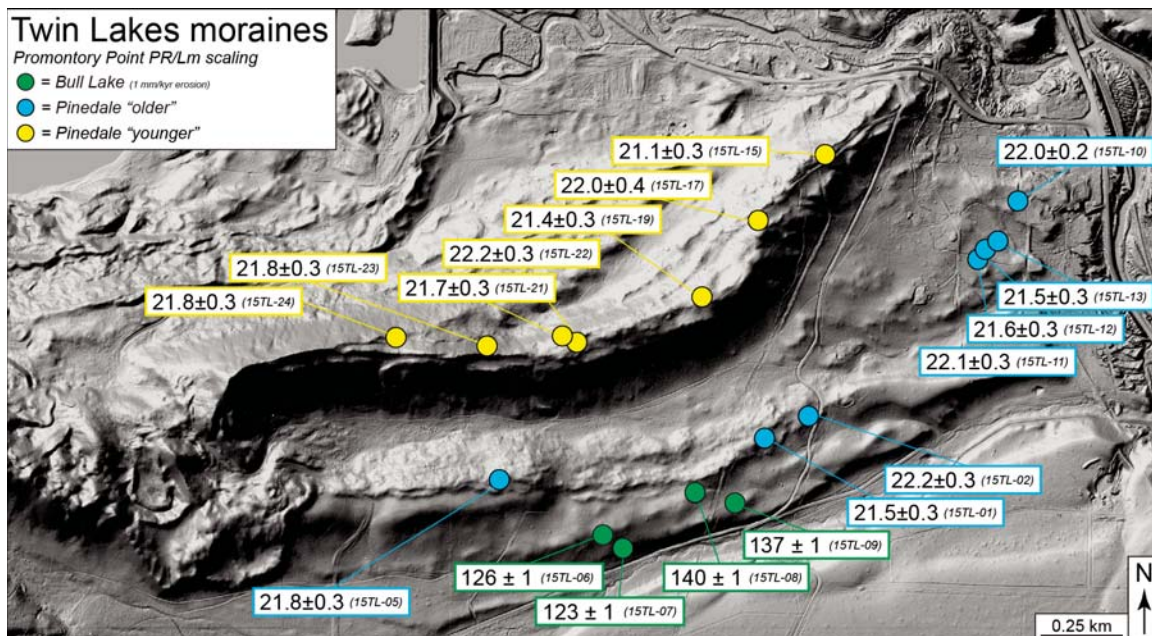


Figure 8. ^{10}Be exposure ages (ka) from the right-lateral moraine sequence southeast of the Twin Lakes Reservoir in the upper Arkansas valley. Green indicates ^{10}Be ages (in ka) from the Bull Lake moraine ($n = 4$), blue indicates ages from the Pinedale “older” moraine ($n = 7$), and yellow indicates ages from the Pinedale “younger” moraine ($n = 7$). Note that the Bull Lake ages were calculated assuming an erosion rate of 1 mm/k.y. Base image is a hillshade generated from LiDAR data acquired in September 2010. PR/Lm—production rate and scaling scheme.

Driving directions from Stop 1-1 to Stop 1-2. Depart Stop 1-1 to CR 10. Turn right onto CR 10 and stay on it to the bottom of the hill, and then take a left (east) onto the paved road, Colo. 82. Take a right (south) onto Lost Canyon Road (gravel road; County Road 30) across from the northern continuation of County Road 30 (old Colorado Highway 82). Drive ~0.3 km (~0.2 mi) to a turnout on the right associated with an old jeep road and park.

STOP 1-2. Hike through the Twin Lakes Moraine Sequence (Lunch Stop)

This stop includes a 1.5–3.0 km (~1–2 mi) hike across the right-lateral moraine sequence south of the Twin Lakes Reservoir to discuss new ^{10}Be exposure dating results and visit some of the dated boulders. *Please wear sturdy footwear and warm clothes.* The suggested hike starts at Stop 1-2 and begins by following an old jeep road up the sharp-crested Pinedale “younger” moraine (elevation gain is ~70 m) to highlight the geomorphology and visit five of the seven dated boulders. From the highest point, there is a view of the Pinedale “older” and Bull Lake moraines, and associated outwash deposits to the south, and the Twin Lakes Reservoir and recessional moraines to the northwest. After lunch atop the Pinedale “younger” moraine crest, we will hike southeast to the Pinedale “older” moraine. The hike will conclude with walking along the Bull Lake moraine ridge and visiting some of the dated boulders from this position. Vans will be shuttled for pick up. During this hike, we will discuss moraine morphostratigraphy, mapping of the glacial deposits, characteristics of dated boulders, and the newly developed ^{10}Be moraine chronology.

Prior to obtaining numerical ages of these deposits, the mapped distribution of surficial map units was based on field mapping, age-related soil properties (Shroba, 1977; Nelson and Shroba, 1998), and the interpretation of both LiDAR imagery acquired in 2010 and 1:40,000-scale, black-and-white aerial photographs taken in 1999 (Shroba et al., 2014). As a result of these activities, moraines enclosing the Twin Lakes Reservoir were mapped as three distinct map units (Figs. 6 and 7; Bull Lake, Pinedale “older,” and Pinedale “younger”; Shroba et al., 2014). Eighteen new ^{10}Be exposure ages from these three map units (moraine groups) provide direct age control for the tills that comprise these map units (Fig. 8). ^{10}Be ages boulders on the Bull Lake moraine range from 123 to 140 ka ($n = 4$) and are the first Bull Lake ages from the upper Arkansas valley. These ages are calculated with an assumed erosion rate of 1 mm/k.y., which has been estimated for granitic lithologies in this climate in previous work (Benedict, 1993; Gosse et al., 1995). The ages of the Pinedale “older” and “younger” moraines were determined to be 21.8 ± 0.3 ka ($n = 7$) and 21.7 ± 0.4 ka ($n = 7$), respectively, which are equivalent within 1σ uncertainty and average 21.8 ± 0.3 ka ($n = 14$; Fig. 8). These age calculations incorporated no rock surface erosion rate because many of the boulder surfaces exhibited original glacial smoothing (i.e., polish, striations).

The near-simultaneous deposition of the tills in the two Pinedale moraines south of Lake Creek was an unexpected finding. The distinct difference in morphology between the much larger, sharp-crested Pinedale “younger” moraine and the subdued and hummocky Pinedale “older” moraine suggests that Lake Creek glacier may have favored the extent demarcated with the Pinedale “younger” moraine and only advanced to the Pinedale “older” position for a short period of time. In this scenario, the till that comprises the bulky Pinedale “younger” moraine may have been largely deposited during a brief and late advance to the Pinedale “older” limit. Following recession from the Pinedale “older” limit, the glacier may have “topped off,” or draped, the outermost Pinedale “younger” moraine with debris, and subsequently retreated and deposited the till that comprises dozens of additional moraines between the outermost Pinedale “younger” moraine and the entrance to the steep valley of Lake Creek. This scenario is consistent with the recent suggestion of Anderson et al. (2014) that interannual variability of temperature and precipitation may lead to a complex sequence of short-term glacier margin fluctuations and resultant moraines, even under a period of constant climate.

Driving directions from Stop 1-2 to Stop 1-3. Depart Stop 1-2 by backtracking on Lost Canyon Road northward toward Colo. 82. At the intersection of Lost Canyon Road and Colo. 82, turn right (east) and drive 0.8 km (0.5 mi). Turn right (south) on U.S. 24 and drive 7.1 km (4.4 mi). The turnoff for Stop 1-3 will be on your left and is a dirt road with a circular pullout directly off U.S. 24. The left turn for this dirt road is ~0.16 km (0.1 mi) past the turnoff for County Road 390 (CR 390), which is north of the Clear Creek Reservoir.

STOP 1-3. Clear Creek Valley—Paleoglacier and Paleoflood Modeling

This stop allows for views of till of the Pinedale glaciation located east of the present course of the Arkansas River (Fig. 9). It has been hypothesized that the eastward-flowing Clear Creek glacier crossed the upper Arkansas River valley and abutted the opposing valley wall, damming a large lake upvalley (north) of the glacier terminus (Scott, 1975, 1984; McCalpin et al., 2010, 2012; Lee, 2010). If so, a lake(s) would have formed in the Arkansas River valley upriver from this site, and dam failure would have catastrophically released floodwaters laden with debris sourced from the granodiorite cliff and Clear Creek Pinedale lateral moraines. The presence of two boulder-rich terrace deposits downstream of Clear and Pine Creek valleys suggests that the hypothesized lake(s) would have drained catastrophically at least twice during Pinedale times (McCalpin et al., 2010).

At this stop, we will discuss numerical modeling of Sawatch Range paleoglaciers that has been conducted for Pine Creek, Clear Creek, Lake Creek, Halfmoon Creek, and Rock Creek valleys (see Appendix A), as well as paleoflood modeling of the Clear Creek glacier dam (see Appendix B).

Driving directions from Stop 1-3 to Stop 1-4. Depart Stop 1-3 and return to U.S. 24. Turn right onto U.S. 24, drive 0.16 km (0.1 mi), and turn left onto CR 390. Drive 8.5 km (5.3 mi) west on CR 390 and pull into a dirt pullout on the left.

STOP 1-4 (Optional). Clear Creek Valley—Glacially Sculpted Bedrock

This optional stop is just inboard of the range front and is the first outcrop of glacially sculpted bedrock that was sampled from the Clear Creek valley bottom (Fig. 9; Young et al., 2011). We plan to discuss the merits of dating boulders versus bedrock, and the deglaciation history of the upper Arkansas River valley. The ^{10}Be ages from the terminal Pinedale moraine indicate that the Clear Creek glacier retreated from its Pinedale maximum ~ 20.5 ka, whereas the ^{10}Be age from sculpted bedrock at this stop suggests ice remained near its maximum extent until $\sim 15.3 \pm 0.3$ ka (Fig. 9; Young et al., 2011).

Driving directions from Stop 1-4 to Stop 1-5. Turn around and drive 8.5 km (5.3 mi) east on CR 390 to U.S. 24. Turn right

(south) onto U.S. 24 and drive 7.1 km (4.4 mi). Turn left (east) onto a dirt road, cross the railroad tracks and Arkansas River, and turn right (south) onto County Road 371 (CR 371) after 0.4 km (0.25 mi). Drive for another 0.4 km (0.25 mi) and turn left onto a dirt road that cuts through the farm fields. Follow this road across the lower terrace and up onto the upper terrace and park in the dirt pullout.

STOP 1-5. Upper and Lower Flood Terraces

This site hosts some of the largest flood boulders in the Arkansas River valley, which range in size from ~ 1.5 to 15 m in length (McCalpin et al., 2010). We will discuss ^{10}Be dating of flood boulders on the two lowermost terraces, visit some of the dated flood boulders, and discuss how the Pinedale glaciation chronology reported here fits in with the broader patterns of Pinedale glaciation across the western United States. This will be the final geology-oriented stop of the field trip and thus the conclusion of the field trip prior to dinner and the drive back to Denver.

Initial work in the Arkansas River valley identified four flood terraces south of the Pine Creek moraine (Scott, 1975, 1984). The



Figure 9. Clear Creek valley and Clear Creek Reservoir in the upper Arkansas valley. Yellow and green dashed lines demarcate the outer limit of the Pinedale and Bull Lake moraines, respectively. Also shown are the approximate locations of ^{10}Be exposure ages (in ka) from boulders on the Pinedale moraine (yellow circles) and valley bottom bedrock (white circle; Young et al., 2011), the Arkansas River, and stop locations 1-3 and 1-4 (red circles).

two older (higher) terrace gravels were deposited during the early and middle Pleistocene, whereas the two younger (lower) terrace gravels comprise Pinedale-age outwash and are strewn with large boulders that were likely dislodged from moraines and valley walls during glacial outburst flood events. Flood deposits of Bull Lake age have not been identified in the Arkansas River valley, suggesting that glaciers in Lake Creek and Clear Creek valleys did not extend far enough east to impound the Arkansas River during the Bull Lake glaciation (Shroba et al., 2014). Here, the discussion focuses on the lower two terraces and their associated deposits (i.e., late Pleistocene terraces), which are referred to as the upper and lower terraces in the text and figures in this paper.

Upper terrace flood gravels are ~18 m thick and grade to ~5 m in thickness ~11 km downstream of the mouth of Pine Creek valley (Lee, 2010). Similarly, gravels associated with the lower terrace range from ~9 m in thickness near the mouth of Pine Creek to ~6 m downstream. At present, the upper terrace is ~10–20 m above the modern Arkansas River channel, with a median height of 15 m, and the lower terrace has a median height of ~6 m above the modern-day river channel (Lee, 2010). At this location, Young et al. (2011) obtained four ^{10}Be ages from boulders

positioned on the upper terrace and two ^{10}Be ages from boulders resting on the lower terrace (Fig. 10). ^{10}Be ages on the upper terrace date to 22.2 ± 1.0 ka, 20.6 ± 0.5 ka, 20.4 ± 0.7 ka, and 20.4 ± 0.9 ka, and boulders on the lower terrace date to 18.5 ± 0.5 ka and 18.4 ± 0.9 ka (Young et al., 2011; all previously published ^{10}Be ages are recalculated for consistency with the new data reported here), indicating both of the lower terraces are Pinedale in age as originally suggested by Scott (1975). Three ^{10}Be ages on the upper terrace average 20.5 ± 0.1 ka and correlate with the subset of ^{10}Be ages that averages 20.5 ± 0.2 ka from the Clear Creek Pinedale moraine, implicating the Clear Creek glacier as the most likely ice dam (Young et al., 2011). Two additional ^{10}Be ages were obtained from the lower terrace at a different site farther downvalley. These have a mean age of 19.5 ± 0.2 ka, which differs from the ages on the lower terrace closer to the location shown in Figure 10. Although a bimodal distribution of ^{10}Be ages exists across the sampling locations for the lower terrace, all ^{10}Be ages overlap with 2σ uncertainty and average 19.0 ± 0.6 ka ($n = 4$). It is possible that the two lower terrace sites, which are separated by ~5 km, represent deposits that differ in age. Although the age of the lower flood terrace is not expressed in the Clear

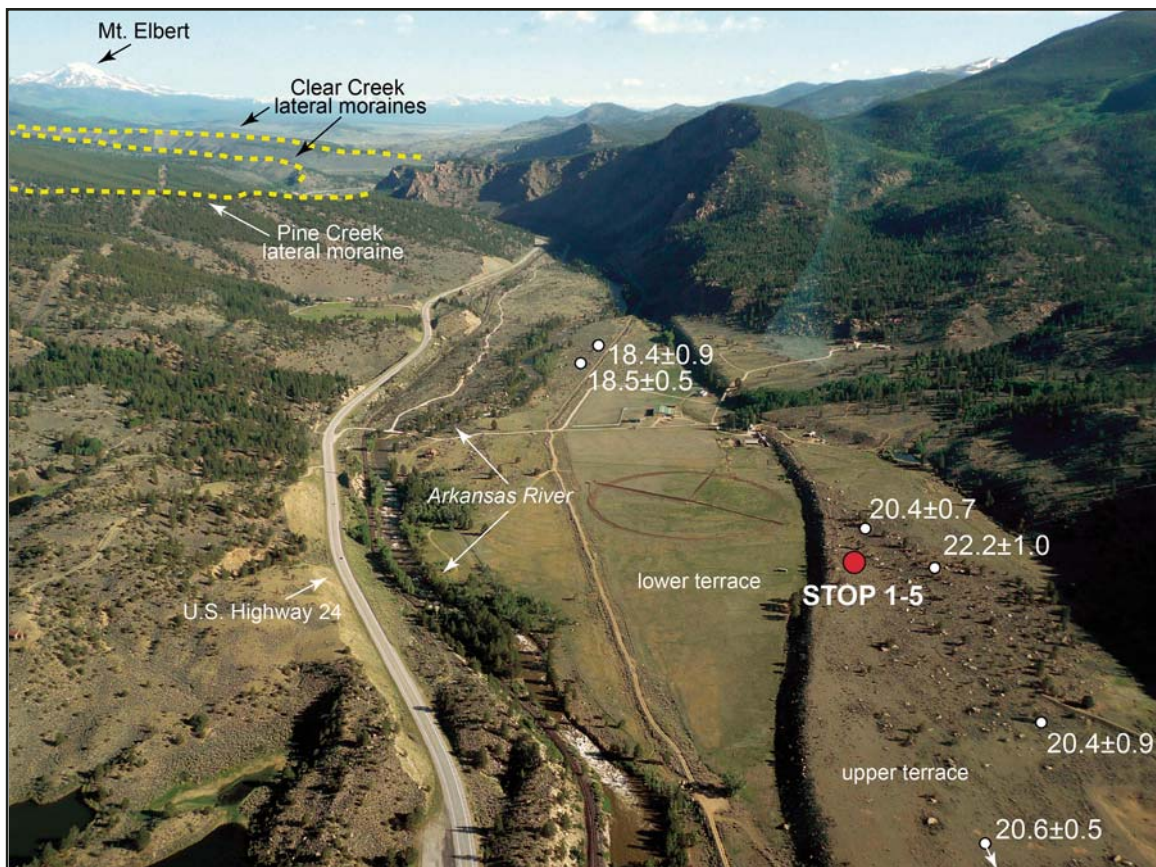


Figure 10. Upper and lower flood terraces in the upper Arkansas River valley. ^{10}Be ages (in ka) from flood boulders on each terrace are shown by the white circles (Young et al., 2011). Also shown are the Stop 1-5 location, modern-day Arkansas River, and Pine Creek and Clear Creek Pinedale moraine crests (yellow dashed lines). A dirt road with access to Stop 1-5 (red circle) has been created since this photo was taken and allows us to drive directly to the upper flood terrace.

Creek or Pine Creek moraine records, the close correspondence of the timing of moraine abandonment in Clear Creek valley and the upper flood terrace ages suggests that the Clear Creek glacier was near its Pinedale maximum and damming the Arkansas River ~ 19.0 ka (Young et al., 2011).

Although chronologies of Pinedale terminal moraines, and particularly Bull Lake moraines, are relatively sparse in the western United States, the timing of Pinedale maxima in Lake Creek valley is in accord with Pinedale maxima in neighboring valleys in the upper Arkansas River valley (24–16 ka) and across the western United States (24–15 ka). ^{10}Be ages from the Lake Creek Pinedale moraines (averaging 21.8 ± 0.3 ka; $n = 14$) are equivalent with the outermost Pinedale moraine located east of Mt. Massive and north of Twin Lakes (two ^{10}Be ages of 21.5 ± 0.5 ka and 21.8 ± 0.6 ka; Ruleman et al., 2013). ^{10}Be ages from Lake Creek valley indicate that Lake Creek glacier abandoned its Pinedale terminal position ~ 2 k.y. after the earlier advance of Pine Creek glacier (23.6 ± 1.4 ka; $n = 5$) and just prior to the earlier advance of Clear Creek glacier (20.5 ± 0.2 ka; $n = 3$; Young et al., 2011). Clear Creek and Pine Creek glaciers may have reoccupied their terminal positions at 19.0 ± 0.2 ka ($n = 3$) and 16.6 ± 1.0 ka ($n = 7$), respectively, which have not yet been identified in the Lake Creek valley moraine record (Young et al., 2011). However, the readvances documented in Clear Creek and Pine Creek valleys may be correlative to moraine(s) located directly inboard of the dated Pinedale “younger” moraine in Lake Creek valley (Fig. 7).

Within the western United States, moraine chronologies reveal significant age differences among Pinedale terminal moraines (24–15 ka; e.g., Licciardi et al., 2004; Licciardi and Pierce, 2008; Ward et al., 2009). For example, Pinedale terminal moraines in the northern Rocky Mountains are ~ 5 k.y. younger than Pinedale terminal moraines in the Wind River Range and at some locations in Colorado (Gosse et al., 1995; Licciardi et al., 2004; Thackray et al., 2004; Benson et al., 2005). Multiple hypotheses for the asynchronous timing of Pinedale termini have been proposed, including altered westerly atmospheric flow driven by North American ice sheets (Licciardi et al., 2004; Thackray et al., 2004, 2008), enhanced moisture delivery to downwind mountain ranges from the Great Basin paleolakes (Laabs et al., 2009), and the influence of nonclimatic factors intrinsic to individual glacier systems (i.e., hypsometry, response time; Licciardi and Pierce, 2008; Ward et al., 2009; Young et al., 2011).

Glacier fluctuations are ultimately driven by climate change; therefore, significant age differences between widely spaced Pinedale terminal moraines across the western United States are likely a result of spatially variable climate forcing (e.g., Licciardi et al., 2004; Laabs et al., 2009). However, the exact position of a glacier terminus is filtered by nonclimatic factors intrinsic to each glacier valley system (Young et al., 2011), as well as the influence of interannual climate variability (Anderson et al., 2014). Our findings from Lake Creek valley, which yield a terminal moraine age that differs from other terminal moraine ages in adjacent valleys, reinforce Young et al.’s (2011) argument that local nonclimatic factors (i.e., hypsometry and glacier response

time) can contribute to asynchrony of glacial maxima within each individual region or mountain range. Our new chronology from the Lake Creek valley contributes toward elucidating broader-scale patterns of Pinedale glaciation in the western United States, and may provide insight on the role and relative importance of climate forcing, such as insolation or carbon dioxide concentration (Shakun et al., 2012), in driving glacier change during the last deglaciation.

Driving directions from Stop 1-5 to Stop 1-6. Return to U.S. 24 (~ 1.4 km [~ 0.9 mi]). Continue southward on U.S. 24 for 16.9 km (10.5 mi). Turn left on East Main Street and drive 0.8 km (0.5 mi). Take another right onto South Main Street (baseball fields will be on your left). Drive down South Main Street for ~ 0.5 km (~ 0.3 mi) until the intersection with Riverpark Road. The Eddyline Restaurant is on the corner.

STOP 1-6. Dinner in Buena Vista, Colorado (Eddyline Brewery and Restaurant)

Dinner is scheduled for 7 p.m. at the Eddyline Restaurant at 926 South Main, Buena Vista, Colorado 81211 [<http://eddylinerestaurant.com/>; 719-966-6000]. We hope to return to Denver by ~ 11 p.m.

Driving directions from Stop 1-6 to the Colorado Convention Center, Denver, Colorado. Depart the Eddyline Restaurant parking lot and turn right (west) onto Riverpark Road. Turn right onto Ramsour Road and then make a quick left onto Arizona Street (County Road 313). Follow Arizona Street (County Road 313) for 3.2 km (2 mi), and then turn left (east) onto U.S. 24 (U.S. Highway 285). Continue northward on U.S. 285 for 191.5 km (119 mi). Take the Colorado 470 exit toward I-70/Grand Junction and then merge (west) onto Colorado Highway 470 (Colo. 470; 8.5 km [5.3 mi]). Exit toward I-70 W/I-70 E/Grand Junction/Denver (0.3 km [0.2 mi]), keep right at the fork and follow signs for I-70 E/Denver and merge onto I-70 E. Drive 2.1 km (1.3 mi) and take Exit 261 for U.S. 6 (6th Avenue). Continue eastward on U.S. 6 for 12.9 km (8.0 mi), and then take the exit north toward I-25. Merge onto I-25 (2.1 km [1.3 mi]) and then take Exit 210A for U.S. Highway 40 (U.S. Highway 287 or Colfax Avenue; 0.6 km [0.4 mi]). Turn right onto Colfax Avenue (1.1 km [0.7 mi]), which will continue as Stout Street and then 14th Street. Arrive at the Colorado Convention Center.

ACKNOWLEDGMENTS

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early career professionals to attend our field trip. We would like to thank the local landowners that allowed us access to their land for ^{10}Be sampling and conducting this field trip. We also thank Bob Finkel and Susan Zimmerman of Lawrence Livermore National Laboratory for ^{10}Be measurements. Finally, we thank Cal Ruleman, Charles Porreca, and Joseph Tulenko for assistance in the field; William Caffee and Vincent Carsillo for helping with lab work; and Theodore Brandt for generating the hillshades shown in Figures 7 and 8.

APPENDIX A

Numerical Modeling of Sawatch Range Paleoglaciers

Eric Leonard and Charles Russell (Colorado College)

In addition to revised geologic mapping and a new ^{10}Be chronology, we also carried out numerical glacier-modeling experiments to constrain the patterns of glaciation and their climate controls in the upper Arkansas River valley. We employed a 2-D coupled energy/mass balance and flow model (Plummer and Phillips, 2003; Laabs et al., 2006; Leonard et al., 2014) to investigate both ice dynamics and climate conditions during the Pinedale glaciation in this region. We paid particular attention to the magnitude and rates of change associated with subsequent deglaciation from Rock Creek valley to Pine Creek valley (Fig. A1). The timing and rate of deglaciation is constrained by ^{10}Be exposure ages of glacially sculpted bedrock in Lake Creek, Clear Creek, and Pine Creek drainages (Fig. 4; Young et al., 2011). Modern

climate input to the model is derived from 1971 to 2000 A.D. monthly values from PRISM (Parameter-elevation Regression on Independent Slopes Model) grids clipped to the eastern flank of the Sawatch Range, as well as from nearby meteorological and SNOTEL (Snowpack Telemetry) station records. We model mass balance and flow only for the eastern flank of the Sawatch Range and do not consider possible ice flow across the range divide in either direction.

Figure A1-A shows the modeled extent of glaciers on the eastern flank of the Sawatch Range during the Pinedale maximum. Assuming no change from modern precipitation, this ice extent is associated with a temperature depression of 5.4°C . If precipitation during the Pinedale maximum was more or less than that of the present, temperature depressions smaller or larger than 5.4°C , respectively, would have been necessary to sustain the glaciers. Model output indicates that equilibrium-line altitudes (ELAs) for individual glaciers during the Pinedale maximum ranged from ~ 3430 to ~ 3480 m. These values are ~ 1000 – 1100 m lower than the theoretical modern snowline along the eastern flank of the Sawatch Range suggested by degree-day modeling (Brugger, 2010). The model output closely resembles the mapped Pinedale maximum ice extent in each of the five main valleys of the model domain (Fig. A1-A; Lee, 2010; McCalpin et al., 2012; Shroba et al., 2014). Consistent with recent mapping, the Lake Creek glacier closely approaches, but does not reach the Arkansas River, while the Clear Creek and Pine Creek glaciers flow across the present course of the Arkansas River and impinge on the bedrock wall on the eastern side of the valley. Modeled paleoglaciers in Clear Creek valley, and especially in Halfmoon Creek valley, slightly overtop their mapped Pinedale lateral moraines, but closely match their terminal positions (Fig. A1-A).

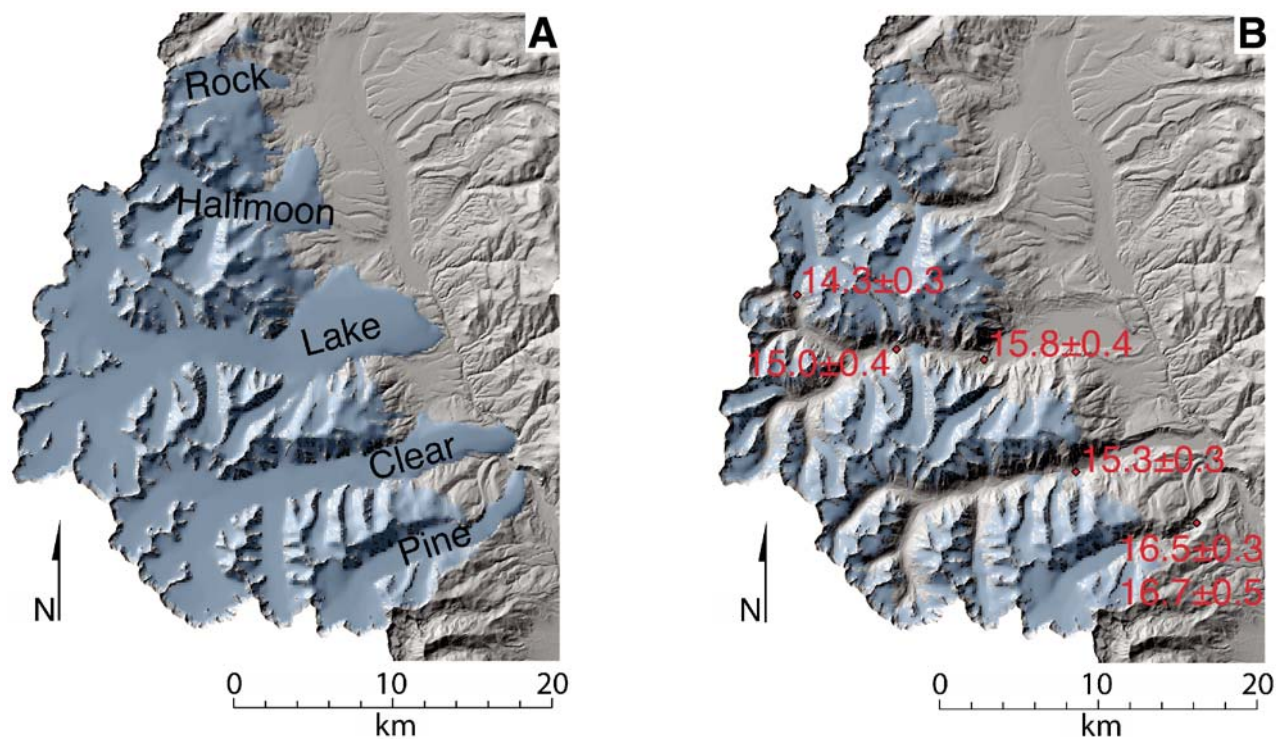


Figure A1. (A) Modeled extent of glaciers on the east flank of the Sawatch Range with a temperature depression of 5.4°C from modern temperatures and no change from modern precipitation. (B) Modeled extent of glaciers on the eastern flank of the Sawatch Range with a temperature depression of 4.2°C from modern temperatures and no change from modern precipitation. Red diamonds indicate ^{10}Be sample sites from glacially polished bedrock and associated exposure ages in ka (Young et al., 2011).

We combine ^{10}Be ages of bedrock located upvalley from Pinedale end moraines in Lake Creek valley (Fig. 4; Young et al., 2011) with our modeled glacier history to interpret the timing and patterns of change in ice extent and climate during deglaciation (Fig. A1-B). In addition, we recently sampled polished-bedrock outcrops from cirque lips farther upvalley, very close to the range divide, but those ages are still pending and thus not included in these model runs. The most upvalley and youngest of the available bedrock ages indicates that extensive deglaciation of Lake Creek valley had occurred by 14.3 ± 0.3 ka (Fig. 4; Young et al., 2011). At this time, the glacier margin had receded more than 25 km from its Pinedale moraine (or more than 75% from its Pinedale maximum length). Using the Lake Creek glacier margin as a control, we modeled the likely extent of ~ 14.3 ka glaciers along the eastern flank of the range (Fig. A1-B). Model results indicate that the Clear Creek glacier would have receded even more (~ 26 km from its Pinedale moraine, or 93% of its Pinedale maximum length) than the Lake Creek glacier, with the main valleys in the Clear Creek drainage essentially ice free at that time. By contrast, more limited recession occurred in Pine Creek valley (~ 7.6 km from its Pinedale moraine, or 43% of its Pinedale maximum length). Surprisingly, this rather extensive retreat from Pinedale terminal moraines by ~ 14.3 ka is forced by a temperature increase of only 1.2°C from Pinedale maximum temperature depression from modern of 5.4°C . Model-derived ELAs for ~ 14.3 ka glaciers are ~ 3650 m, which is only 200 m higher than the Pinedale ELA, and ~ 800 m below the modern snowline.

Young et al. (2011) noted that the timing of the Pinedale maximum of the Pine Creek glacier and its behavior following its maximum stand differed from that of the Clear Creek and Lake Creek glaciers, with the former arriving at its Pinedale maximum position earlier and remaining at or very close to that position longer than the others. This observation is consistent with our model results. Young et al. (2011) attributed this dissimilar response to differences in slope and hypsometry between the valleys occupied by the glaciers, differences also observed in our analysis.

APPENDIX B

Discharge Estimates for a Glacial-Lake Outburst Flood in the Upper Arkansas River Valley

Keith Brugger (University of Minnesota, Morris)

The possibility of glacial-lake outburst flooding in the upper Arkansas River Valley was first proposed by Scott (1975) based on the presence of large, presumed flood-transported boulders. Scott (1984) later attributed flooding to failure of an ice dam created by ice flowing out of Pine Creek valley that impounded water of the upper Arkansas River catchment. Scott (1984) also recognized four episodes of flooding based on the association of large boulders and outwash-terrace gravel, two of which occurred during Pinedale glaciation. Subsequent work by Lee (2010) and McCalpin et al. (2010) provided additional evidence for a Pinedale-age ice-dammed lake and its eventual catastrophic drainage, including the possibility of preserved shoreline gravels, potential lacustrine sediment, and ice-rafted boulders. The authors suggest these deposits are equivocal and acknowledge that scant evidence exists for the lake and/or its level.

Lee (2010) concluded that the main ice dam during Pinedale glaciation was created by the advance(s) of the Clear Creek glacier (Fig. B1); this does not, however, preclude minor ice damming from Pine Creek glacier. Moreover, Young et al. (2011) demonstrated a close temporal relationship between the age of the Pinedale terminal moraine in Clear Creek valley ($\sim 20.5 \pm 0.2$ ka) and flood boulders on an upper flood terrace dated at 20.9 ± 0.9 ka. A lower, boulder-strewn terrace deposit dated at 19.0 ± 0.6 ka cannot unequivocally be related to dated moraines in the area, but may document a younger advance and ice damming by the Clear Creek glacier that is not represented in the moraine record (Young et al., 2011).

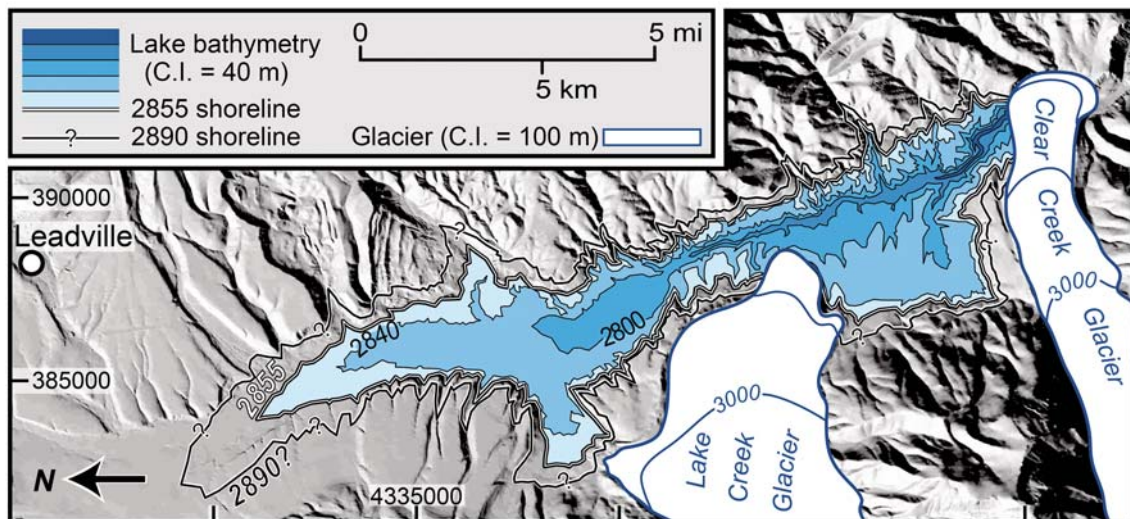


Figure B1. Hypothesized extent and bathymetry of lake waters impounded by the ice dam created by the advance of the Clear Creek glacier ca. 20.5 ka (Young et al., 2011). A possible highstand of the lake is also shown.

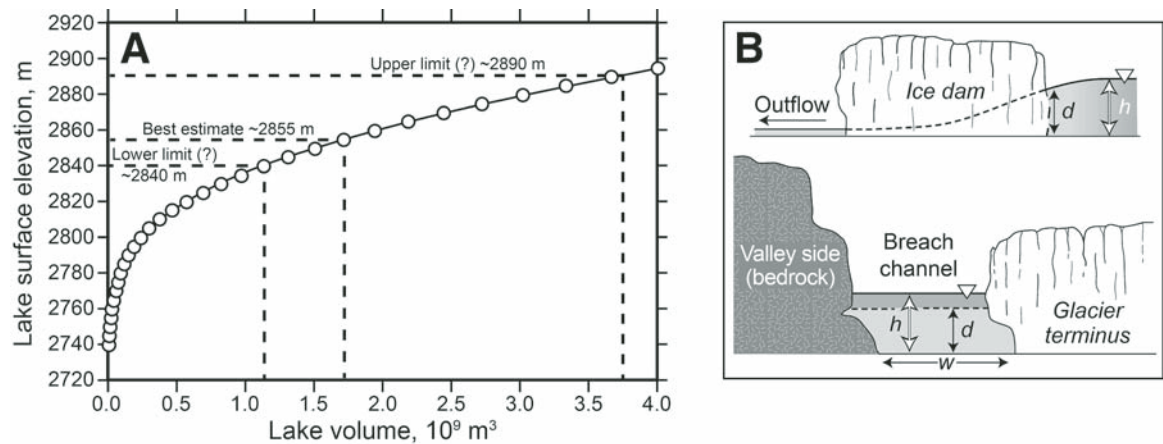


Figure B2. (A) Variation of lake volume and surface elevation (adapted from Bush, unpublished). (B) Depiction of breach geometry. d —depth; h —height; w —width.

Lake Levels and Volume

Based on reconstructions of the terminus of the Clear Creek glacier at its Pinedale extent, the paleolake level is estimated to have been ~2855 m (Figs. B1 and B2-A), ~150 m above the modern channel of the Arkansas River at this location. This is consistent with Lee’s (2010) estimate, and corresponds to a lake volume of $1.72 \times 10^9 \text{ m}^3$ (Fig. B2-A). Using reasonable values for river discharge into the lake (e.g., that of the modern Arkansas River and multiples thereof), filling times could have been on the order of years to a few decades, suggesting that the lake(s) may have been transient. Comparable filling times have been reported for modern ice-dammed lakes having volumes of similar magnitude (e.g., Anderson et al., 2003; Trabant et al., 2003; Huss et al., 2007; Mayr et al., 2014). This may explain the lack of shorelines preserved around the hypothesized paleolake.

Ice-Dam Failure Mode and Modeling

Possible triggers of outburst floods include flotation of the glacier’s terminus (Lee, 2010), initial establishment of a localized subglacial tunnel (Clarke, 1982) or subaerial channel (“breach”; Walder

and Costa, 1996), or failure of a moraine dam. Given the age correspondence of flood boulders and the LGM position of the Clear Creek glacier, it seems more likely that water was impounded by ice, and thus only the subaerial breach is used to estimate paleoflood discharges. It is noteworthy here to point out that Clarke’s (1982) model of subglacial tunnel enlargement yields significantly lower discharge estimates. In the model, it is implicitly assumed that there was (1) a single ice dam, and (2) no mechanical failure (e.g., fracture, calving, and/or collapse) of the ice dam. The model—a “minimum model” wherein not all dissipated energy is used for melting and widening of the breach—is only briefly summarized here; additional details and derivations can be found in Walder and Costa (1996). While empirical and other relationships (discussed subsequently) are useful in estimating peak flood discharges, the advantage of modeling is that flood hydrographs are generated, and the influence of other factors that potentially control the discharge can be explored.

Outflow from the impounded lake is controlled by breach geometry (Fig. B2-B) determined by channel width w and depth d at the entry to the breach. Widening of the breach through time t is

TABLE B1. RESULTS OF THE SIMULATIONS

Constants	Lake level (m)	Lake volume, V_0 (m^3)	f	θ_L ($^{\circ}\text{C}$)	Q_{peak} ($\text{m}^3 \text{ s}^{-1}$)
$\rho_w = 1000 \text{ kg m}^{-3}$					
$\rho_i = 900 \text{ kg m}^{-3}$					
$g = 9.8 \text{ m s}^{-2}$					
$L = 333.5 \text{ kJ kg}^{-1}$					
$k_w = 0.56 \text{ W m}^{-1} \text{ deg}^{-1}$					
$\eta = 1.79 \times 10^{-3} \text{ kg m}^{-1} \text{ s}^{-1}$					
$c_1 = 0.07$					
$c_2 = 0.23$					
$w_0 = 0.5 \text{ m}$					
$u_i = 0 \text{ m s}^{-1}$					
	2855	1.72×10^9	0.05	1.0	46,050
				5.0	47,070
				0.0	45,780
			0.01	1.0	21,020
			0.10	1.0	64,740
	2840	1.14×10^9	0.05	1.0	31,270
	2890	3.77×10^9	0.05	1.0	90,890

$$\frac{dw}{dt} = -u_i + c_1 \left(\frac{\rho_w}{\rho_i} \right) \frac{f}{L} (gh)^{1.5} + c_2 \frac{k_w (\theta_i - \theta_w)}{\rho_i L} \left(\frac{g^{0.5} \rho_w w h^{0.25}}{\eta (4/3h + w)} \right)^{0.8}, \quad (1)$$

where u_i is the advance terminus due to ice flow, ρ_w and ρ_i are the densities of water and ice, respectively, f is a friction factor within the breach channel, L is the latent heat of fusion for ice, g is gravitational acceleration, h is the water height above the outlet base, k_w is the thermal conductivity of water, θ_i and θ_w are the water and ice temperatures, respectively, η is the viscosity of water, and c_1 and c_2 are constants. In short, the first term describes closure of the breach, the second term reflects the breach widening due to viscous dissipation, and the third term describes the widening via stored thermal energy in the water.

Flow velocity at the breach is $(gd)^{0.5}$ and $d = 2/3h$, thus discharge Q is

$$Q = (2/3h)^{1.5} g^{0.5} w. \quad (2)$$

Assuming negligible water loss by evaporation over the lake surface and little gain by inflow, both reasonable given the short duration of the resulting floods, the discharge through the breach is related to the change in lake volume

$$\frac{dV}{dt} = Q. \quad (3)$$

Equations 1 and 3 are coupled by a relationship between water height h and lake volume V (Fig. B2-A). Using the initial conditions of $V = V_0$, that is full lake volume, and a small but nonzero existing breach width w_0 , Equations 1 and 3 are simultaneously integrated using a 4th-order Runge-Kutta technique to simulate the flood hydrograph for the resulting outburst flood and estimate the associated peak discharge Q_{peak} .

Constants for all simulation were: $\rho_w = 1000 \text{ kg m}^{-3}$, $\rho_i = 900 \text{ kg m}^{-3}$, $g = 9.8 \text{ m s}^{-2}$, $L = 333.5 \text{ kJ kg}^{-1}$, $k_w = 0.56 \text{ W m}^{-1} \text{ deg}^{-1}$, $\eta = 1.79 \times 10^{-3} \text{ kg m}^{-1} \text{ s}^{-1}$, $c_1 = 0.07$, $c_2 = 0.23$, $w_0 = 0.5 \text{ m}$, and $u_i = 0 \text{ m s}^{-1}$. A control simulation was run with “preferred” values set as follows: $V_0 = 1.72 \times 10^9 \text{ m}^3$ corresponding to $h = 2855 \text{ m}$, $\theta_L = 1^\circ \text{C}$, $f = 0.05$, $w_0 = 0.5 \text{ m}$. Closure of the breach due to ice flow was assumed to be negligible. Subsequent simulations focused on how lake volume, water temperature, and friction within the breach channel affected Q_{peak} and the duration of flood flow. Values for θ_L ranged from 0 to 5°C and those for f from 0.01 to 0.1, based on those reported in the literature or used in similar modeling (Clarke, 1982; Walder and Costa, 1996; Björnsson, 2004; Anderson et al., 2003; Dussallant, et al., 2010). Results are partially summarized in Table B1 and the flood hydrographs are shown in Figure B3.

Lake water temperature plays a minor role in widening of the breach (Table B1), indicating the breach widening is dominated by viscous dissipation of heat. Greater channel roughness could have generated peak discharges in excess of $60,000 \text{ m}^3 \text{ s}^{-1}$ (Fig. B3) by virtue of increasing “melt erosion,” again by viscous dissipation. Conversely, a smoother breach would lead to substantial reductions in Q_{peak} values, ca. $20,000 \text{ m}^3 \text{ s}^{-1}$. Model results are insensitive to reasonable values of w_0 . At the extreme, Q_{peak} values could have been on the order of $\sim 90,000 \text{ m}^3 \text{ s}^{-1}$ for a plausibly greater lake volume (Fig. B2) or higher when combined with greater channel roughness.

In summary, model results for a subaerial breach of the Clear Creek glacier terminus suggest peak discharges in the upper Arkansas River valley may have been on the order of $45,000 \pm 20,000 \text{ m}^3 \text{ s}^{-1}$. To put these discharges in perspective, the mean annual discharge of the Congo is $\sim 40,000 \text{ m}^3 \text{ s}^{-1}$. In contrast, the long-term mean annual discharge of the Arkansas River near Salida is $\sim 18 \text{ m}^3 \text{ s}^{-1}$ (data from <http://nwis.waterdata.usgs.gov>, accessed March 2016). High discharges

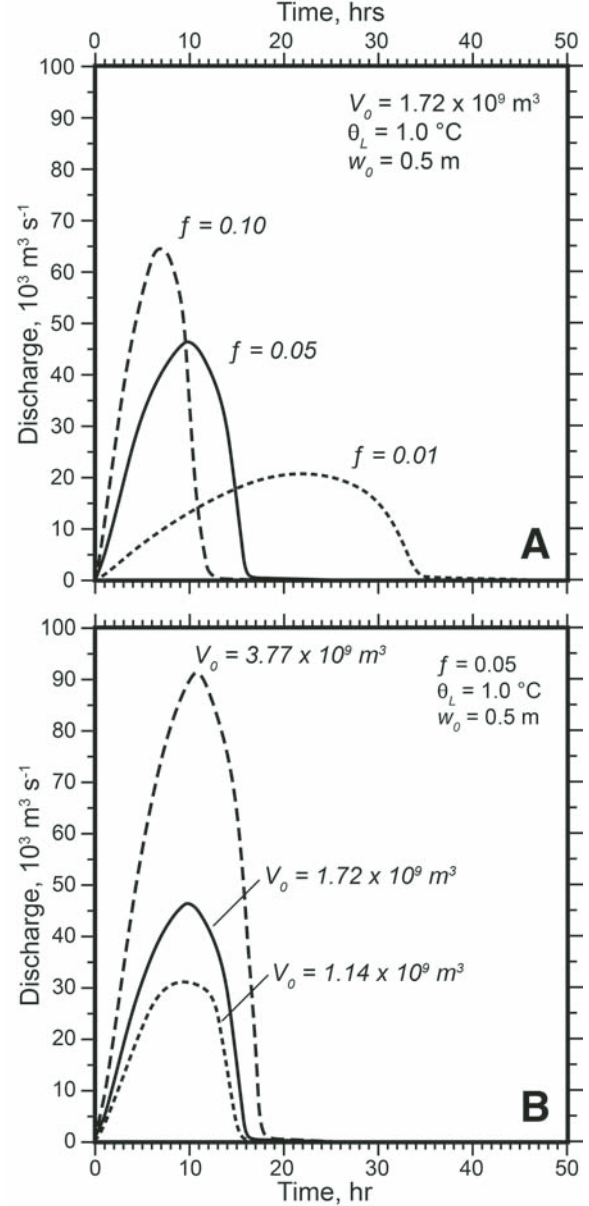


Figure B3. Simulated flood hydrographs. (A) Effects from varying channel roughness. (B) Effects from variations in lake volume.

would have persisted for ~ 15 – 30 h and nearly all water impounded by the ice dam would have drained. The overall shape and duration of the hydrographs (Fig. B3) and the magnitude and timing of Q_{peak} in the simulations compare favorably with those recorded for modern outburst floods draining lakes of comparable volume via subaerial breach (e.g., Russell Lake, Alaska; Trabant et al., 2003).

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