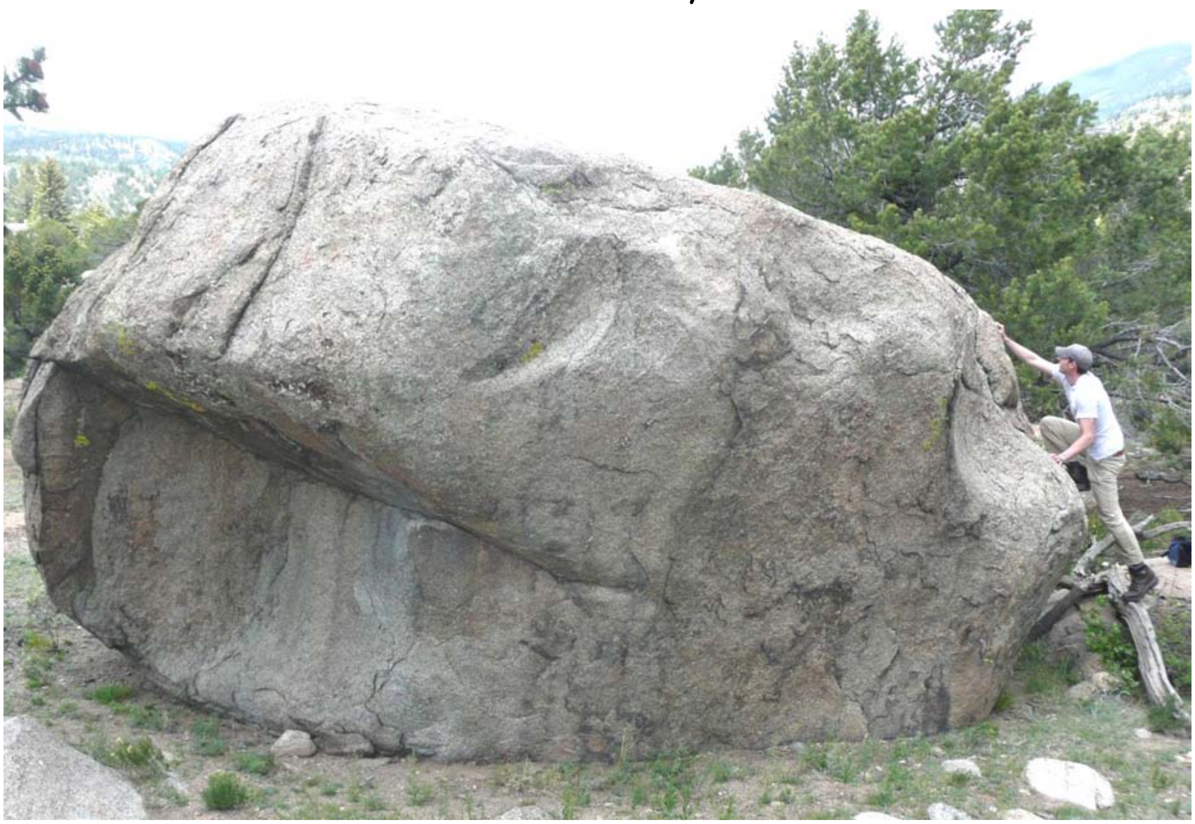
Quaternary Geology and Geochronology of the Uppermost Arkansas Valley— Glaciers, Ice Dams, Landslides, Floods

October 29-30, 2010



Sampling an outburst flood boulder

Edited by J.P. McCalpin

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General Itinerary: a 257-mile loop starting and ending at the Colorado Convention Center, downtown Denver, CO.

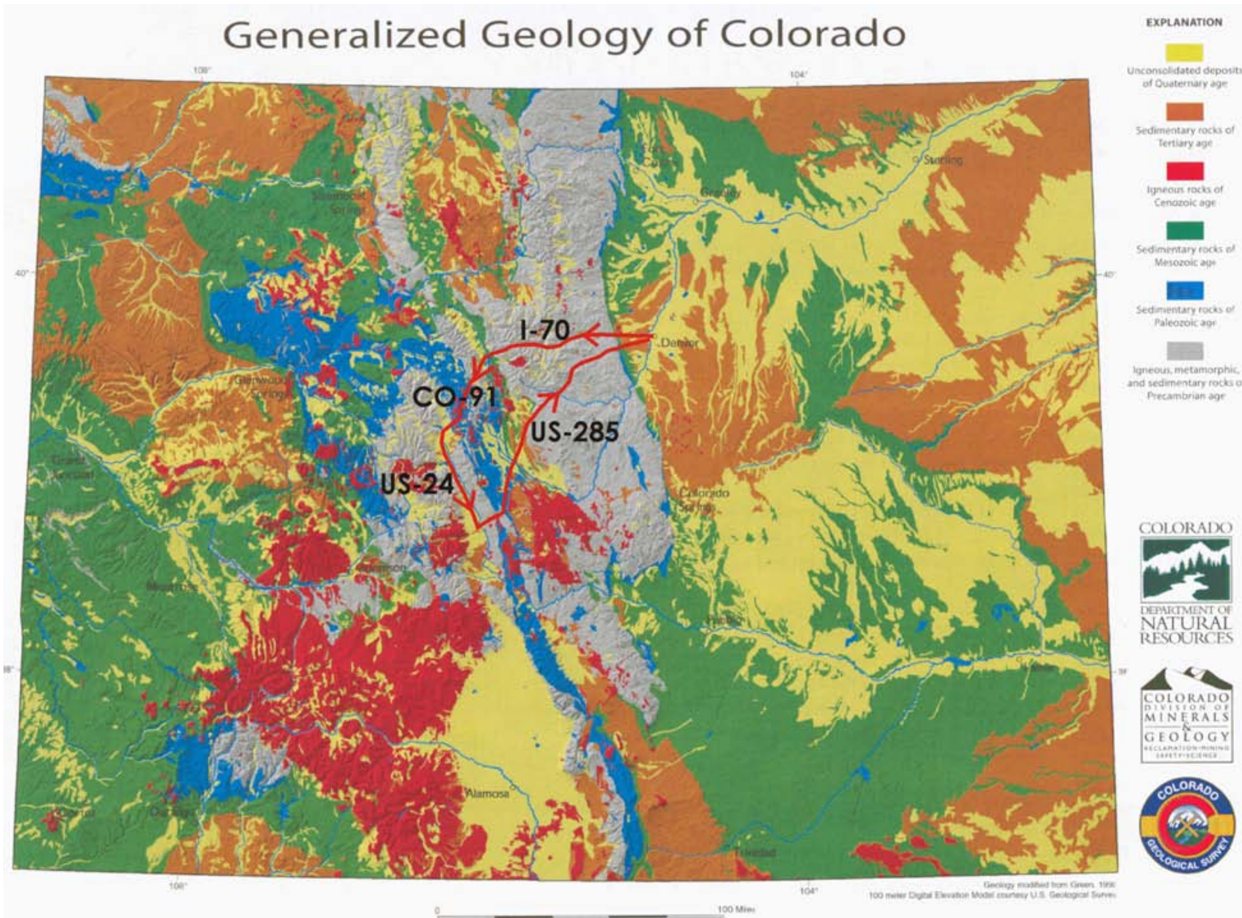


Figure 1. Route of field trip (red line with arrows) superimposed on the Geologic Map of Colorado postcard. The Day 1 route goes west from Denver on I-70 to Copper Mountain, then SW on CO-91 to Leadville, CO (100 miles). Overnight stay in Leadville. The Day 2 route takes US-24 south from Leadville to Buena Vista (34 miles), then NE on US-285 back to Denver (123 miles). The detailed road log begins at Frisco, CO (72 miles west of Denver on I-70).

Introduction and Acknowledgements

This field trip to the uppermost Arkansas Valley of central Colorado is based mainly on work performed since 2008, in an area with a long history of bedrock mapping by renowned geologists of the U.S. Geological Survey beginning in the 1880s (e.g. Emmons, 1886). The Quaternary features of the region were first described by Capps in 1909. Despite over 100 years of intermittent geologic studies here, there were still new discoveries to be made in the past few years.

Day 1 of the trip emphasizes landslides (in the morning) and glacial deposits (in the afternoon). The “take-home messages” of the morning’s stops are: (1) that previously published bedrock maps of the area (e.g. Tweto, 1974a) severely underestimated the extent of landsliding in the region, and (2) there is a continuum of gravitational deformation between sackung, incipient landsliding, and full landsliding (which creates a rubbilized landslide deposit). Areas subjected to the first two stages appear to be intact bedrock at first glance, and have been mapped as such on almost all published geologic maps. This was done without regard for the rather obvious post-glacial landforms (downslope-facing scarps, upslope-facing scarps, linear closed troughs, closed depressions) indicating extensional spreading/toppling at their heads.

On the afternoon of day 1 we look at the Quaternary geology around the town of Leadville, emphasizing the moraines and outwash of Pinedale, Bull Lake, and pre-Bull Lake episodes. The subsurface geology of unconsolidated deposits was described by Emmons (1886) and Emmons et al. (1927) based on mine shafts and tunnels, but it is difficult to relate what they saw to what we can see today at the surface.

Day 2 concentrates on the glacial ice dams that dammed the upper Arkansas River during various ice advances, the lake that formed upstream from the dam (Three Glaciers Lake), and the flood gravels deposited downstream from the ice dams when they failed catastrophically. The evidence for the flood was first described by Scott (1975, 1984), but he only speculated on the location and height of the ice dams, and he never identified any features associated with the ice-dammed lake. His dating of the floods had to rely mainly on relative-age dating and correlation; neither luminescence dating nor cosmogenic surface-exposure dating had been developed at the time of his studies.

The “new” flood story was outlined in large part by Keenan Lee, working independently of the STATEMAP mappers (McCalpin and crew), the moraine mappers and hydrologic calculators Eric Leonard and crew), and the cosmogenic daters, (Briner, Young and crew). USGS geologists Cal Ruleman and Ralph Shroba got all the parties together in 2008. We now have some more quantitative answers about the sequence and numerical ages of moraine deposition, river damming, lake formation, dam failure, and outburst flooding. However, there are still unresolved issues, for which we solicit comments, and hopefully, brilliant solutions, from our field trip attendees.

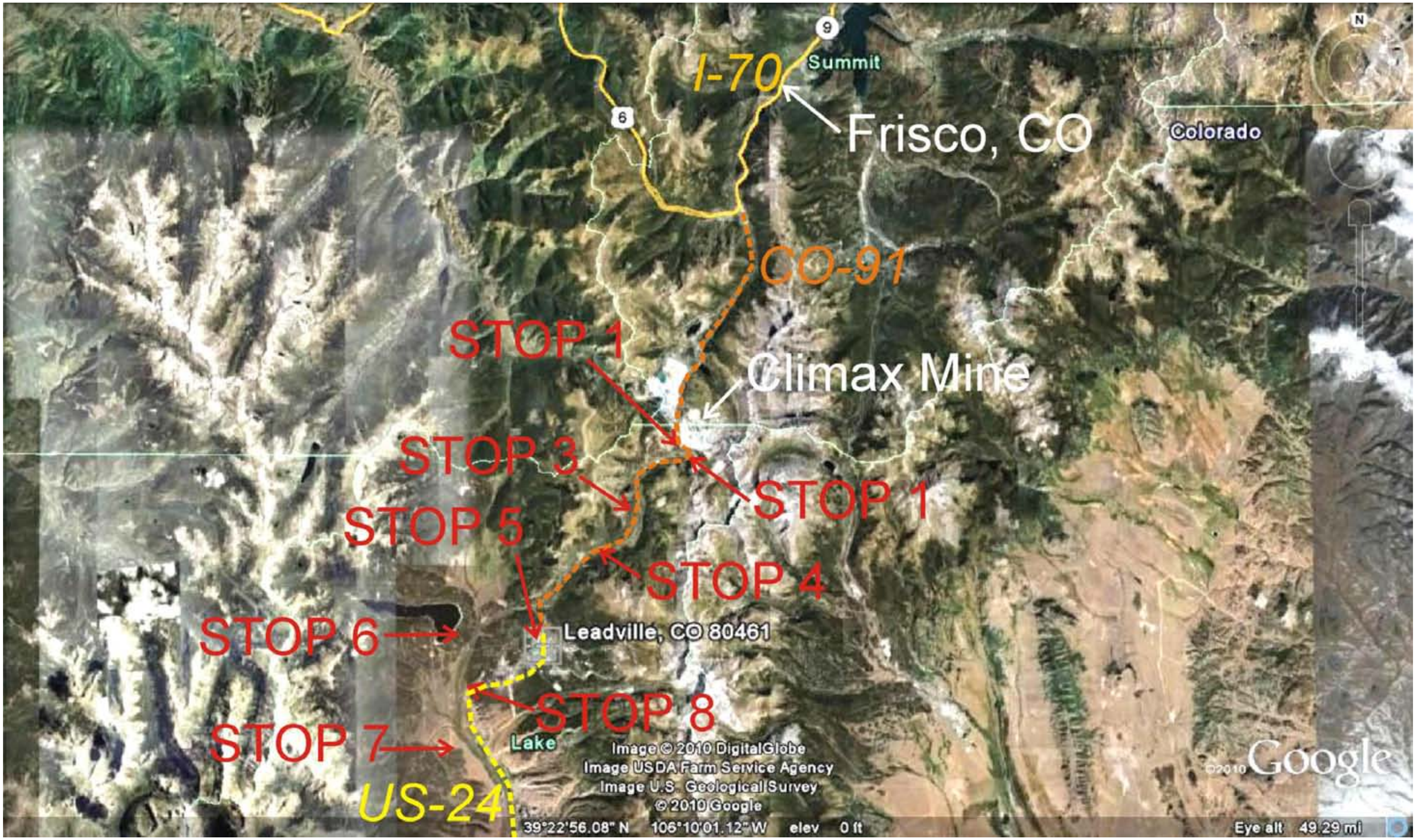


Figure 2. Field trip STOPS of DAY 1 plotted on a Google Earth image; north is at top.

Day 1 (Friday, Oct. 29, 2010): drive west from downtown Denver on US-6 to I-70; continue west on I-70 and up Clear Creek, passing through the mountain towns of Idaho Springs, Georgetown, and Silver Plume; cross beneath the Continental Divide via the Eisenhower Tunnel; descend I-70 to Silverthorne; take Exit 203 to Frisco for a coffee stop; ROAD LOG BEGINS HERE.

From Frisco continue west on I-70 to Copper Mountain; exit I-70 onto CO-91 and ascend to Fremont Pass [STOP 1]; make another 6 stops between here and Leadville, and in the vicinity of Leadville. Day 1 involves about 120 miles of driving, almost all on paved roads.

In the evening, the group will have an hour-long "Happy Hour" slide show at the National Mining Museum in Leadville, followed by a catered dinner.

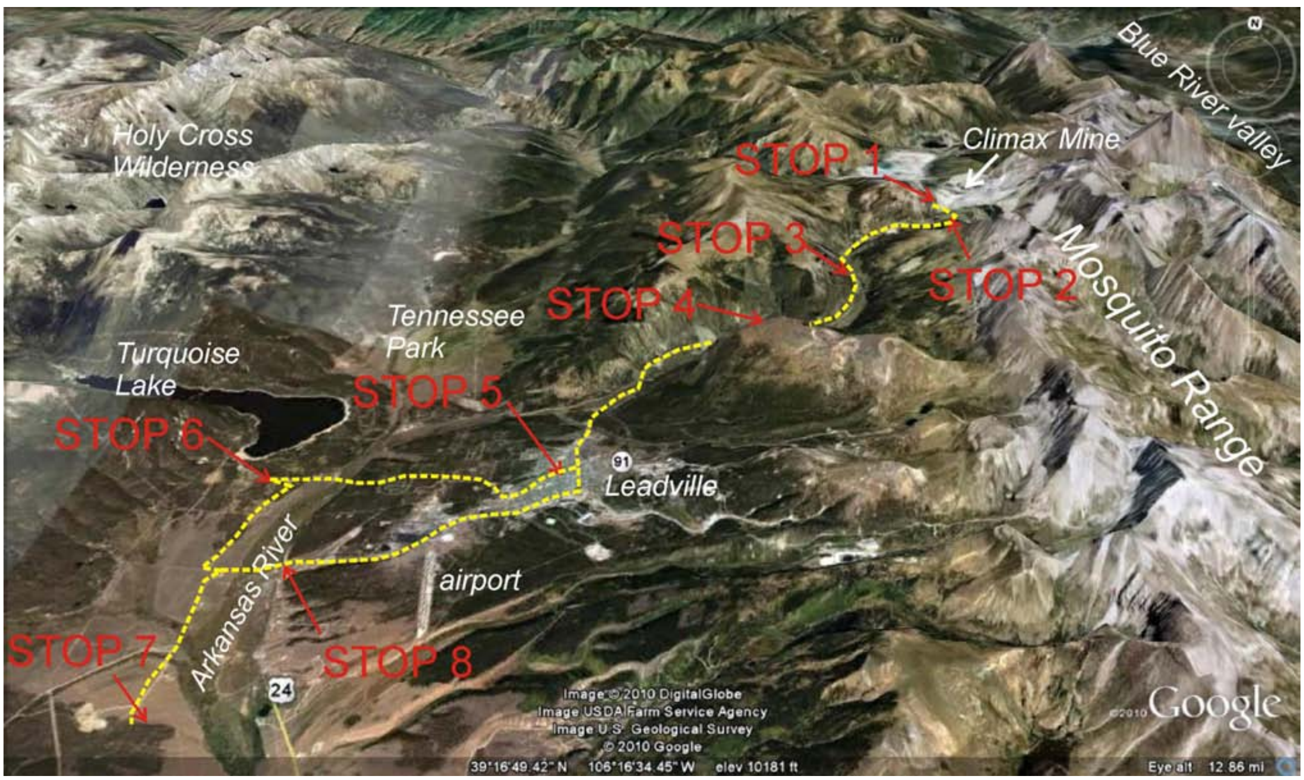


Figure 3. Oblique satellite view of DAY 1 Stops. North is to the top. The Mosquito Range (right) is the eastern flank uplift of the Rio Grande Rift. The Arkansas River (left) flows south down the rift axis.

ROAD LOG
Day 1

Mi 0.0: Exit 203 on I-70 (CO-9 to Frisco and Breckenridge)
Exit I-70 and go 270 degrees around roundabout; drive south on CO-9 to first stoplight; turn right and go into Safeway parking lot.

Mi 0.3: Safeway store, Frisco, CO [N end of CO-9]; at Signature café, coffee and restrooms. Leave Safeway and drive south on CO-9 to Main Street.

Mi 1.7: Turn right (W) onto main Street and take it all the way through downtown, to its intersection with I-70.

Mi 2.8: Drive beneath the Interstate and turn left onto ramp; continue west on I-70 to the Copper Mountain/Leadville exit (CO-91).

Mi 8.2: Exit onto CO-91 and continue south, past Copper Mountain base area, to Fremont Pass.

Mi 10.0: Pinedale lateral moraine of the Tenmile Creek paleoglacier can be seen on valley wall to left, about 400 ft above the valley floor.

Mi 11.1-11.5: Pinedale till is exposed in roadcuts to right.

Mi 12.4: CO-91 crosses Tenmile Creek and begins ascending a long grade. The grade section was built in the early 1970s.To the right old CO-91 goes west on the valley floor toward the old town of Kokomo, now buried under the tailings piles from the Climax Mine. Old CO-91 is gated and locked; all land beyond the gate is part of the Climax Mine Industrial Area owned by Freeport McMoRan.

Mi 14.5: to left, view up Mayflower Creek, a glaciated tributary in the Tenmile Range.

Mi 14.9: to left, roadcut is failing in a large landslide. For the next mile all roadcuts on left have slope stability problems, some involving major regrading during highway construction.

Mi 15.8: to left, view up Clinton Creek, a glaciated tributary in the Tenmile Range.

Mi 16.1: highway cuts through Minturn Formation (red Pennsylvanian sandstones) intruded by Tertiary dikes and sills (light tan porphyry). These Tertiary intrusives are part of the Colorado Mineral Belt and become more common toward Leadville, where they are responsible for the mineralization.

Mi 17.3-17.5: to left, failing roadcuts are in Pleistocene till with a matrix of ground-up red Minturn Formation. To right, good views of the Climax tailings piles.

Mi 19.7: STOP 1-1. (Jim McCalpin) Fremont Pass [elevation 11,318 ft]. Pull off highway to right (W) and park in front of interpretive signs. This 20-minute stop will be an overview of the pre-Quaternary geology of the Mosquito Range.

DISCUSSION ITEMS:

1—Fremont Pass was scoured out by glacial ice that overflowed from the East Fork paleoglaciers (to the south) and flowed north into the Tenmile Creek valley. We know this because the lower slopes of the pass are mantled with sporadic till and round erratics of Precambrian gneisses from the south side of the Mosquito fault, which could only have come from the cirque of the East Fork Arkansas River.

2—The Mosquito fault is a major Neogene normal fault associated with Rio Grande Rift extension. The fault has displaced the Tertiary Climax ore body (24-33 Ma) 9000 ft vertically. The fault has created a steep range front held up by the Precambrian granites and gneisses of the upthrown block; the downthrown block is more subdued forested topography underlain by the softer Minturn formation (Pennsylvanian).

From Fremont Pass, you can see the fault exposed beneath the range-front slope break, on the north wall of the Climax Pit. There the gray rocks of the Minturn Fm. are faulted down against altered Precambrian rocks (tan to orange) (see Fig. 4).

Figure 4. The Mosquito fault in the north wall of the Climax Pit, juxtaposes Minturn Formation (left) against lighter Precambrian granite and Tertiary porphyry (right). Here the fault zone dips steeply west and is approximately 50 ft wide. Widmann and others (2004a) describe the eastern margin of the fault zone as a 1 to 1.5 foot wide zone of light-gray to white, clayey fault gouge, probably derived from crushing and pulverizing of the Precambrian granite in the footwall block. The footwall block has been altered to a jasperoid mass for a distance of at least 50 ft from the fault. The wide fault zone is mainly composed of 3-6 foot wide slivers of Minturn Formation. The slivers are lens shaped, elongated parallel to the fault zone, and appear to have tapered ends that dovetail into each other. Photo by Vince Matthews.

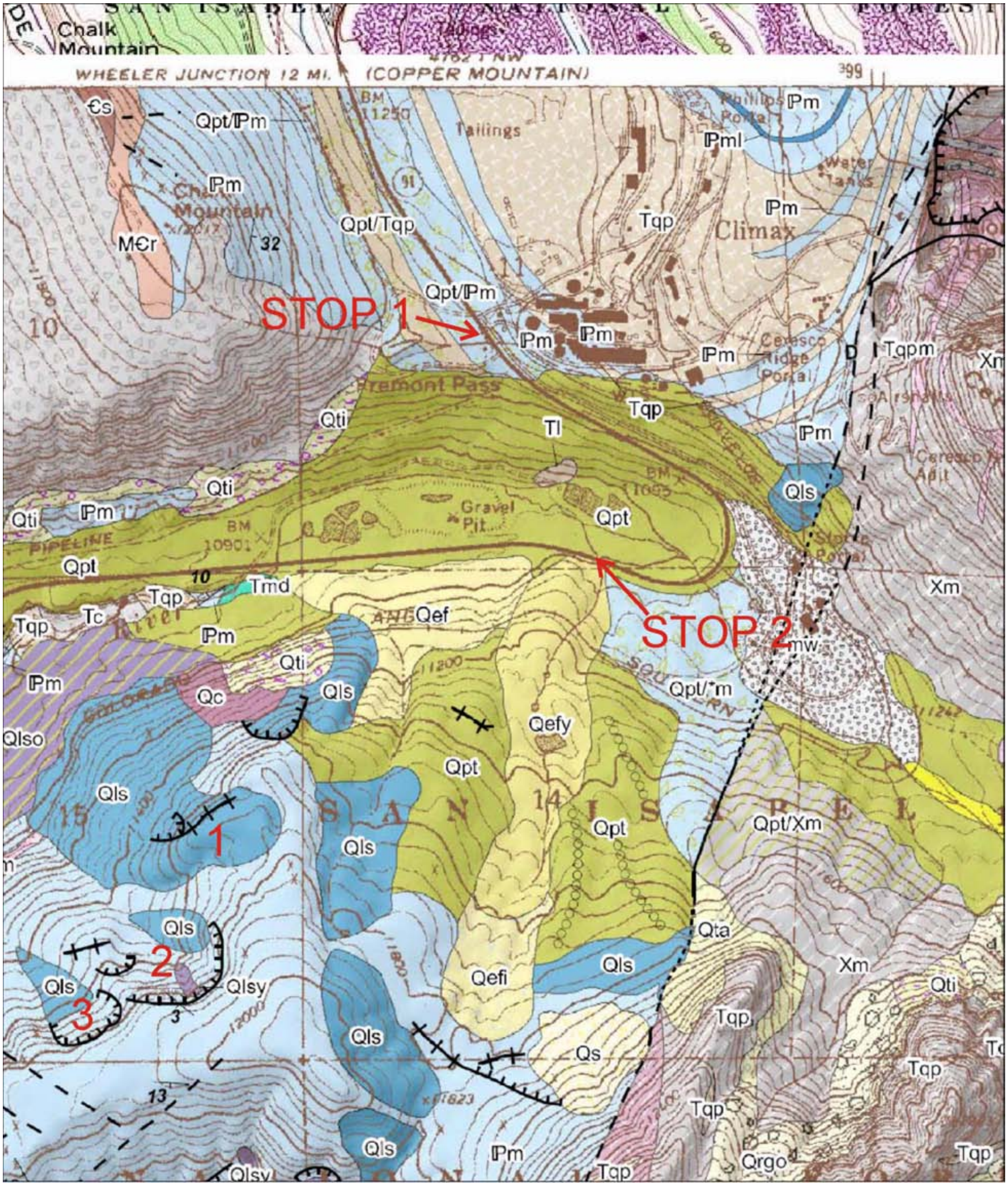


Figure 5. Map of landslides in the Minturn Fm., south of Stops 1 and 2. Earthflows (young, Qefy; intermediate, Qefi; undivided, Qef) and slumps (Qls); Fig. 7a is a photo of the upper part of Qefi; Fig. 7b shows slumps numbered 1, 2, and 3 at lower left.

From the Pit the fault crosses the glaciated valley of the East Fork (Fig. 5, lower photo), and continues along the base of the range front to the SW, below Mt. Arkansas (the high peak). McCalpin et al. (2010) concluded that there was no evidence for late Quaternary movement on this fault.

3—The north pit wall has cut away much of Bartlett Mountain. Due to the block-caving method of mining used at Climax before the open-pit phase, the base of the north wall was undermined, leading to an incipient mountain flank collapse (Fig. 5, upper photo). The headscarp of this collapse is over 100 feet high and resembles natural “sackung” landforms created by deep-seated gravitational spreading.

4—The Climax Mine (Fig. 5) was once the largest producer of molybdenum in the world. It has been in a temporary shut-down mode for the past 10 years, but will be reopened once molybdenum reserves of the nearby Henderson Mine are exhausted. This was supposed to have occurred in 2008, but has been postponed. When it was operating, the Mine was the largest employer in Lake County and provided high-wage union jobs.

5— The Minturn Formation (middle Pennsylvanian) underlies the hanging wall of the Mosquito fault here, and is prone to landsliding due to its lithology. According to McCalpin et al. (2010), the Minturn is:

Predominantly white, tan, greenish-gray, or dark purplish-gray arkosic, micaceous pebble- and cobble-conglomerate, sandstone, and shale, interbedded with dark-gray, limestone beds typically less than 30 ft thick. Black shale is most prevalent near the base of the sequence and is interlayered with thinly bedded (platy) dark-purple, gray, or buff micaceous sandstone…. Total thickness of the Minturn Formation in the quadrangle is estimated to be over 5,000 ft however, the upper part of the section and contact with the Maroon Formation is not observed.

South of Stops 1 and 2 there are two types of landslides. The landslides below the glacial limit are dominantly earthflows, and are composed of remobilized till which has a matrix of clayey, ground-up Minturn shale. The long Qefi/Qefy earthflow is fed by springs arising downslope of the Mosquito fault, possibly coming out of an unmapped parallel fault in the hanging wall.

However, the more common landslides in the Minturn are slumps and slump-earthflows. These landslides dominate above the glacial limit, where there is no blanket of till to soak up infiltration and become saturated. Some of the slumps in the Minturn east of the East Fork are dipslope failures, caused when the LGM glaciers oversteepened the valley walls. When the ice retreated there was no restraining force to hold in place weak beds that dipped outward into the steep valley wall. However, slumps also occur in the Minturn where slopes are not dipslopes.

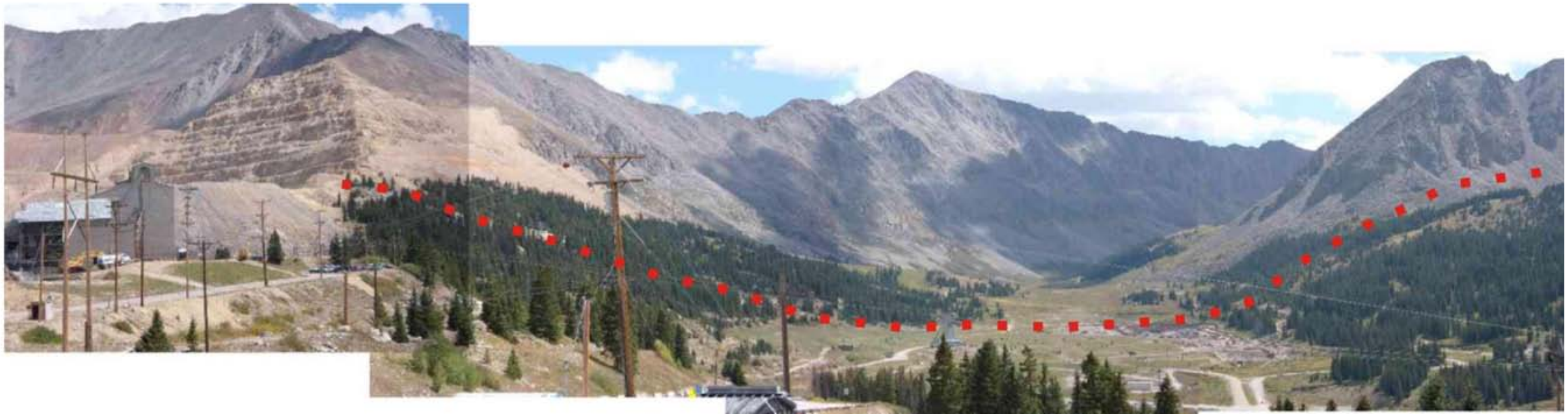
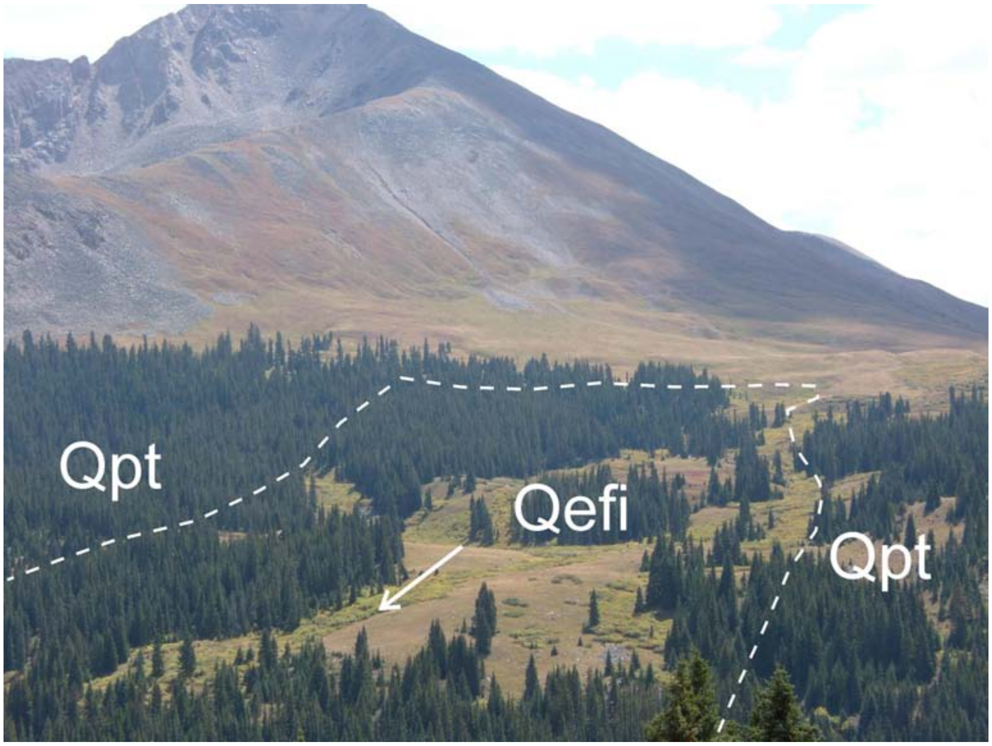
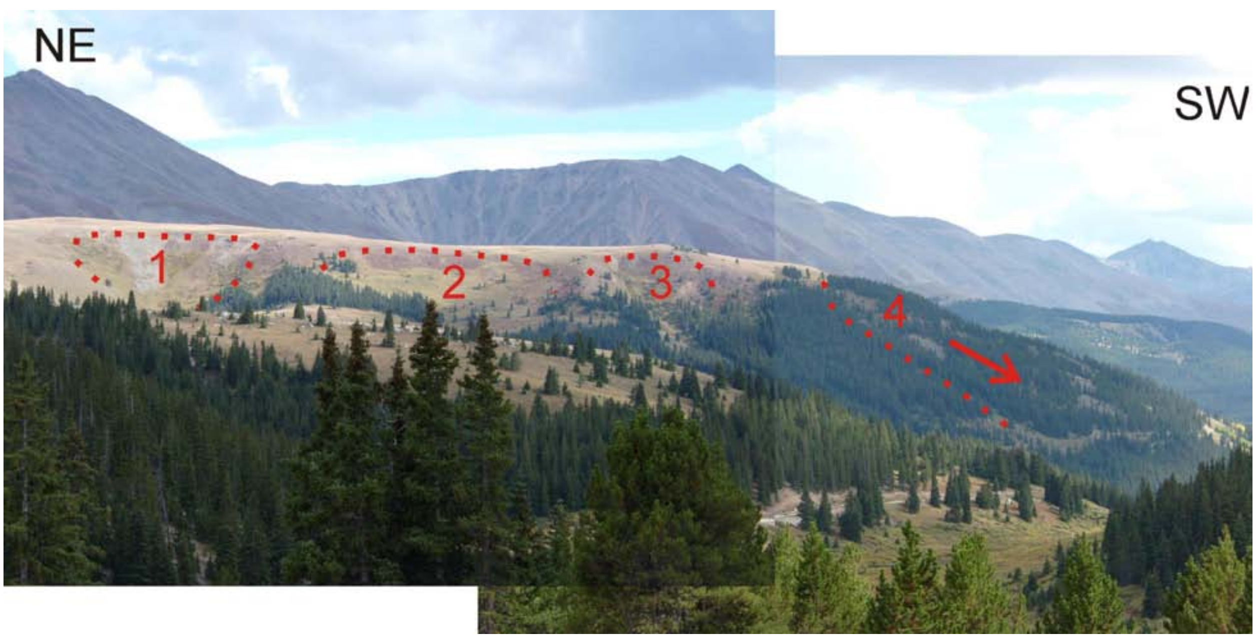


Figure 6. Panoramic photographs of the Climax Mine from Fremont Pass. UPPER PHOTO: the Climax Mine, looking east. The Mosquito fault is exposed between the red arrows at far left. The red dashed line is at the top of the gravitational collapse scarp on Bartlett Mountain. LOWER PHOTO: a continuation of the upper photo, looking to the south. The Mosquito fault (red dotted line) crosses the glaciated valley of the East Fork Arkansas River, and then continues SW at the base of the steep Precambrian range front.



(a).



(b).

Figure 7. (a) Telephoto of the head of earthflow Qefi. This earthflow is composed of remobilized till (Qpt) overlying Minturn Fm.; **(b)** photomosaic of slumps in Minturn Fm. above the glacial limit.

Mi 20.7: STOP 1-2. (Jim McCalpin) Toe of earthflow Qefy. Pull off highway to right (N) and park on narrow shoulder. This 5-minute stop will examine earthflows derived from till overlying Minturn Fm. The toe of this earthflow protrudes onto the valley floor here, forcing the highway to bend to the north. Looking back toward the Climax Mine, you can see a small slump in till, crossed by power lines above the highway.

Mi 21.9-23.3: CO-91 descends a step in the valley bottom, and the East Fork is incised into the valley floor to the left (southeast). Most of the gully walls expose a thin veneer of lodgement till underlain by Tertiary intrusives (Tqp, Tc, Tmd) and Minturn Fm.

Mi 24.0: STOP 1-3. (Jim McCalpin) Landslides occur on both valley walls of the East Fork. Pull off highway to right (W) and park on small dirt road. This 20-minute stop will discuss mega-block landslides and range-front sackung.

West-Side landslide (Qlso/Qlsi/Qlsy/Qlsw): Stop 1-3 is on the west side of CO-91 (Fig. 8), opposite the mouths of French and English Gulches. The highway cuts through the toe of a rocky landslide which protrudes 800 feet onto the valley floor (Fig. 9). The slide deposit is composed exclusively of angular blocks and smaller rubble of Minturn sandstone. The landslide deposit is composed of three lobes, with the lowest one (Qlsy) exhibiting the freshest topography. We will walk up onto the toe of Qlsy and examine the irregular topography of depressions, mounds, and trenches. This microtopography is reminiscent of that found on rock glaciers, and it is possible that it was formed post-deposition by development of interstitial ice.

Figure 8. Panoramic

However, several other lines of evidence suggest the topography is primary:
1—The elevation here (10,400 ft) is too low for a rock glacier to develop in modern climates.
2—The landslide deposit is clearly post-glacial, because it lies atop post-glacial terrace alluvium (Qat) in a formerly glaciated valley.

3—The distal part of the toe has an anomalous slope angle and texture compared to the rest of the deposit, suggesting that it was deposited in standing water. Both this landslide and one across the valley to the NE exhibit this odd toe topography, and were mapped as unit Qlsw by McCalpin et al. (2010a), described below:

Qlsw: Landslide deposits, deposited into standing water (late Pleistocene) –*Chaotically arranged debris ranging from clay to boulder size (diamicton), but dominated by large angular blocks of Minturn Formation from 1.5 to 16 ft in diameter. Derived from catastrophic rockslide avalanches that fell from oversteepened glaciated valley walls, into temporary lakes in the valley of the East Fork Arkansas River, during the Pinedale deglaciation. Surface of deposit is extremely hummocky and blocky, and source area of landsliding is easily identifiable (top of scarp area indicated by thick dashed lines with ticks in direction of sliding). Deposit is at least 16 ft thick.*

Evidence for deposition into running water:
1—Long runout across a flat valley floor.
2-- Abrupt change of slope of landslide deposit on valley floor.
3—On some toes, an abundance of clasts on surface, compared to a clast-poor surface elsewhere. We speculate that fine matrix at surface of those landslide deposits was transported away by water.

East-Side landslides (Qls of various ages, and the French-English block slide): Opposite Stop 1-3 are several isolated hills rising up on the valley floor. These hills are composed of Tertiary porphyry (Tqpm on Fig. 9) and protrude 500-600 feet onto the glacial valley floor; the porphyry is well exposed in the vertical cliff facing west. Upslope from these hills are more irregular hills of porphyry. The hills are veneered with glacial erratics of Precambrian gneiss and have a sculpted shape, indicating they were overridden by Pinedale ice.

The presence of these hills of porphyry on a glaciated valley floor is puzzling. The East Fork Arkansas River has experienced many episodes of valley glaciation in the Quaternary, as have all other high valleys in central Colorado, but very few of the other valleys display bedrock hills on their valley floors, even where the local rock type is as hard (or harder) than the porphyry (e.g., Precambrian granite or gneiss). The sculpting and erratics indicate the knobs were overridden by Pinedale ice, but it seems unlikely that the knobs could have survived being overridden by all the many Quaternary glaciations.

The topography upslope of the knobs is extremely irregular and contains linear closed depressions and scarps in bedrock, similar to those observed at the heads of landslides and in areas of deep-seated gravitational spreading. McCalpin et al. (2010a) concluded that these knobs were the toe of a bedrock landslide that had slid into the valley after the Bull Lake glaciation but before the Pinedale glaciation. This younger bedrock slide block is outlined with open squares on Fig. 9.

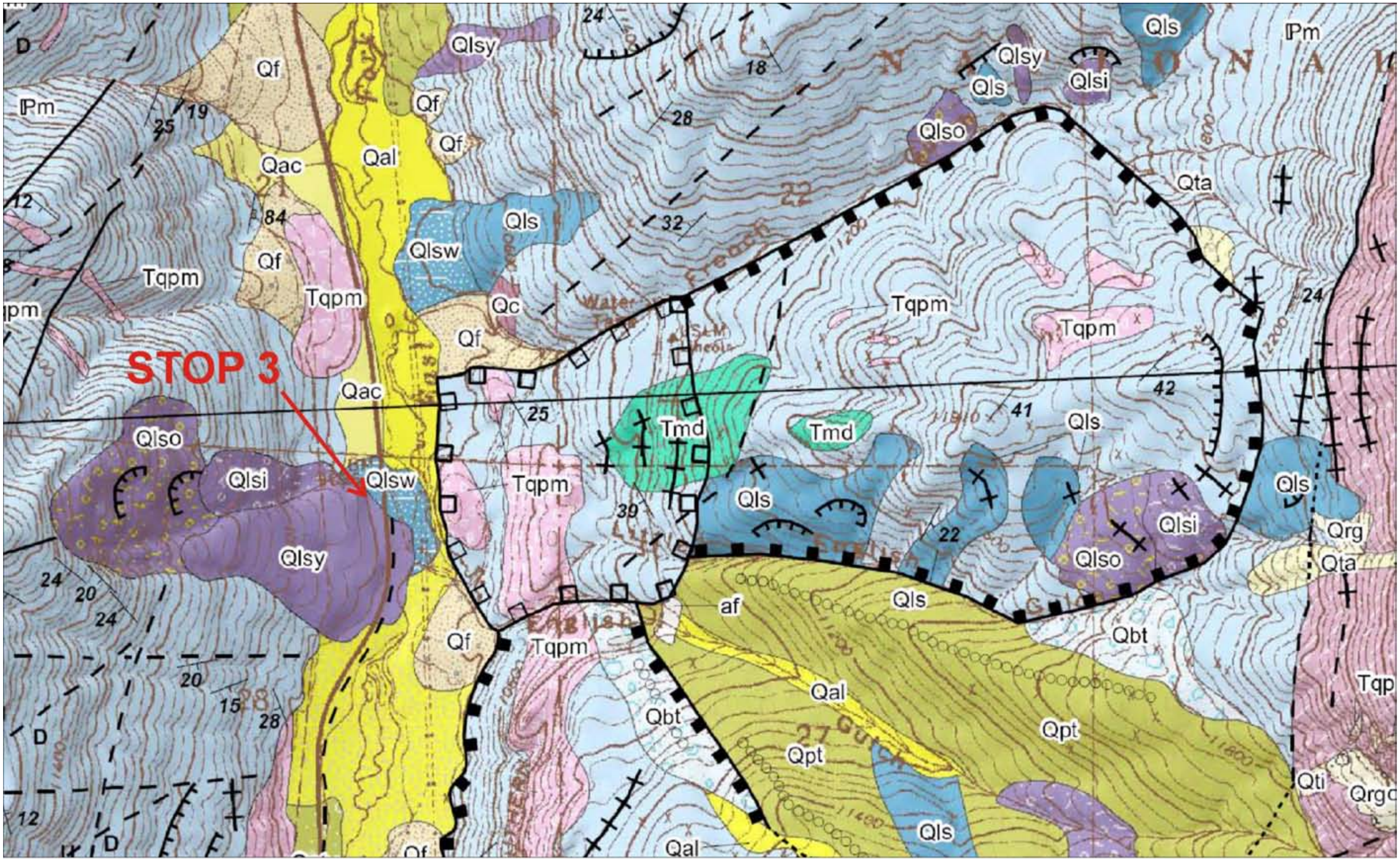


Figure 9. Geologic map of landslides in the Minturn Fm. seen at Stop 3.

The entire ridge between French and English Gulches also has anomalous topography, compared to its flanking ridges (Fig. 10). First, it is 100 ft lower than its flanking ridges, suggesting that it may have also slide valleyward. Second, the ridge is bounded by deeply-incised, linear drainages (French and English Gulches), which are atypical of the hanging wall of the Mosquito fault. Third, the only young deformation along the Mosquito fault occurs directly upslope of this ridge (Fig. 11)

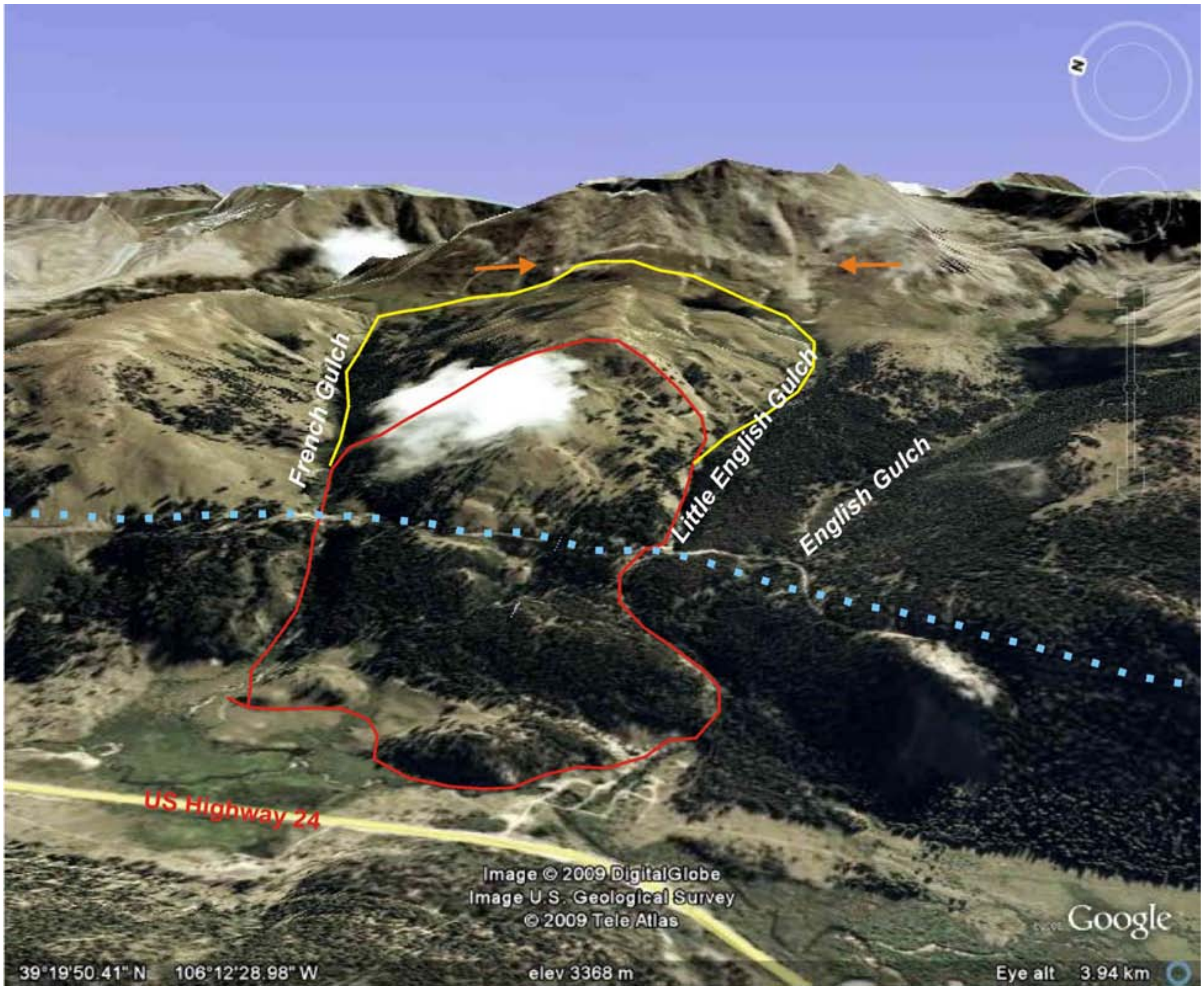


Figure 10. Annotated Google Earth view of the French-English Gulch low-angle landslide block (at center; younger part outlined in red, older part in yellow). View is to the east; the glaciated valley of the East Fork Arkansas River is in foreground. Dotted blue line shows the Pinedale glacial limit. The toe of the younger landslide block is composed of blocks of Tertiary intrusive that were clearly overridden by Pinedale ice, thus this block moved pre-Pinedale. The upper slide block extends all the way to the Mosquito fault, and based on morphology, is older than the lower landslide block. Note that the antislope scarps (orange arrows) along the Mosquito Fault escarpment exist only above the head of the low-angle slide block.

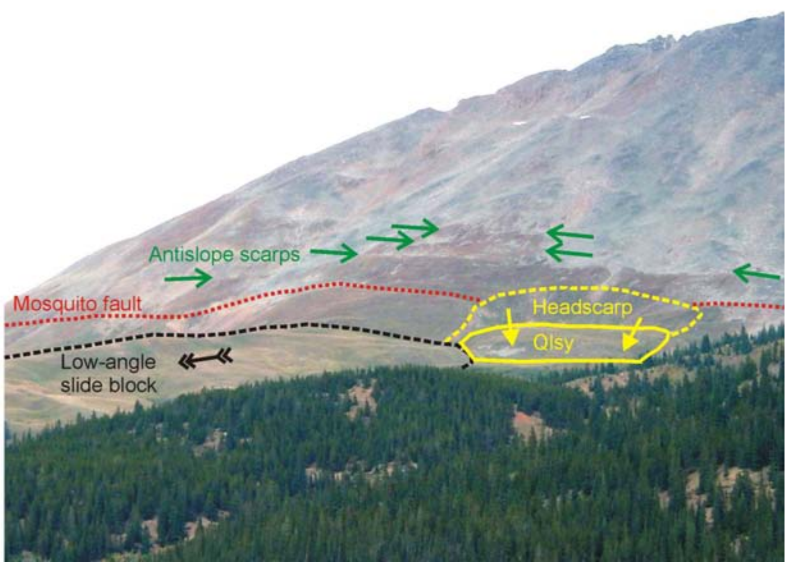


Figure 11. Unannotated (above) and annotated (below) telephoto views of antislope scarps above the Mosquito fault trace between French and English Gulches.

According to McCalpin et al. (2010a), the only scarps along the Mosquito fault in the Climaz quadrangle are related to landsliding or deep-seated gravitational spreading, rather than to coseismic surface faulting in the Quaternary. The largest anomalous landform along the Mosquito fault is the prominent bench and antislope scarp at the toe of the range front escarpment between English and French Gulches (Fig. 12). This linear landform lies roughly 50 m upslope from the trace of the Mosquito fault. The most prominent part of this antislope scarp occurs on the upslope end of the ridge that separates French Gulch and Little English Gulch (French-Little English Ridge). The trough behind that scarp exhibits a sharp fissure in active talus that may be as young as historic.

This anomalous series of three antislope scarps on the French-English Ridge is inferred to reflect gravitational spreading and westward toppling of the (altered) fault footwall rocks and overlying range-front colluvium, rather than tectonic faulting, for several reasons: (1) the antislope scarps only exist at the range front, directly upslope of the large French Gulch-English Gulch low-angle slide block, (2) there are no antislope scarps, or valley-facing scarps, elsewhere along the mapped trace of the Mosquito fault in the Climax quadrangle, and (3) the young fissure shown in Fig. 12 cannot be coseismic, because there have been no late Holocene or historic earthquakes on the Mosquito fault. The driving force for the inferred gravitational toppling is the debuttressing effect from a large, low-angle slide block directly downslope, a slide block which contains all of the French-Little English Ridge. The toe of this ridge-slide block protrudes 500-600 ft into the glaciated valley of the East Fork, but was overridden by Pinedale ice; thus it probably last had major movement between the Pinedale and Bull

Lake glaciations. However, the very fresh appearance of the trough fissure implies that stress adjustments may still be continuing.

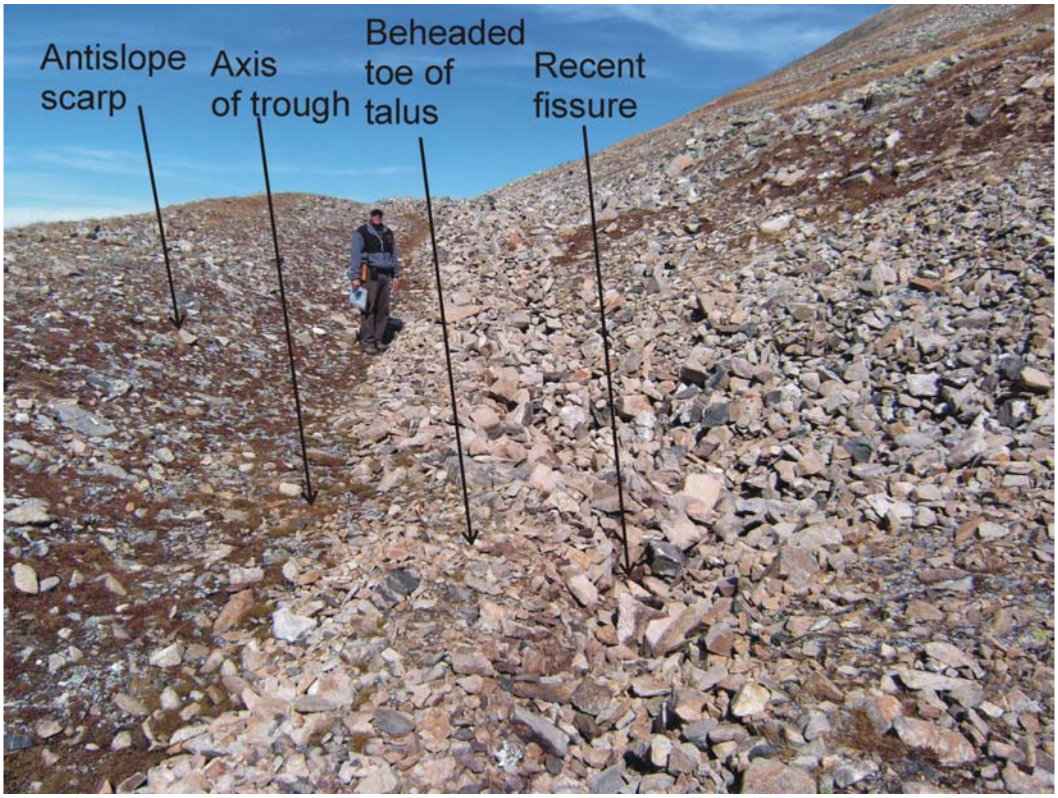


Figure 12. Photograph of the most prominent antislope scarp and trough between English and French Gulches. Active, unvegetated talus at center has been beheaded from its source at right, by the development of a fresh tension fissure. The trace of the Mosquito fault is located about 50 m downslope (to the left). Person for scale is 6 ft 2 inches tall.

Mi 25.2: recessional moraine to left; contains round Precambrian boulders

Mi 26.6: STOP 1-4. Deep-seated bedrock landsliding.

Pull off CO-91 on a short paved driveway to right, which leads to the 10th Mountain Huts Parking area (BLM).

Across the valley to the south, the north-facing valley wall is composed of a 1.1 mile-wide zone of landslides (Fig. 13). The landslides originate in Minturn Fm. that is riddled with Tertiary dikes and sills. To the east of the slide complex, dips in the Minturn are gentle and easterly in the lower part of the slope, but steeper (25-37°) and northerly in the upper part of the slope. Thus, the failures may have begun as dipslope failures. Another contributing factor could be glacial meltwater flowing down this slope from the Bull Lake paleoglaciers flowing north from Prospect Mountain; note the Bull lake moraine (Qbt) perched at the head of the slide complex.

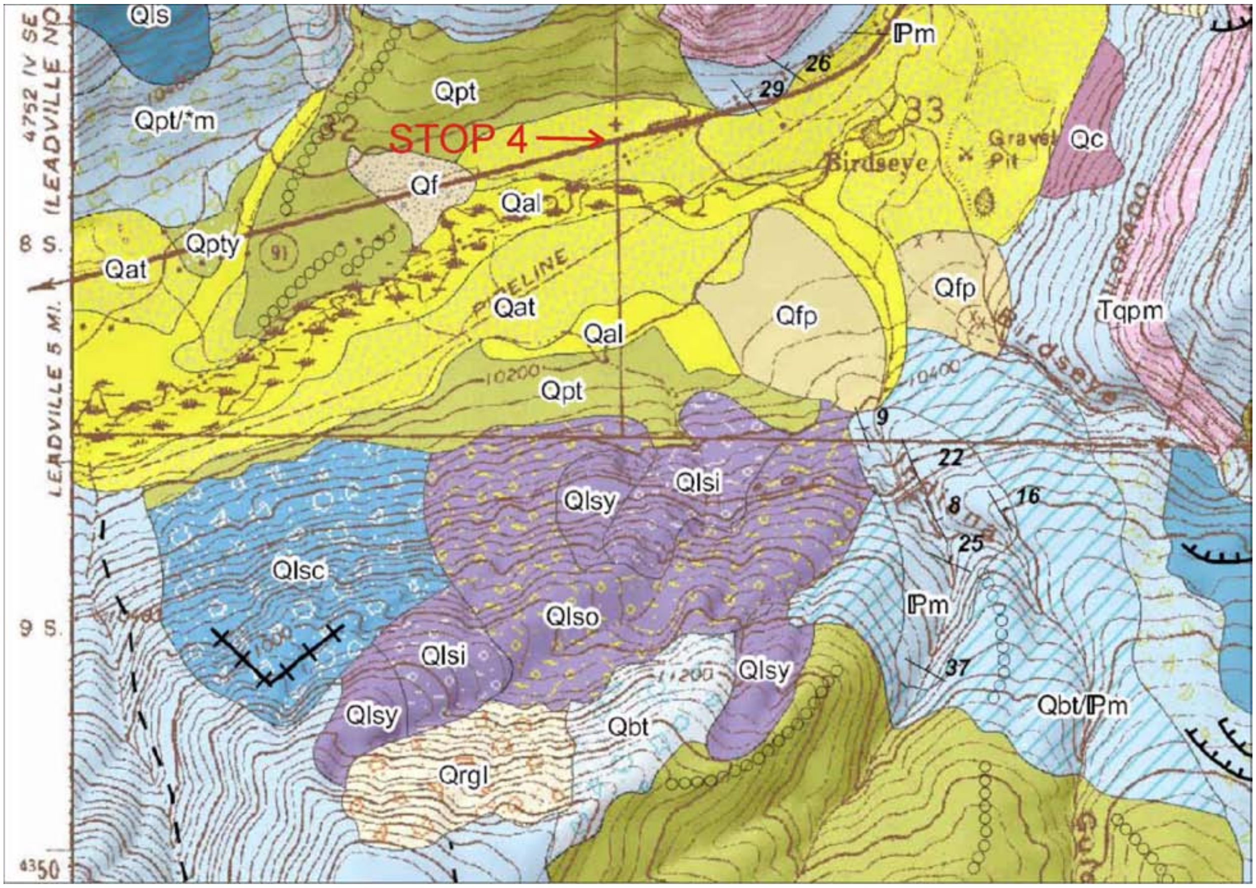


Figure 13. Geologic map of the slide complex visible from Stop 4.

The slide complex is composed of three parts. The eastern 2/3 is a complex of rock slides divided (based on morphology) into age classes Qlsy, Qlsi, and Qlso. At the head of the complex, a former rock slide has transformed into a rock glacier (Qrgl, rock glacier derived from landslide deposits). Due to the high elevation of this deposit (11,200 to 11,600 ft), it is likely it developed interstitial ice, probably during the Pinedale glaciation.

The western 1/3 of the complex is a large incipient rock failure that generally lacks the hummocky topography of the eastern part, and is not easily divisible into separate lobes. The headscarp pull-away zone is composed of two linear troughs trending at nearly right angles, with the fall line lying between them. This pattern looks like a pair of conjugate extensional fractures, across which the lower valley wall is pulling away from the upper valley wall.

This eastern 2/3 was one of the 3 landslides mapped by Tweto (1974a), probably because he was able to walk along the tracks of the Leadville and Southern Railroad which traverses across the entire complex at an elevation of 10,500 to 10,600 ft. The railroad cuts expose broken rock rubble floating in a matrix, rather than intact Minturn Fm. or porphyry. Fig. 14 shows a cut into the western 1/3 in unit Qlsc.

Figure 14. Photograph of a railroad cut into the landslide complex deposit (unit Qlsc) near the western map boundary, on the south wall of the East Fork Arkansas River. Note large intact blocks of Tertiary porphyry floating in a crushed matrix of porphyry and Minturn Formation. Height of cut shown is approximately 26 ft.

This slide complex is an example of the continuum that exists between normal landslides and deep-seated gravitational deformation (DSGD), where part of a mountain flank has begun to detach and topple away from the rest of the mountain mass. In the latter case the toppling mass may be completely intact rock and mapped as such by bedrock mappers. The only hint of the toppling will be anomalous extensional landforms in the pull-apart zone at the head of the slope; these may be downhill-facing scarps, uphill-facing scarps, linear troughs, or closed depressions. The key to recognizing such landforms as evidence of DSGD is that they could not have been formed by in their locations by normal hillslope/fluvial erosion processes.

Return to CO-91 and continue driving west toward Leadville. For the next 3 miles the floor of the East Fork valley is dominated by a wide, nearly horizontal floodplain, scarcely incised and invaded by willows. This valley floor may have been the site of a post-glacial lake, although the key evidence for that would be downstream in the Leadville North quadrangle (last mapped by Tweto in the 1970s, as part of the Holy Cross 15’ quadrangle).

Mi. 29.7: CO-91 crosses the East Fork. For the next 2 miles the highway ascends through the Pinedale lateral moraines, and merges with US-24 north of town. The highway finally tops out in Bull Lake moraines where commercial buildings appear along the highway.

Mi 32: enter Leadville; as it approaches downtown, US-24 bends to the right (west) and then to the left (south) at Harrison Street. Instead of turning left onto Harrison (the main drag of Leadville), continue straight west on 9th Street across Harrison Street, and up to the National Mining hall of Fame and Museum on the right. Park there.

Mi 32.3: STOP 1-5. National Mining Hall of Fame and Museum, 120 W. 9th Street. Overview of Quaternary geology of Leadville.

The Quaternary geology of Leadville (Fig. 15) could be called “a dog’s breakfast”, for three reasons. First, the city lies on a piedmont of old (Tertiary?) basin fill that has been veneered and channeled by glacial outwash coming from Evans Gulch to the east. Second, these deposits have been so modified by mining and urbanization that the original landforms have been obscured. Third, the most detailed geological mapping was performed more than 100 years ago when the mines were still active, but scant attention was paid back then to Quaternary deposits, compared to the economic bedrock units. Emmons divided the Neogene deposits into two units, a younger “Wash” (which probably includes all of the Quaternary), and an older “Lake Beds” (probably includes all the late Tertiary), which were exposed only in mine shafts and tunnels (Fig. 16). Even as late as the 1970s Tweto (1974c) surprisingly ignored the Quaternary cover deposits on the Leadville piedmont and mapped the underlying bedrock units, which are not actually exposed (modifying his contacts from Emmons, 1886 and Emmons et al., 1927). In 1984 Nelson and Shroba prepared a reconnaissance Quaternary map for an AMQUA field trip, but based on little field checking. In 2008 McCalpin et al. mapped the Leadville South quadrangle for Colorado Geological Survey, but did not map the Leadville North quadrangle. This awkward legacy is portrayed in Fig. 15, where the city is shown as lying mainly on older Bull Lake outwash from Evans Gulch (lower panel, unit Qboo), with the southwestern part of town lying on a slightly lower terrace (younger Bull Lake outwash, unit Qboy).

The most enigmatic Quaternary feature is Capitol Ridge, on which Stop 5 is located. This 2.3 mile-long ridge rises about 150 feet above the flanking outwash surfaces, and was been mapped by Capps (1909), Nelson and Shroba (1984), and McCalpin et al. (2010b) as a moraine (the latter as a younger pre-Bull Lake moraine).

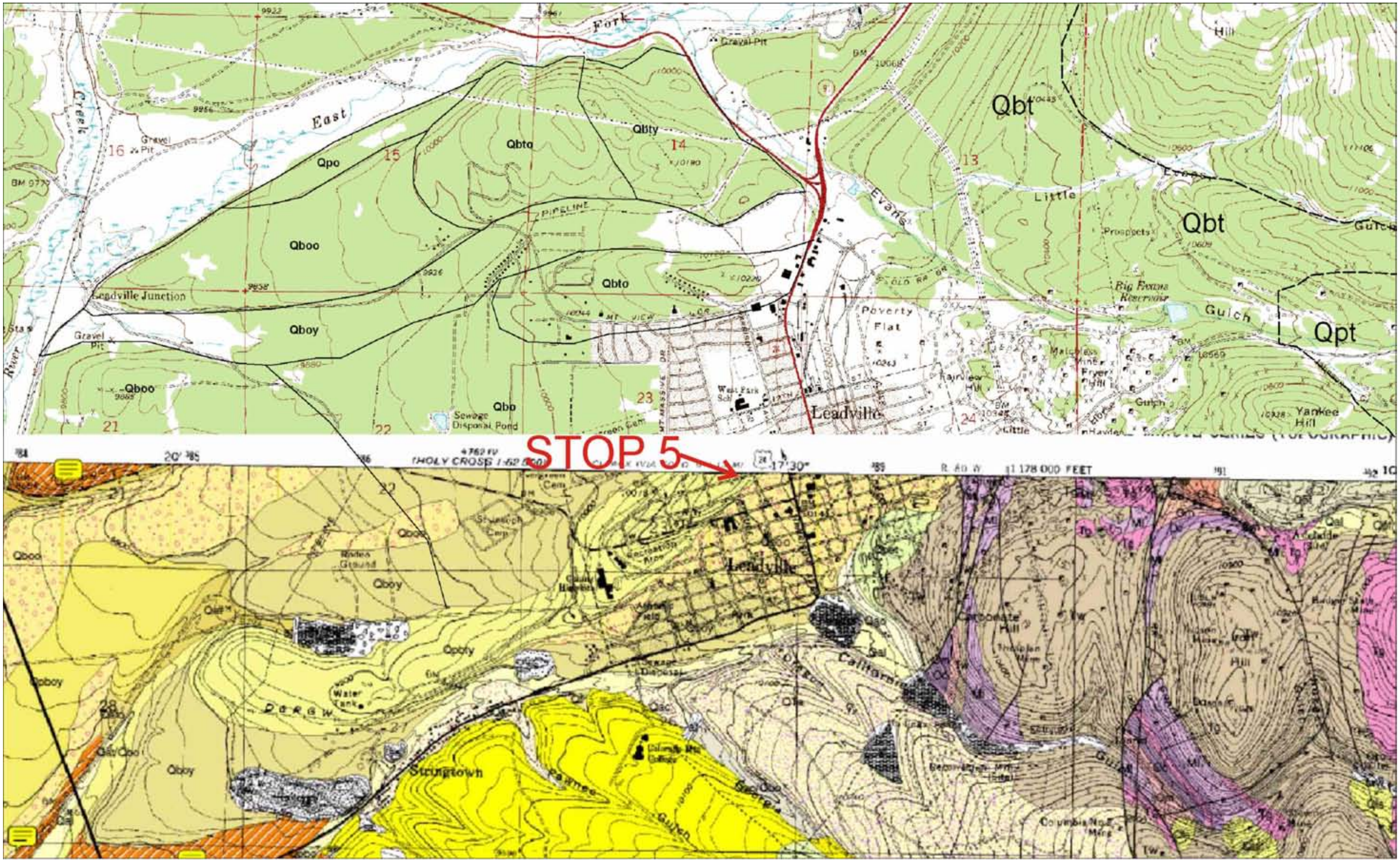


Figure 15. Composite geologic map of Leadville South quad (below, color) and Leadville North quad (above, contacts from Nelson and Shroba, 1984).

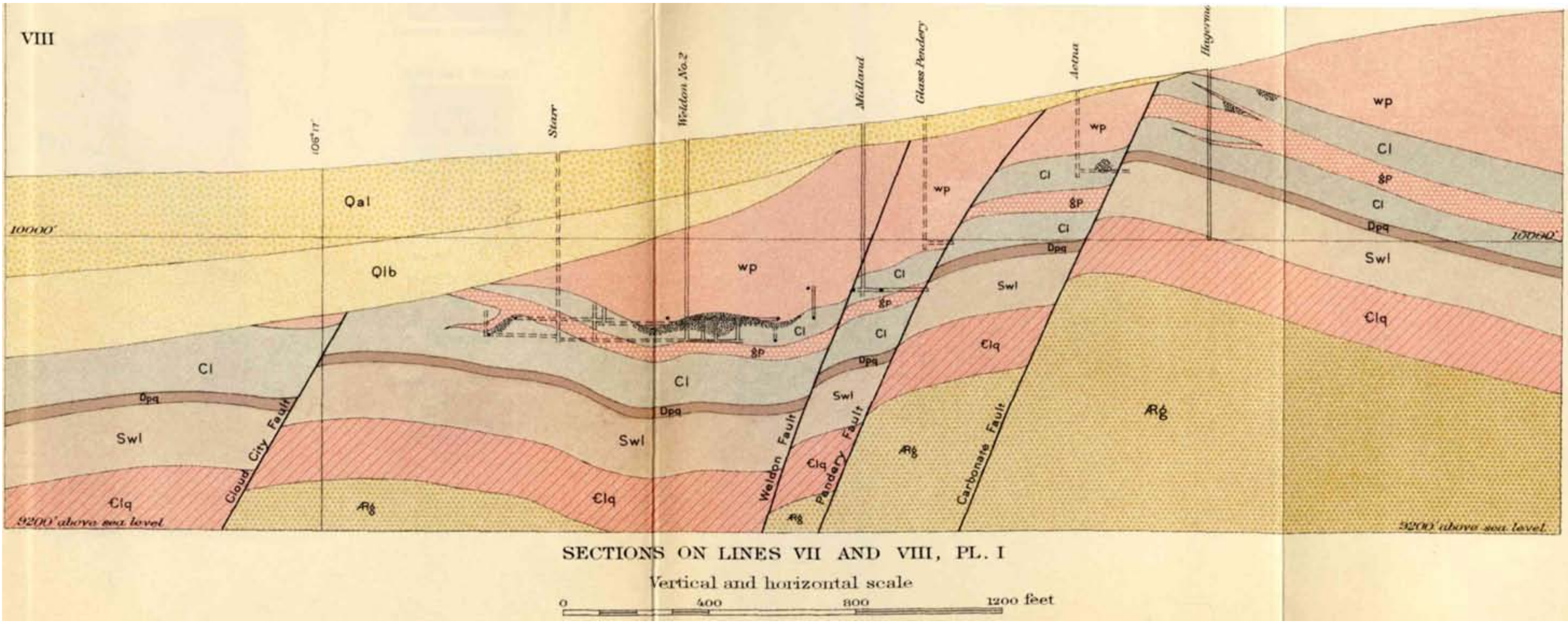


Figure 16. Cross-section across the Late Cenozoic normal step-faults at Carbonate Hill, just east of downtown Leadville, from Emmons and Irving, 1907 (from left to right, Cloud City fault, Weldon fault, Pendery fault, Carbonate fault). The latter three faults merge into a single Pendery fault just south of this section line. Although not shown in this section, the Pendery fault displaces the Late Cenozoic "Lake Beds" (lower part of the Dry Union Formation?) farther to the north (Section IV of Emmons and Irving, 1907). Qal, the "Wash" (our units Qpbo, Qboo, Qboy); Qlb, the "Lake Beds" (nowhere exposed at the surface); wp, White Porphyry (our unit Tw); Cl, Carboniferous limestone (our unit Ml, Leadville Dolostone); gp, Gray Porphyry (our unit Tg); Dpq, Parting Quartzite (our unit Dc); Swl, White Limestone (Manitou Dolomite of modern usage, Om); Clq, lower quartzite (Sawatch Quartzite of modern usage, Cs); ARg, Archean granite (our unit YXg). From McCalpin et al., 2010b.

However, several uncertainties arise if Capitol Ridge is a moraine. First, if it is the south lateral moraine from Evans Gulch, where is the corresponding north lateral moraine? Could it be the ridge behind the Safeway Store in north Leadville that was mapped by nelson and Shroba (1984) as older Bull Lake (Qbto)? But those two ridges come so close at their upper ends, that there would not be much room for a glacier terminus between them.

Second, Capitol Ridge does not expose many boulders. This could be partly the result of urbanization, or partly the result of long weathering since pre-Bull lake time. For example, the pre-Bull Lake till mapped on the Airport pediment south of Leadville has very few boulders at the surface, and is deeply weathered.

Third, the ridge is about the same elevation as the pre-Bull Lake pediment surface south of downtown Leadville (the Airport surface). One alternative origin is that capitol Ridge is merely an erosional remnant of the same pre-Bull Lake pediment that exists south of the city. However, the trend and shape of the ridge is discordant with that pediment.

Continue driving west on W. 9th Street; turn left (south) on Spruce Street and drive 1 block to W. 8th Street; turn right and continue west on W. 8th St.. It curves to the south, becomes Washington Street, and intersects W. 6th Street.

Mi 32.9: junction of Washington and W 6th Streets; turn right (W) and drive west on W. 6th Street past the school and pool.

Mi 42.2: junction of W. 6th St. and CR-4; continue west on CR-4 across the younger Bull Lake outwash surface.

Mi 44.3: road descends down to low terraces and floodplain of the Arkansas River. Continue west across valley floor.

Mi 44.7: bear right at forks, staying on CR-4, and ascend onto the terminal moraine of Turquoise Lake. Continue west across Sugarloaf Dam.

Mi 46.0: West abutment of Sugarloaf Dam; park here; **STOP 1-6**.

Retrace route back to east; at forks (Mi 47.3) turn right (south) onto CR-5a and proceed 2.0 miles south.

Mi 48.0: junction of CR-5a and CO-300; turn right (west) onto CO-300

Mi 48.3: junction of CO-300 and CR-11; turn left (south) onto CR-11 and proceed south

Mi 49.4: CR-11 crosses the Lake Fork of the Arkansas River

Mi 49.7: CR-11 crosses Halfmoon Creek. At the junction with Halfmoon Road continue on CR-11A.

Mi 50.2: turn off onto 2-track road that forks to left (SE) and park; **STOP 1-7.** Halfmoon Creek, its moraines and outwash terraces.

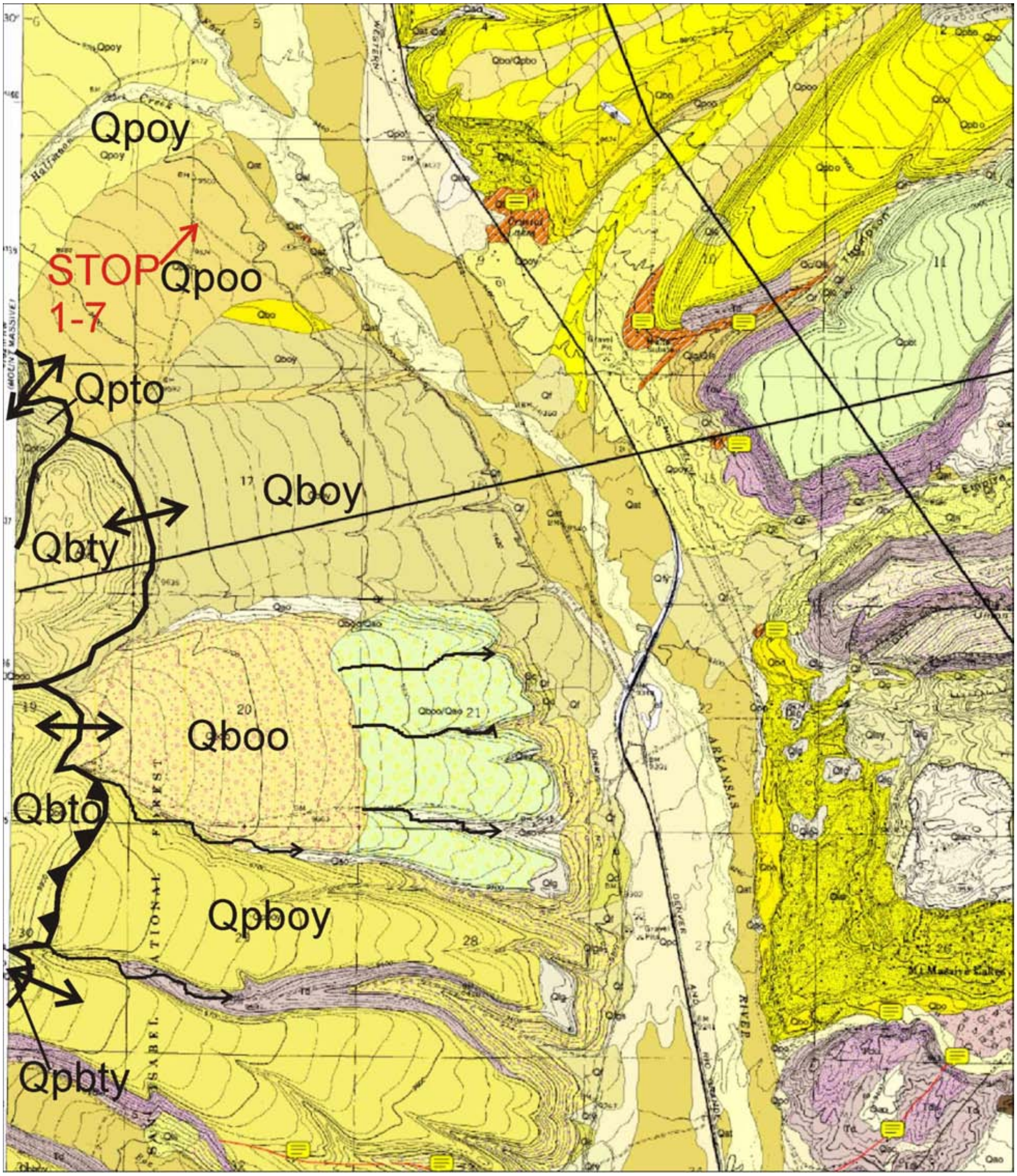


Figure 17. Geologic map south of Stop 1-7, showing the eastern tips of moraines of Halfmoon Creek, and their associated outwash surfaces.

Halfmoon Creek, its moraines and outwash terraces

Halfmoon Creek flows east out of the crystalline core of the Sawatch Range, but then makes a right-angle turn to the north when it crosses the range front. It then flows 4 miles northward before emptying into the Arkansas River. This 4-mile reach cannot be seen from the floor of the Arkansas Valley, because it is hidden behind the heavily-forested, east lateral moraine ridges of the Halfmoon paleoglacier.

A small part of these east lateral moraines lies within the Leadville South quadrangle, so McCalpin et al. (2010b) attempted to subdivide the moraines by age and relate them to the outwash surfaces that form a 2 mile-wide piedmont slope west of the Arkansas River. Based mainly on moraine surface morphology and boulder frequency, they subdivided the moraines into lobes of different inferred ages, from north to south, older Pinedale (Qpto), younger Bull Lake (Qbty), older Bull Lake (Qbto), and pre-Bull Lake (Qpbty) (Figs. 17, 18). In most cases the moraines appeared to grade smoothly onto the heads of the outwash surfaces, and were thus deemed to be contemporaneous (shown by double-headed arrows in Fig. 17). However, in one case it appeared that a younger moraine (Qbto) had advanced out over the head of an older, pre-existing outwash surface (Qpboy); this relationship is shown by a thrust symbol.

It also appears that the piedmont surface onto which the outwash was shed as a thin veneer, already contained incised tributary gullies to the axial Arkansas River. In that case, meltwater spread a veneer of outwash gravel over the western (unincised) part of the piedmont slope, but eastward the meltwater flowed off the piedmont surface and down into pre-existing gullies. This occurred on both the Qboo and Qpboy surfaces.

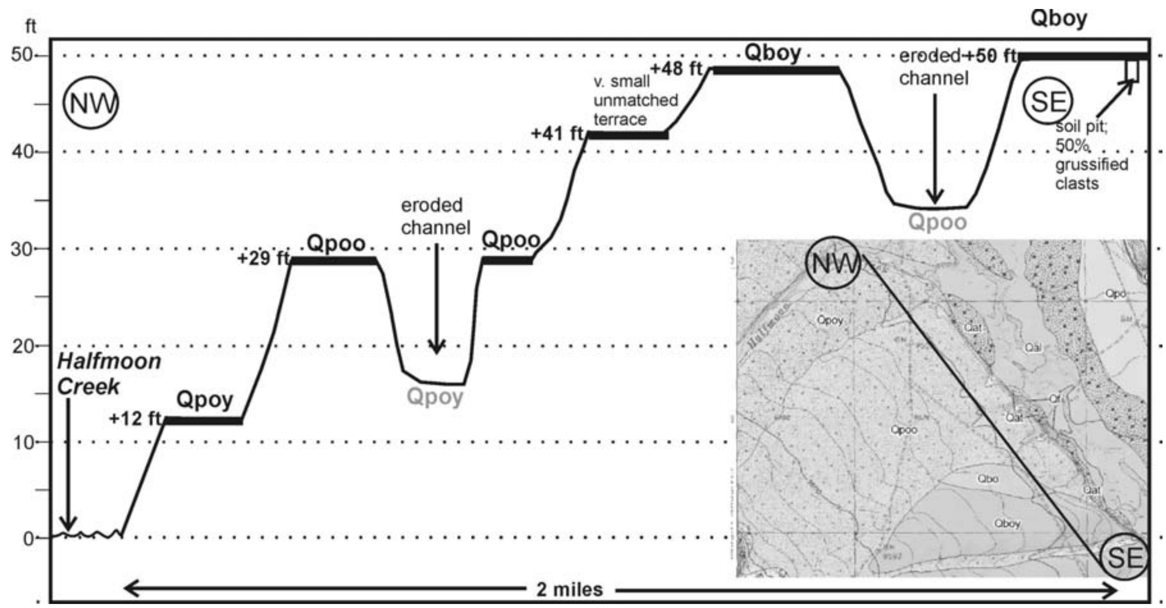


Figure 18. North-south topographic profile across the sequence of younger glacial outwash terraces of Halfmoon Creek. Qpoy, younger Pinedale outwash; Qpoo, older Pinedale outwash; Qboy, younger Bull Lake outwash. Gray labels indicate erosional channels that may have little or no backfill. All terrace heights (e.g., +12 ftm) are measured from the bed of Halfmoon Creek, except for Qboy at far right, which is measured above the Arkansas River.

Retrace route back north on CR-11.

Mi 52.4: junction of CR-11 and CO-300; turn right (east)

Mi 53.2: junction of CO-300 and US-24; turn left (north)

Mi 53.3: turn left into the entrance of Saturday's Market; proceed straight north past the gas pumps on a gravel road (CR-36) until it turns sharply to the right; continue straight ahead to the old railroad grade. If gate is unlocked, proceed north to the base of the exposure.

Mi 53.7: STOP 1-8 (optional). (Jim McCalpin). Type locality of the Malta Gravel of Tweto (1961).

The abandoned townsite of Malta lies at the mouth of California Gulch where it empties into the Arkansas River. There was a large railroad switching yard here in the mining days, and to widen the switching area, workers cut a long excavation (Fig. 19) into the toe of the pediment surface to the east (mapped by McCalpin et al., 2010b, as capped by pre-Bull Lake outwash, unit Qpboy).

Figure 19. Photograph of the type locality of the "Malta gravel" of Tweto (1961) (our map unit QTa), in an old cut at the Malta railroad siding. [UTM Z13, NAD27, 383530m E, 4342580m N].

Tweto (1961) applied this name to all of the gravel underlying the high pediments flanking the Arkansas River, so it would include McCalpin et al.'s (2010b) Qboo and Qpboy surfaces of Halfmoon Creek (previous stop), and the Qboy, Qboo, Qpbo, Qpboy, and QTa surfaces east of the Arkansas River.

The age of the gravel exposed at the type locality is uncertain. The gravels that cap the high surface to the east are mapped as pre-Bull Lake outwash (middle Pleistocene). However, the gravels in the cut lie 80 feet beneath the Qpboy surface, and could be early Pleistocene, or even late Pliocene. McCalpin et al. suggested the latter by placing this small outcrop in their unit QTa, Quaternary-Tertiary alluvium.

Close examination of the exposure reveals several clast-free, sandy beds that might be amenable to luminescence dating. If the beds yielded middle-Pleistocene ages, that would tell us something about the sedimentation rate in the graben.

Retrace route back to US-24.

Mi 54.1: junction of CR-36 and US-24; turn left (east) and proceed to Leadville. Drive through downtown on US-24, which turns to Polar street in north Leadville.

Mi 59.1: pull into the Silver King Inn, 2020 N. Poplar Street.

ROAD LOG
Day 2
Reset trip meters to zero!!

Day 2: INTRODUCTION

Much of the research presented here builds on a 1984 AMQUA-sponsored field trip through the same region (Nelson et al., 1984) and thus we will visit several of the same locations of interest. However, unlike the AMQUA excursion that focused primarily on the relative ages of Quaternary deposits, this trip will benefit from newly emerging, numerical age control on the same deposits.

Upper Arkansas River valley glacial geomorphology and previous work

The Upper Arkansas valley encompasses the headwaters for the Arkansas River, which drains the Mosquito Range situated to the east and the Sawatch Range to the west (Fig. 20). The first detailed characterization of the region was conducted the late 1800s (Hayden, 1874), and subsequent descriptions by Westgate (1905) and Capps (1909) established Arkansas River valley as ‘a classic area of alpine glacial deposits’ (Nelson et al., 1984). Since these early descriptions, detailed geologic mapping has been conducted by Tweto (1961, 1979), Tweto and Case (1972), Tweto and Reed (1973), Tweto et al. (1976, 1978), Keller et al. (2004), McCalpin and Shannon (2005), and McCalpin et al. (2010).

Bedrock in the region is composed primarily of Precambrian plutonic and metamorphic rocks (granites and gneisses), with exposures of late Cretaceous and Tertiary plutonic rocks (Tweto et al., 1976; 1978; Scott et al., 1978). During the latest Pleistocene glaciation (e.g. Pinedale; Marine Isotope Stage, MIS 2), Sawatch Range glaciers extended eastward towards the north-south trending upper Arkansas River valley. Notably, glaciers that flowed down Clear Creek, Pine Creek and Lake Creek valleys extended beyond the range front and onto the Upper Arkansas valley floor, depositing large latero-frontal moraines. These valleys are all east facing, and at their Pinedale maximum extent glaciers emanating from Clear and Pine Creek valleys dammed the Arkansas River (Scott, 1984; Nelson and Shroba, 1998; Lee, 2010). Clear Creek and Pine Creek valleys each host a pair of sharp-and single-crested lateral moraines that are >150 m high. These moraines are easily distinguishable from more extensive Bull Lake (MIS 6) moraines that are characterized by a more subdued surface topography and higher degrees of weathering on boulder surfaces (Nelson and Shroba, 1998). Unlike Clear Creek and Pine Creek valleys, where single Pinedale moraines are preserved, the Lake Creek valley contains numerous Pinedale end moraines (Nelson and Shroba, 1984). Glacier retreat from the terminal moraines in the Clear Creek and Pine Creek valleys likely led to a series of outburst floods that deposited two boulder-rich terraces downstream of Pine Creek, resting ~ 15 and 6 m above the modern day Arkansas River (Scott, 1984; Lee, 2010).

Initial work in the area identified four flood terraces along the Arkansas River south of Pine Creek (Scott, 1975, 1984). Terraces were tentatively correlated to glacier maxima and attendant outburst floods occurring during the early Pleistocene (ca. 1.4

Ma), middle Pleistocene (ca. 0.6 Ma), and late Pleistocene (two lowermost terraces; Scott, 1984).

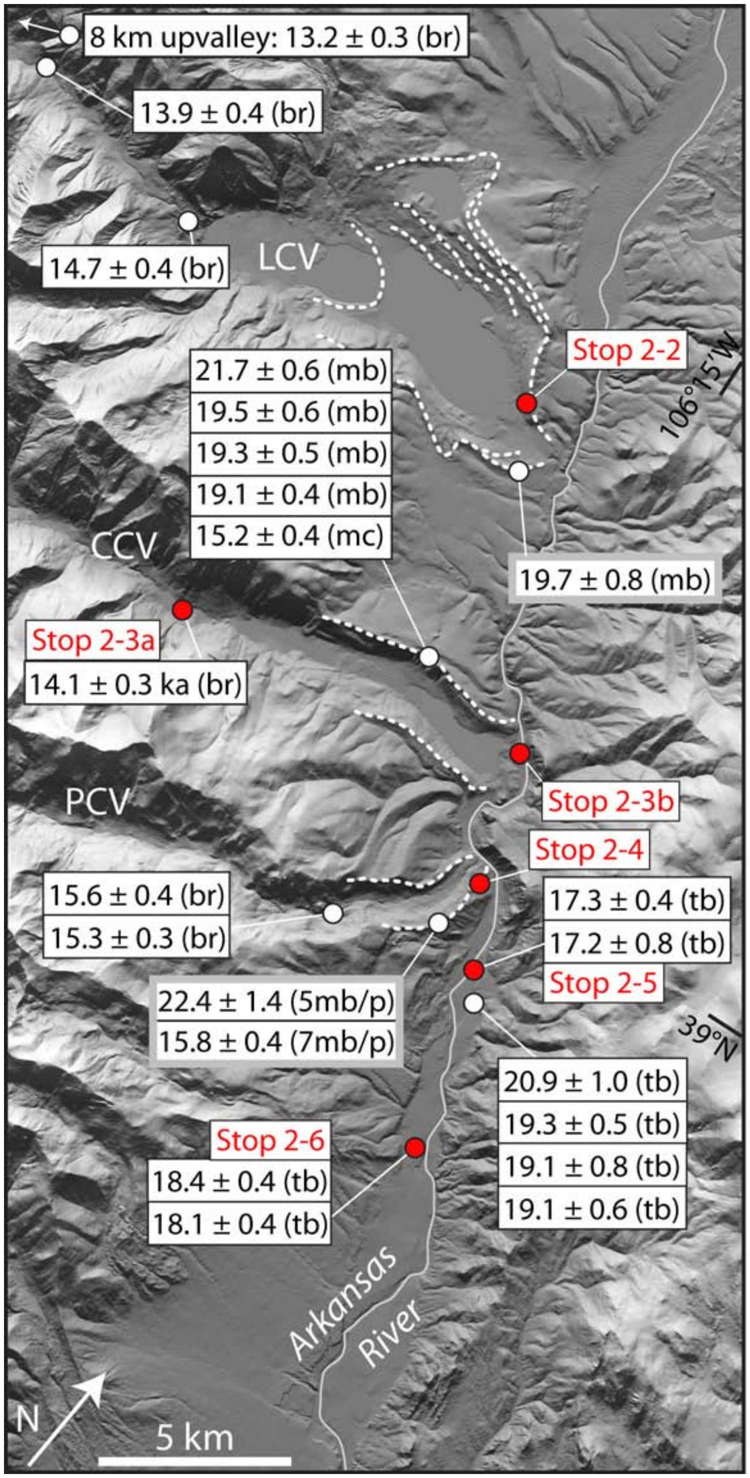


Figure 20. Upper Arkansas River study area showing ^{10}Be ages (ka) with 1σ uncertainty; gray boxes are ^{10}Be ages from previous studies (br=bedrock; mb=moraine boulder; mc=moraine clast; mb/p=moraine boulders and pebbles; tb=terrace boulder); LCV= Lake Creek valley; CCV=Clear Creek valley; PCV=Pine Creek valley.

The age of the uppermost terrace with flood boulders is constrained by correlation to a surface ~20 km away containing Bishop Ash (ca. 760 ka; Sarna-Wojcicki et al., 2000). Tributary alluvial fans sourced from the Mosquito Range, containing Lava Creek B ash (ca. 640 ka; Lanphere et al., 2002), are correlated with the next highest flood deposit (Scott, 1984). The two lower terraces lie within Pinedale-age outwash and are strewn with large boulders, almost all of which are composed of granodiorite (Scott, 1975); smaller clasts are composed of both granodiorite and metamorphic lithologies. These boulders were likely dislodged from moraines and valley walls during flooding. On this trip, we focus on the lower two terraces of the four (i.e., the late Pleistocene terraces), which we refer to as simply as the 'upper' and 'lower' terraces.

Cosmogenic nuclide exposure dating

Cosmogenic nuclide exposure dating (Gosse and Phillips, 2001; herein surface exposure dating) has emerged as the preeminent tool to directly date surficial deposits. Briefly, cosmic rays entering the earth's atmosphere collide with nuclei triggering a cascade of high-energy neutrons. These neutrons eventually reach the earth's surface and collide with target nuclei hosted within certain minerals, resulting in the splitting of target nuclei (spallation) and the creation of new *in situ* nuclides. The concentration of these *in situ*' cosmogenic' nuclides within the uppermost surface of the crust (~3 m) is directly related to how long that surface has been exposed to cosmic ray bombardment. Thus, the longer a surface has been exposed to cosmic rays, the greater the number of nuclides that surface will contain. By measuring the total abundance of a particular *in situ* cosmogenic nuclide in a rock surface, combined with knowing the rate at which the nuclide is produced, it is possible to determine the time elapsed since initial rock surface exposure. Although several *in situ* cosmogenic nuclides are produced within the earth's crust (i.e. ^{3}He, ^{10}Be, ^{14}C, ^{21}Ne, ^{26}Al, ^{36}Cl), ^{10}Be is the most commonly measured nuclide because: 1) ^{10}Be is primarily produced in quartz and quartz-bearing lithologies are relatively widespread, 2) isolation of ^{10}Be from quartz is relatively simple, and 3) the production of ^{10}Be in quartz is straightforward and the production rate is reasonably well constrained. All ages from the Arkansas River valley are ^{10}Be surface exposure ages. Ages are presented in ka at 1-sigma confidence.

Determining the surfaces exposure ages of moraines is one of the most common applications of the method. However, moraines degrade over time and boulders that were once resting directly on the moraine crest can be transported downslope. In addition, boulders that were initially covered by till can be exhumed. Both of these processes lead to surface exposure ages that are younger than the actual age of moraine deposition. As a result, several surface exposure ages are needed from any given moraine crest (typically 5 or more), and preference is often given to the oldest surface exposure ages within a dataset as these ages mostly likely record the 'true' timing of moraine deposition. Within the Arkansas River valley, we obtained five ^{10}Be surface exposure ages from the Clear Creek terminal Pinedale moraine and twelve ^{10}Be surface exposure ages from the Pine Creek terminal Pinedale moraine (Briner, 2009).

We also obtained ^{10}Be surface exposure ages from glacially polished bedrock surfaces located in the valley bottom upvalley from Pinedale terminal moraines in the Lake Creek, Clear Creek, and Pine Creek drainages. Unlike moraines, bedrock surfaces do not experience mass degradational processes that can result in erroneously young surface exposure ages. Thus, in certain settings, thoughtfully sampled bedrock yields surface exposure ages that accurately constrain the timing of deglaciation.

Depart Silver King Inn **(Mi 0.0 for Day 2)** and drive south on US-24 through Leadville, and on to Malta (Stop 1-8 of previous day)

Mi 5.0: junction of US-24 and CR-36 at Malta; continue south on US-24. From here south, cuts behind houses to the left (east_ expose poorly-sorted, oxidized sand and gravel of the Malta Gravel.

Mi 6.2: US-24 ascends a slope to rise up onto the Pinedale fan of Iowa Gulch. This slope is too steep to be a constructional part of the fan, and it is unlikely that the Arkansas River cut the slope by erosion. The slope might be the subaqueous slope of a fan-delta deposited into Three Glaciers Lake, but its elevation is too high compared to the supposed Pinedale ice dam at 9360 ft.

Mi 8.6: turn left (east) off of US-24 at a large bend onto CR-7. Proceed south. The dirt road on the left (east) goes up Empire Gulch to the Beaver Lakes subdivision, which is built on a huge landslide complex.

Mi 9.2; CR-7 curves to the left and begins ascending onto the Mt. Massive Lakes landslide.

Mi 9.6: turn left off CR-7 onto Forest Road xx; go through gate and continue 0.2 miles on a dirt road.

Mi 9.8: park on left (N) side of road. **STOP 2-1.** (Jim McCalpin). Three Glaciers Lake shore zone features of Three Glaciers Lake

Stop 2-1: Wave-cut benches and other shorezone features of Three Glaciers Lake; Mt. Massive Lakes landslide and Kobe landslide; type section of the Dry Union Formation.

Keenan Lee first noticed anomalous shorezone features in the Arkansas valley upstream of Twin Lakes, the most prominent of which is the Kobe landslide (discussed later). During STATEMAP mapping of the Leadville South quad in 2008, more features and deposits were observed at about the same elevation as the top of the Kobe landslide deposit. At Stop 2-1 we will look at two levels of horizontal platforms cut onto the surface of the hummocky topography of the Mt. Massive Lakes landslide (Fig. 21). these platforms lie between about 9400 ft and 9480 ft, and affect the entire surface of the landslide between those elevations, giving it a different morphology than the rest of the landslide. In the geologic map (Fig. 23) this zone of platforms covered with lag gravel was mapped as a different age (Qlsi) than the rest of the slide. Although we now think that the morphology was created after landsliding by wave action, we have kept the original unit designation. These platforms are best developed on the east side of the Arkansas River, where the fetch of waves would have been in the longest in the Lake.

Qlg Shoreline gravel deposits (middle Pleistocene) – Coarse gravel deposited on poorly-preserved shoreline platforms of Pleistocene moraine-dammed lakes ("Three Glaciers Lake" of Lee, 2008) between about 9,400 and 9,480 ft elevation. Deposits are well sorted, well stratified, clast-supported, small pebble to small cobble gravel. Clasts are subrounded and slightly weathered on the surface although this may be a resistant lag overlying more weathered gravel in the subsurface. Occasional boulders exist that may have rolled down onto the active platform from the wave-cut cliff; these are partly buried by colluvium. Shoreline platforms range in morphology, from gently sloping benches eroded into the high terraces, to flat hilltops planed off by wave action on landslide complexes (Fig. 13) such as the Mt. Massive Lakes complex. Thickness ranges from near zero (a thin gravel lag) to as much as 5 ft. Three Glaciers Lake was probably dammed by the Pine Creek glacier (south of the quadrangle) in pre-Bull Lake time (Nebraskan, ca. 1.4 Ma; Kansan, ca. 600 ka), Bull Lake time (ca. 150 ka), and Pinedale time (35 ka to 15 ka) (Scott, 1984; Lee, 2005, 2008).

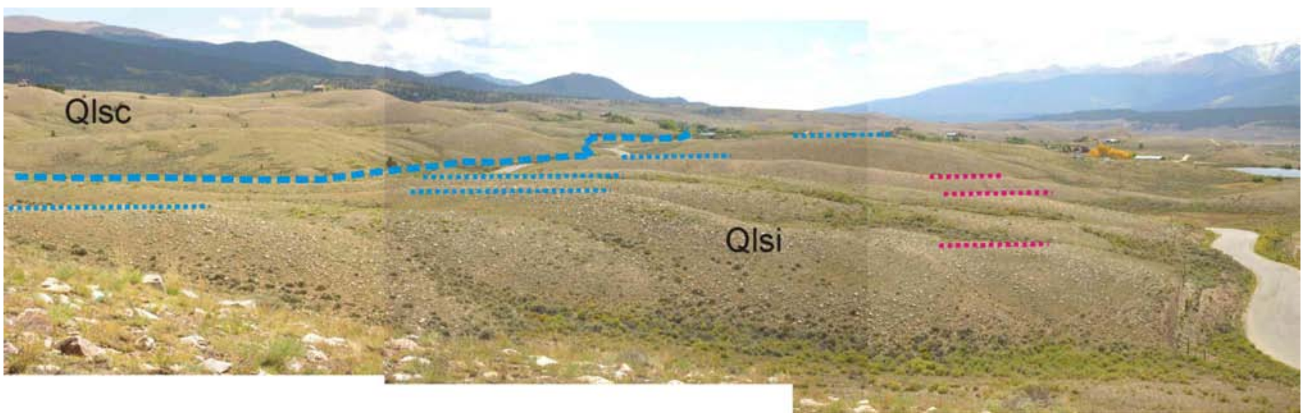


Figure 21. View of beveled, gravel-capped hilltops on the western flank of the Mt. Massive landslide complex, planed off by wave action of Three Glaciers Lake. View is to south from the northwest corner of the landslide complex (Qlsc, Qlsi); note County Road 7 at far right. Thick blue line shows upper limit of lake erosion (ca. 9480 ft elevation); thin blue dots and thin pink dots show two different shoreline levels. [UTM Z13, NAD27, 386880m N, 4336040m N].

Farther to the south on both sides of the Arkansas Valley, embayments created by tributary valleys contain fine-grained, well-stratified deposits that we interpret as nearshore lacustrine deposits (Fig. 22). These deposits were placed in map unit Qlgs (see description below). We do not have time to visit those localities today, which is unfortunate, because there is controversy about both their origin and age. Present OSL ages indicate the beds are 65ka to >204 ka (pre-Pinedale), which accords with their elevation well about the ice dam at 9360 ft. So far we have not identified any shoreline features or deposits that would correlate with a Pinedale lake at 9360 ft elevation.

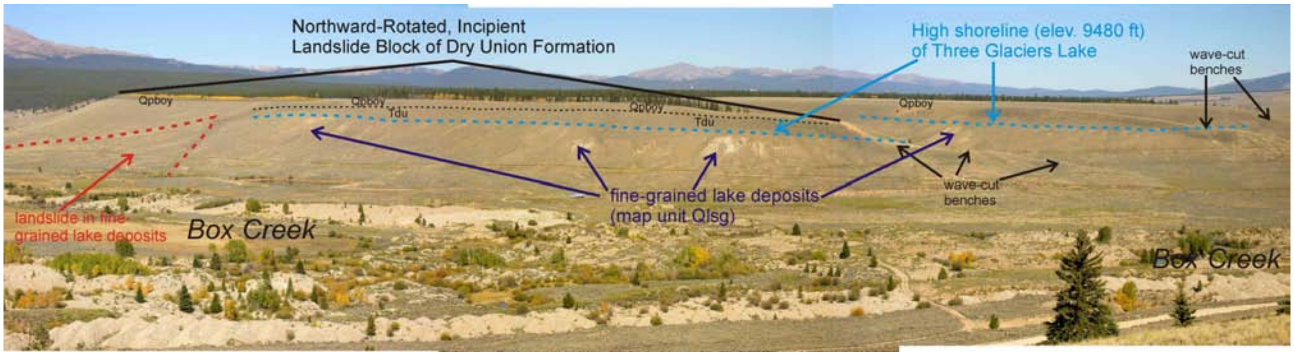


Figure 22. Annotated photo showing fine-grained lacustrine deposits (our map unit Qlgs) of Three Glaciers Lake mantling mesa sideslopes in the valley of Box Creek. View is to the north. Deposition rates were high in this valley because it formed a shallow reentrant bay in the lake that was fed by large meltwater streams from Sawatch Range paleoglaciers. By comparison, slopes at the same elevation east of the Arkansas River have discontinuous lacustrine deposits or erosional landforms. [UTM Z13, NAD27, 384660m E, 4331000m N].

Qlgs Littoral sand and gravel deposits (middle Pleistocene) – Lacustrine sand, silt, clay, and minor gravel deposited in shallow water below the shorelines of Three Glaciers Lake between about 9,340 and 9,480 ft elevation. Deposits are well sorted, well stratified, generally clast-supported. Alternating beds of green-gray clayey sand to sandy clay; cross-bedded coarse sand and granules; small pebble gravel. Mantles all the slopes in the Box Creek reentrant in the southwest corner of the quadrangle below 9,480 ft, which erodes into badlands and gullies (Fig. 14). May correlate with “older lacustrine deposits” that lie beneath the Pinedale terminal moraine of Lake Creek (south of the quadrangle) as described by Nelson and others (1984, p. 10). Also present east of the Arkansas River. Dated by optically-stimulated luminescence at 143-146 ka near top (Bull Lake age), and 204 ka near middle. Exposed thickness is up 33 ft.

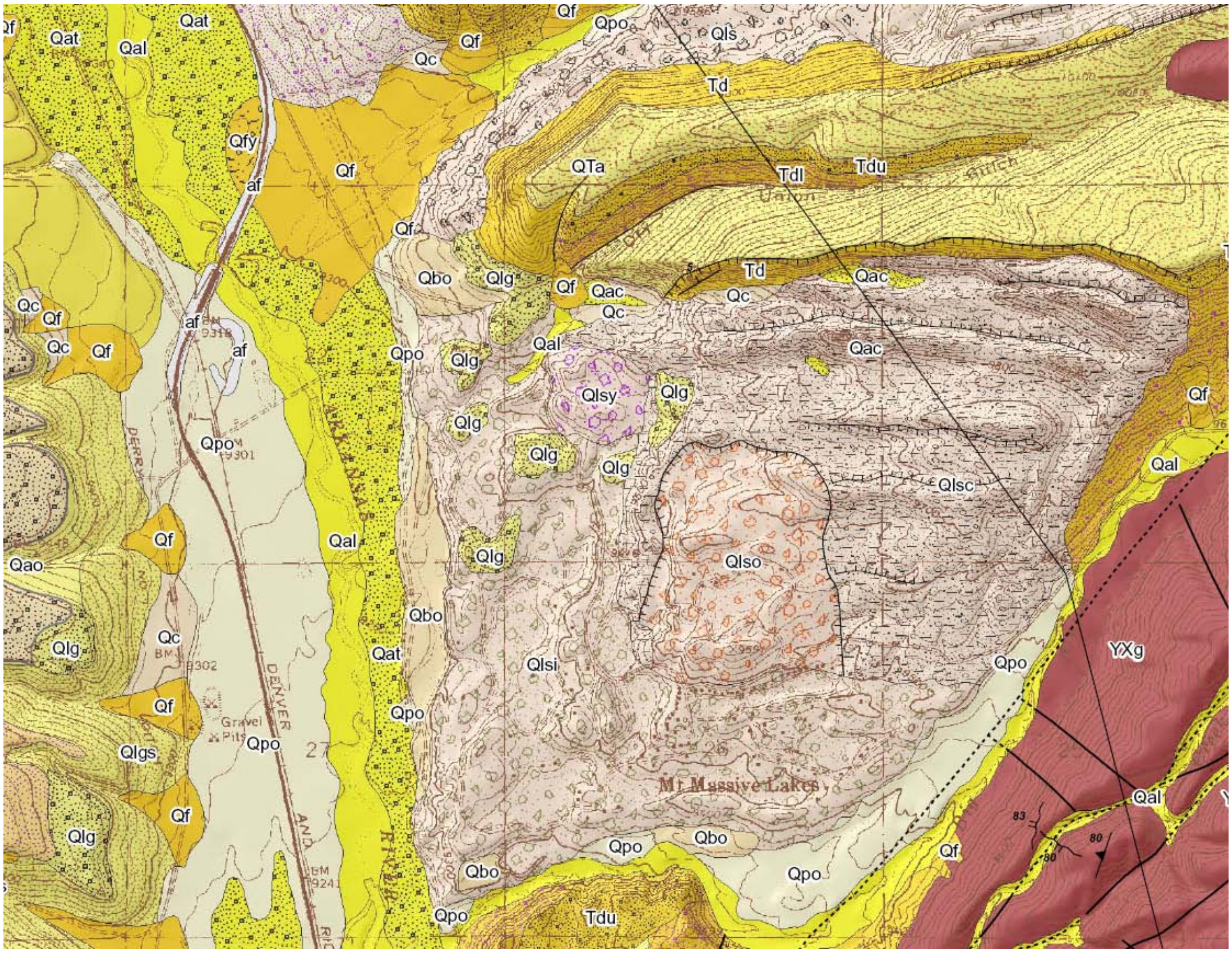


Figure 23. Geologic map of the area around Stop 2-1.

Landslides around the lake rim are another indirect indicator of Three Glaciers Lake. The Kobe landslide is the best discrete landslide example. Its high width-to-length ratio and low-angle deposit composed of multiple slivers is similar to two types of landslides: (1) landslides in sensitive marine clays (quick clays), and (2) landslides caused by rapid drawdown of reservoirs. Lee (2008) surmised that the Kobe landslide formed when Three Glaciers Lake suddenly drained. To date, there have been no attempts to date the landslide.

In addition, the slopes leading down from the pediment surfaces to the Arkansas Valley floor are all composed of hummocky topography, quite unlike the usual slopes below pediments. Although there are no obvious arcuate headscarps or definable landslide lobes in most of this terrain, it all looks “funny.”

Finally, a section of the Qpboy pediment north of Box Creek appears to have rotated and slid southward (Fig. 24). A similar situation affects the ridge of Dry Union Fm. between Dry Union Gulch and the Mt. Massive Lakes landslide. In both cases, the pediment surface has apparently been rotated down-to-the-north.

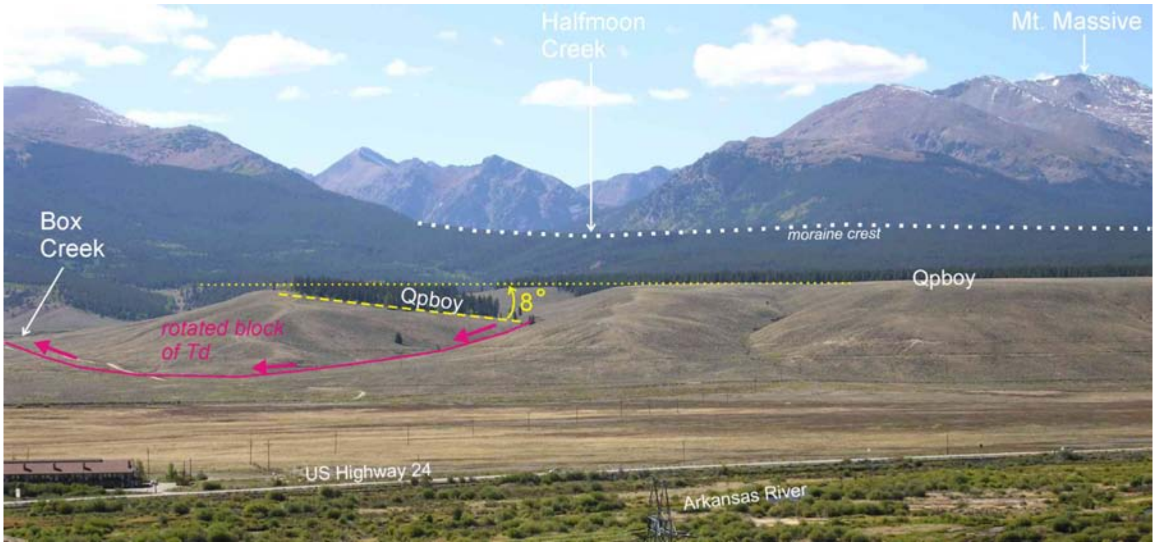


Figure 24. Telephoto view looking west at the rotated slump block (pink) involving the southernmost part of the high terrace (Qpboy) on the west side of the Arkansas River, opposite Kobe. Note the backtilt (down to the north) of the forested pediment surface. The failed slope was mostly submerged by the highstand of Three Glaciers Lake, thus the slump may have been triggered by rapid drawdown.

Type Locality of the Dry Union Formation

Although Tweto defined the Dry Union Formation for exposures near the mouth of Dry Union Gulch, we could find no published measured section for his type section. The best exposure of the Dry Union lies just east of the mouth of Dry Union Gulch (Fig. 25), so in 2008 J. McCalpin measured a 230 ft-high section there (Fig. 26). About half the section is composed of uncemented, massive brown sand that could be fluvial or eolian. The other half of the section is coarser sands and gravels in channels, which have been cemented to sandstone and conglomerate. The entire section is oxidized brown or cemented with calcium carbonate. It looks very different from the unoxidized, uncemented beds mapped by McCalpin et al (2010b) as shoreline deposits of Three Glaciers Lake.

In addition, the beds at the type locality all dip north to northwest, as a result of the rotation of the block at the head of the Mt. Massive Lakes landslide. The lacustrine beds mapped by McCalpin et al. on the north slope of Box Creek (and dated by S. Mahan at 65-204 ka) are not tilted, even though they were deposited on a rotated block of pediment (Fig. 24). For these reasons, McCalpin et al. (2010b) do not think the fine-grained beds at Box Creek, or at other sampled localities, are part of the Dry Union Formation.

Figure 25. Photograph of the type locality of the Dry Union Formation (Tweto, 1961), on the headscarp of the Mt. Massive Lakes landslide, south of lower Dry Union Gulch. View is to the east. The lower 30 ft of the exposure is mantled with slopewash and is not part of the measured stratigraphic section in part b of this Figure. The cemented fluvial channel gravels at center strike N60°E and dip 5°NW. However, the pediment gravel surface at the top of the headscarp (ridge between headscarp and Dry Union Gulch) now dips 10°N, so the entire ridge has presumably been rotated 10° by incipient landslide movement. If so, this implies that the pre-rotation dip of the Dry Union Formation was about 5°S, similar to dips measured in Dry Union strata in Empire Gulch (this report) and at Mt. Elbert Forebay (USBR, 1981).

Tdu Dry Union Formation, upper part (Pliocene?) – Mapped below the tops of the high terraces (mesas) of the Arkansas Valley, from the base of the Quaternary pediment gravels, downward for about 100 ft on the mesa-flanking slopes (i.e., from 40 to 140 ft beneath the terrace surface). On airphotos unit is portrayed by smooth tones, light brown to white north of Big Union Creek, and reddish-brown south of Big Union Creek. Slopes are generally covered by friable gravelly colluvium, suggesting a high gravel component. Rarely exposed in outcrop, except on south-facing slopes north and south of Dry Union Gulch where it is the type locality of Tweto (1961), dominated by fluvial overbank/eolian sands, fluvial channel gravels (cemented into ledges), and rare paleosols (Fig. 17a). Probably correlative to the lower part of the "Wash" defined by early publications of Emmons and Capps, but may partly correlate with the stratified "Lake Beds" unit of the same authors. Correlative with unit Tdu2 of USBR (1981), described as "sand and gravel series, consisting of gravels in a tan to brown clay-silt-sand matrix with some crude stratification." Fluvial channel beds strike N60°E and dip 5°NW, but have probably been rotated by landslide movement. Exposed thickness ranges from about 100 ft in the northern part of the quad to more than 250 ft at the type locality south of Dry Union Gulch (Fig. 17b).

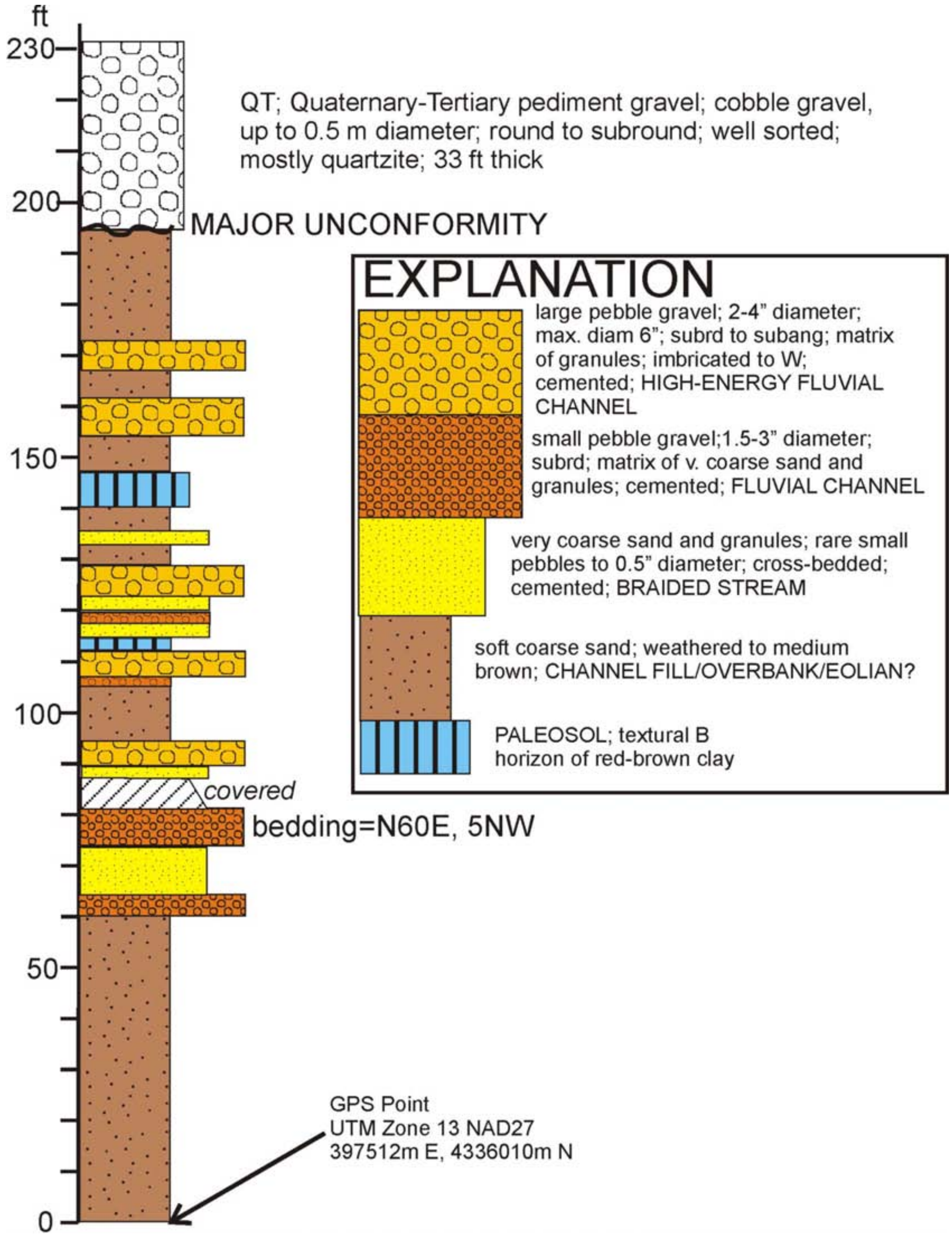


Figure 26. Measured section 233 ft thick at the type locality of the Dry Union Formation (Tweto, 1961), on the headscarp of the Mt. Massive Lakes landslide, south of lower Dry Union Gulch. This exposure is part of the upper part of the Dry Union Formation (Tdu) as defined in this report, correlative with the upper part of the Dry Union Formation ("sand and gravel series") defined by USBR (1981). Degree of cementation is indicated by width of bed. Fluvial channel development and cementation here is anomalously strong, compared to other outcrops of the upper Dry Union Formation.

Depart stop 2-1 by backtracking out to HWY 24. Either turn onto CO RD 10, or drive straight across HWY 24 onto CO RD 10. Reset your odometer. From HWY 24, Stop 2-2 is 6.6 miles. Follow CO RD 10 to the south. After 5.8 miles and a 5-point intersection, take a sharp left onto Edward E Hill DR. After 1.6 miles, turn right onto Reva Ridge RD and park in the pullout on your left a couple hundred feet up the road (6.6 mi).

Stop 2-2: Twin Lakes/Clear Creek moraine sequence; 3 Glaciers dam

This stop involves a short hike (~100 meters) to an overview of the Twin Lakes reservoir (Figs. 27, 28). The viewpoint is atop the Pinedale terminal moraine, with a view of the Pinedale terminal moraines to the south, the Arkansas River to the southeast, and the classic U-shaped morphology of the Lake Creek valley to the west. The stop will serve as an introduction to the moraine morphostratigraphy of the study area.

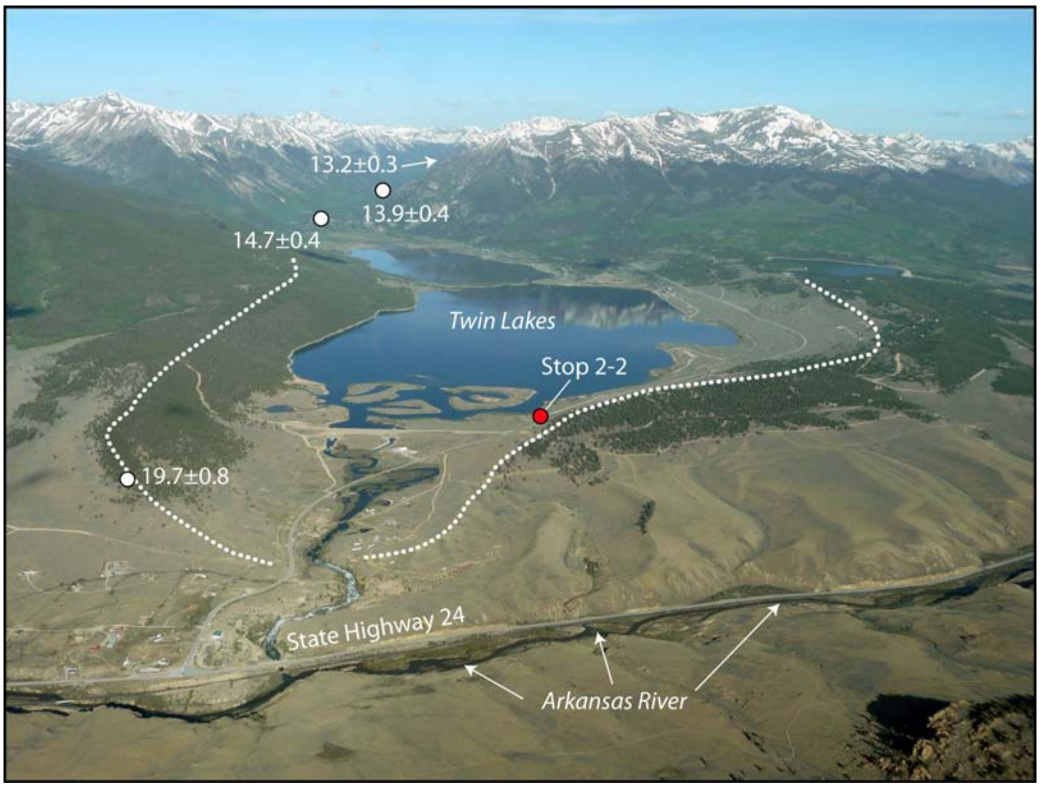


Figure 27. Twin Lakes/Lake Creek reservoir. Dashed line demarcates the terminal Pinedale moraine. Also shown are the approximate locations of ^{10}Be ages (ka) fro the valley.

The Lake Creek valley emanates from the north-south trending Sawatch Range. During the Bull Lake (MIS 6) and Pinedale (MIS 2) glaciations, the Lake Creek glacier extended just beyond the valley mouth onto the Arkansas River valley floor (Fig. 27). However, during both of these glaciations, it is unlikely that the Lake Creek glacier extended far enough to dam the Arkansas River, although the Lake Creek glacier may have forced the river slightly east. Bull Lake deposits within the drainage were initially identified based on soil profiles by Shroba (1977) and Nelson and Shroba (1988).

Deposits of Bull Lake age are relatively scarce as they were overridden and buried by the subsequent Pinedale advance of the Lake Creek glacier. Bull Lake remnants can be found on the north side of the valley (left-lateral moraine) and are characterized by a subdued surface morphology and general lack of moraine boulders (Lee, 2010).

Lee (2010) recognized at least 3 Pinedale advances, with the first being the most extensive, and the second and third advances building the most prominent lateral moraines preserved in valley. In addition, a recessional moraine divides the reservoir into two separate lakes about 5 km upvalley from the terminal moraine complex. Pinedale moraines are sharp-crested with boulders exposed along their crests. The only absolute age control from the Pinedale end moraines in the Lake Creek valley comes from a single ^{10}Be age of 19.7 ± 0.5 ka (Schlingden, 2000). We obtained ^{10}Be ages of 14.7 ± 0.4 ka, 13.9 ± 0.4 ka, and 13.2 ± 0.4 ka from glacially sculpted bedrock at locations ~10, ~16, and ~25 km upvalley of the terminal moraine in Lake Creek valley.

Depart stop 2-2 and go back to the intersection of Reva Ridge RD and Edward E Hill DR. Take a right and stay on Edward E Hill DR to the bottom of the hill and take a right onto CO RD 30 out to the paved road, HWY 82 (7.3 mi). Take a left onto HWY 82 and take it back to HWY 24 (8.5 mi). Take a right onto HWY 24 and proceed south 3.9 miles and turn right onto CO RD 390 (12.4 mi). Drive up CO RD 390 for 5.3 miles, and pull into a dirt pullout on the left (17.7 mi).

Stop 2-3: Clear Creek valley chronology

This stop is just inboard of the range front where the first outcrop of glacially sculpted bedrock is available for sampling on the valley bottom. Our strategy of dating glacially sculpted bedrock to constrain deglaciation will be discussed.

The Clear Creek valley is located one valley south of the Lake Creek valley. Here, Pinedale lateral moraines are the most prominent features, but a large section of the left lateral Bull Lake moraine has been preserved as well as a small section of the right lateral Bull Lake moraine. Pinedale moraines are relatively steep and their crests are boulder-rich compared to Bull Lake moraine crests (Fig. 29). Of particular interest is the oversteepened, granodiorite bedrock cliff opposite the Mouth of the Clear Creek valley and across the Arkansas River (Lee, 2010). We hypothesize that the eastward flowing glacier crossed the Arkansas River valley and abutted against the opposing cliff, damming a large lake upvalley (north) of the glacier terminus. Till is currently visible on the east side of the Arkansas River just north of its confluence with Clear Creek. Dam failure resulted in the catastrophic southerly release of floodwaters carrying debris sourced from the opposing granodiorite cliff and the Pinedale lateral moraines. The presence of two boulder-rich terraces just downstream of the Clear Creek valley indicate that the Arkansas River became dammed at least twice during the Pinedale glaciation (see below). Or, perhaps only once during the Pinedale glaciation (lower terrace) with the upper terrace the result of a Bull Lake age flood event (Lee, 2010).

Based on a projection of the Pinedale moraine crest across the valley, Lee estimates that the river was dammed to a minimum height of about 9360 ft (~2855 m) about 530 ft (150 m) above the present river channel. On the basis of this estimated lake surface height, Bush (unpublished) determined that the impounded Pinedale-age lake would have had an area of about 43.6 km^2, with a volume of about 1.72 km^3. McCalpin et al. (2010) report middle Pleistocene lake shorelines at altitudes up to about

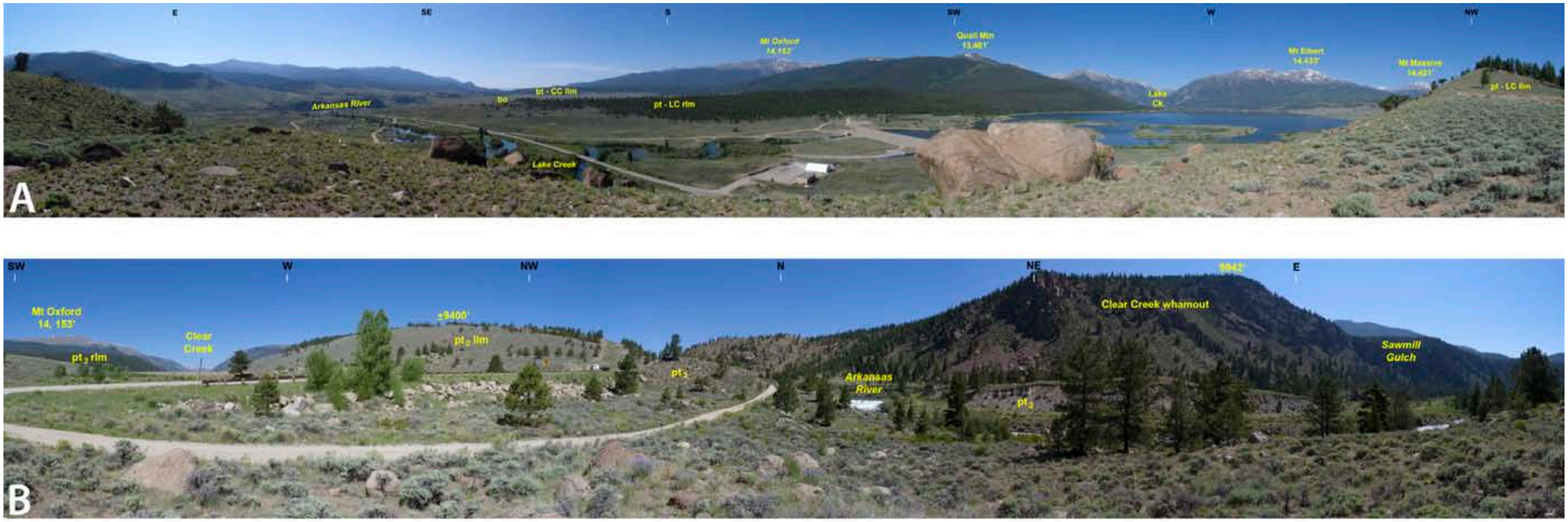


Figure 28. A) Panoramic view of Twin Lakes/Clear Creek valley. Pt – Pinedale-age till, bt- Bull Lake-age till, bo – Bull Lake outwash. B) Panoramic view of the Clear Creek valleymouth where the Clear Creek glacier dammed the Arkansas River.

9480 ft (2890 m) upvalley from the moraine dam. A middle Pleistocene lake at this height would have had an area of about 68.2 km^2, with a volume of about 3.77 km^3 (Bush, unpublished).

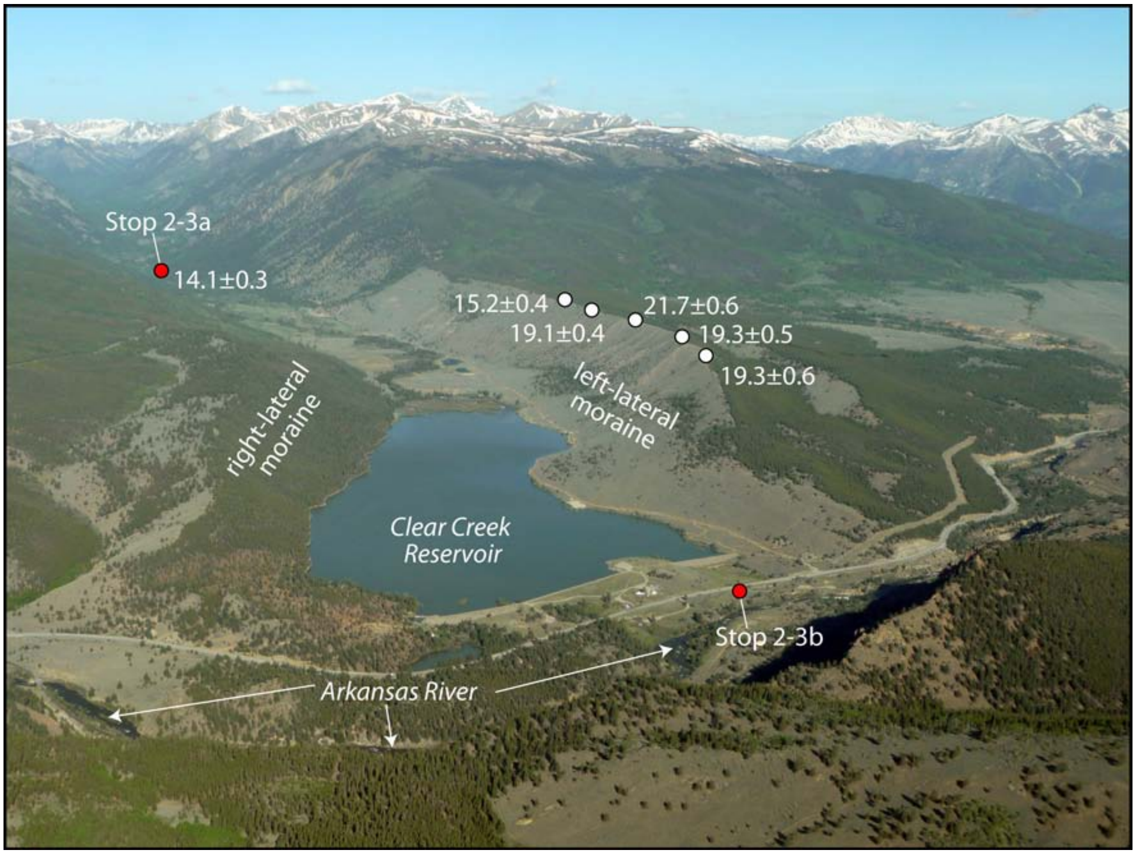


Figure 29. Clear Creek valley with associated ^{10}Be ages from the left-lateral Pinedale moraine and valley bottom bedrock.

We obtained ^{10}Be ages from five moraine boulders/clasts on the terminal left lateral Pinedale moraine. ^{10}Be ages are 21.7 ± 0.6 ka, 19.5 ± 0.6 ka, 19.3 ± 0.4 ka, 19.1 ± 0.4 ka and 15.2 ± 0.4 ka (clast). In addition, a sample collected from glacially polished bedrock located ~9 km upvalley from the Pinedale terminal moraine yielded an age of 14.1 ± 0.3 ka (this stop). While ^{10}Be ages from the terminal Pinedale moraine indicate the Clear Creek glacier retreated from its Pinedale maximum ca. 19.3 ka, the bedrock ^{10}Be age suggests ice remained near its maximum extent until ca. 14.1 ka (Young et al., in press).

Drive 5.3 miles back down CO RD 390 to HW 24 (23.1 mi), take a right, and a quick left onto a dirt road after 0.15 miles (23.2 mi).

Stop 2-3b (optional)

Pinedale till deposits located east of the modern Arkansas River.

Take HWY 24 south for 1.6 miles, and turn right onto CO RD 388 (24.8 mi). Proceed along this dirt road, crossing Pine Creek, to the parking lot and gate (25.6 mi). If congested or too rocky, park earlier. Trailhead for short hike up hill (to the east-southeast) to the overview on the Pinedale moraine crest.

Stop 2-4: Pine Creek valley

This stop involves a ½ mile (0.8 km) hike in each direction to atop the Pinedale left lateral moraine in the Pine Creek valley. The elevation gain is ca. 100 meters. There is a view to the south of the flood terraces. Extensive ^{10}Be dating was done on this lateral moraine, which will be discussed. We will also introduce the terrace stratigraphy that exists immediately downstream of the Pine Creek valley.

Extensive Bull Lake and Pinedale age moraines are preserved in the Pine Creek valley.^{10}Be ages (n = 10) from the left lateral Bull Lake moraine (Dry Creek Gulch) range from ca. 72 ka to 3.4 ka (Briner, 2009). While there is some evidence for a glacier advance during the early- or middle-Wisconsin in the western US (e.g. Colman and Pierce, 1986), most western US records indicate the presence of only Bull Lake and Pinedale age deposits (e.g. Pierce, 2004; Licciardi and Pierce, 2008); this is also thought to be true for the Arkansas River valley (Nelson and Shroba, 1998). Thus, it is likely that ^{10}Be ages from Dry Creek Gulch reflect post-depositional processes (i.e. exhumation), rather than the age of moraine deposition (Briner, 2009).

The position of the Pinedale moraines suggests that the Pine Creek glacier likely crossed the Arkansas River valley and abutted against the opposing valley (Fig. 29). Twelve^{10}Be ages from the right lateral Pinedale moraine range from 24.5 ± 0.7 ka to 13.5 ± 0.4 ka. However, after removing one younger outlier (13.5 ± 0.4 ka), ^{10}Be ages fall into two distinct modes at 22.4 ± 1.4 ka (n = 5) and 15.8 ± 0.4 ka (n = 6). This bimodal distribution is notable because ^{10}Be ages come from a *single-crested* moraine ridge indicating that an advance of the Pine Creek glacier culminated ca. 22.4 ka, followed by a ca. 6000 year period of reduced ice extent, and finally a readvance of the glacier to the same position culminating ca. 15.8 ca. Moreover, ^{10}Be ages of 15.6 ± 0.4 ka and 15.3 ± 0.3 ka from valley bottom bedrock located ~4 km upvalley of the terminal Pinedale moraine, further suggest that the Pine Creek glacier remained near its Pinedale extent until ca. 15.8 – 15.3 ka (Young et al., in press).

Drive the 0.75 miles back out to HWY 25 (26.3 mi). Turn right (south) on HWY 24 and proceed 2.2 miles (28.5 mi). Turn left onto the dirt road, cross the Arkansas River, and turn left 0.25 miles from HWY 24 (28.8 mi) on the dirt track; proceed up the dirt track and park after 0.25 mi (29 mi).

Stop 2-5: Upper and lower flood terraces

This site is host to some of the largest flood boulders found in the Arkansas River valley ranging in size from ~1.5 to 15 m in length (Fig. 30). This will be our first stop at terrace flood boulders, and we will visit boulders that were sampled for ^{10}Be dating.

Near the mouth of Pine Creek, upper terrace flood gravels are ~18 m thick and thin to ~5 m in thickness about 11 km farther downstream (Lee, 2010). Lower terrace flood gravels range from ~9 m in thickness just below the damsite to at least 6 m farther downstream (Lee, 2010). The upper terrace ranges from 10–20 m above the modern river channel, with a median height of ~15 m, the lower terrace 2.5 m above the channel, with a median height of ~6 m (Lee, 2010). Scott (1975) originally mapped the upper and lower terraces as both being Pinedale in age; however, others (e.g. Lee, 2010) have suggested that the upper terrace is Bull Lake in age.

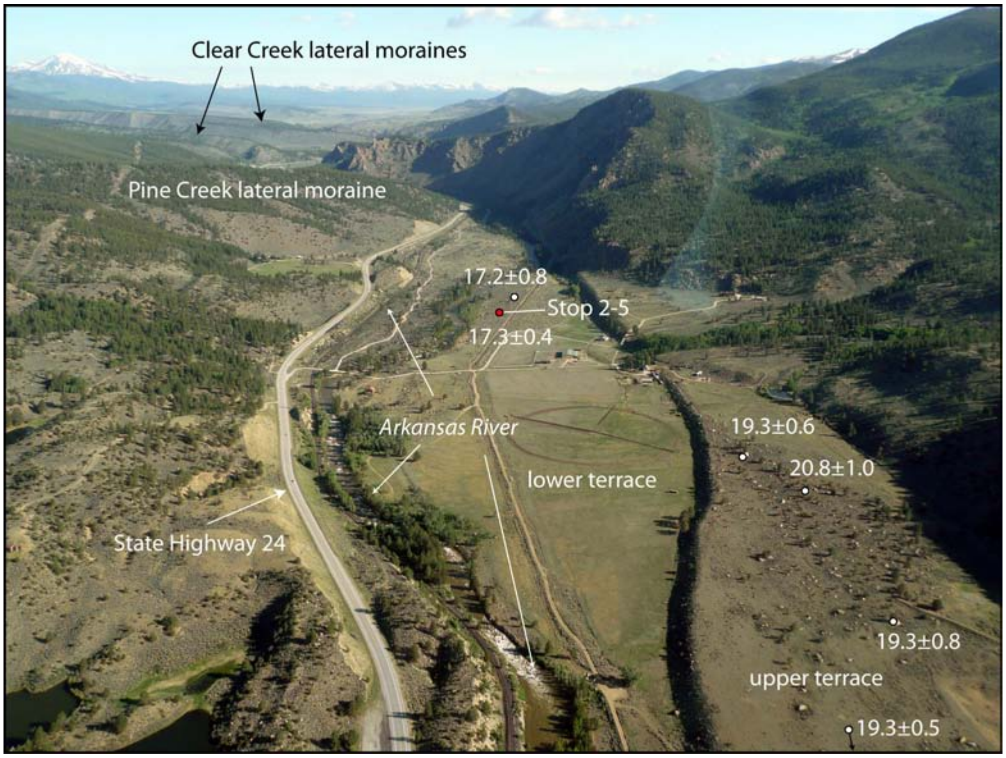


Figure 30. Upper and lower flood terraces with ^{10}Be ages from each surface.

Based on her lake-volume determinations discussed at stop 8 and different empirical models of the relationship between glacier/moraine-dammed lake volumes and outburst-flood magnitudes (Clague and Mathews, 1973; Costa, 1988; Desloges et al., 1989), Bush (unpublished) estimated that peak discharge from the a 2855 m surface elevation Pinedale lake (Lee, 2010) was in the range of 11,000 to 21,000 m^3/sec. Discharge from the higher “middle Pleistocene” lake identified by McCalpin et al. (2010) was in the range of 18,000 to 34,000 m^3/sec. We have not yet identified slackwater deposits or other water-surface indicators associated with the outburst floods, so have not been able to model discharges more accurately using step-backwater methods.

At this location, we obtained four ^{10}Be ages from boulders positioned on the upper terrace and two ages from boulders resting on the lower terrace. On the upper terrace, the^{10}Be ages are 20.9 ± 1.0 ka, 19.3 ± 0.5 ka, 19.1 ± 0.8 ka and 19.1 ± 0.6 ka. Lower terrace boulders returned ages of 17.3 ± 0.4 ka and 17.2 ± 0.8 ka. We note that the cluster of ^{10}Be ages on the upper terrace at 19.2 ± 0.1 ka (*n* = 3) correlates with the mode of ^{10}Be ages at ca. 19.3 ka from the Clear Creek Pinedale moraine, implying that the Clear Creek glacier acted as the main ice dam. Furthermore, all ^{10}Be ages from this sampling location suggest that both the upper and lower terraces were deposited during the Pinedale glaciation.

Turn around and get back to HWY 24, turn left (south). Drive 3 miles, and turn left onto CO RD 385 (32.5 mi). Drive along the dirt road for 0.25 miles and park or turn into the driveway of the farmhouse on the left (33 mi).

Stop 2-6: Lower flood terrace

This stop (Fig. 31) is the second stop at terrace boulders that were sampled for ^{10}Be dating. This stop will serve as the final stop and thus the conclusion of the field trip prior to the drive back to Denver.

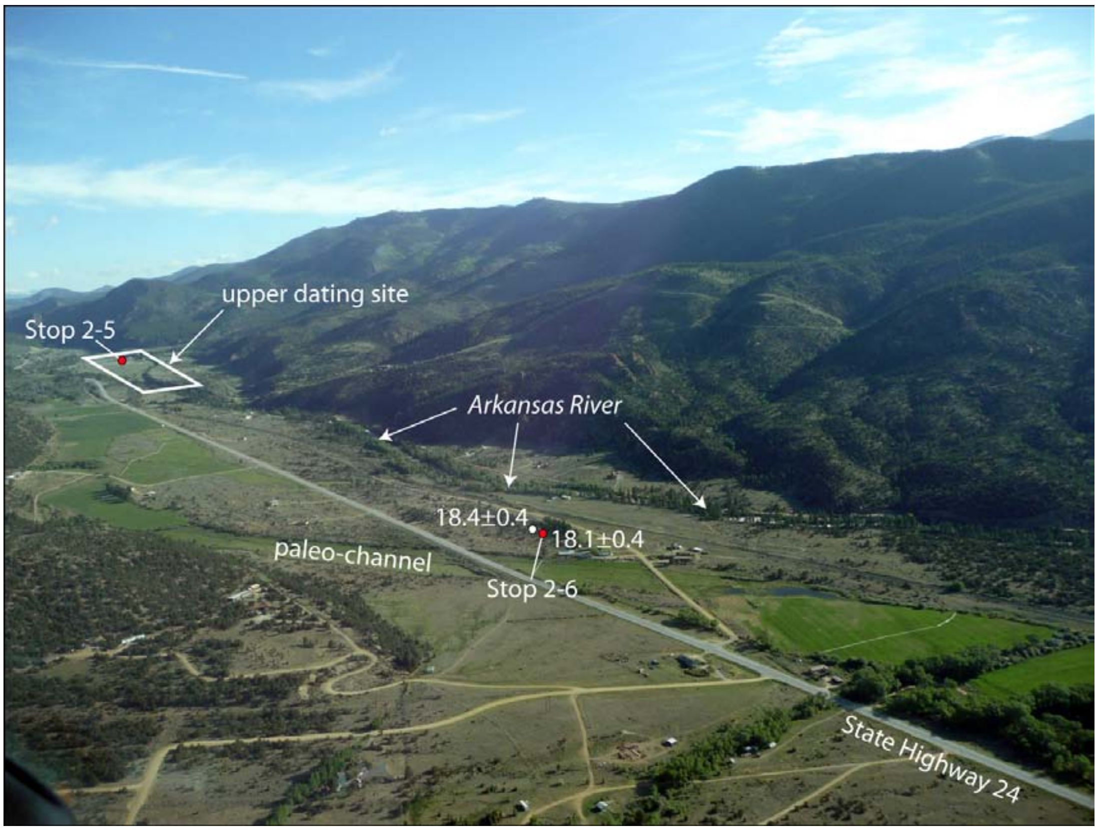


Figure 31. View of lower ^{10}Be dating site along the Arkansas River.

At this location, we obtained two ^{10}Be ages of 18.4 ± 0.4 ka and 18.1 ± 0.4 ka from boulders resting on the lower flood terrace (Fig. 31). However, these ages raise the possibility that boulders at this location are resting on a terrace that is geomorphically and chronologically separate from the lower terrace viewed at stop 2-5 (i.e. a terrace intermediate in height and age between the upper and lower terraces). Outburst floods are geologically instantaneous events, and this should be reflected in a ^{10}Be age distribution from a uniform depositional surface, but ^{10}Be ages from the lower terrace seem to be bimodal across the two sampling locations. Even so, all ^{10}Be ages overlap at 2-sigma confidence and have a mean age of 17.8 ± 0.6 ka. It is interesting that the age of the lower terrace is not expressed in the Clear Creek or Pine Creek moraine records. Nonetheless, deposition of the lower flood terrace requires that glaciers were at or near their Pinedale maxima in order to dam the Arkansas River. Based on the close correlation between Clear Creek moraine abandonment and upper flood terrace ages (19.3-19.2 ka), we hypothesize that the Clear Creek valley glacier was near its Pinedale maximum and damming the Arkansas River ca. 17.8 ka (Fig. 32).

From this stop we will continue downvalley about 10 miles (16 km) through the town of Buena Vista, before turning northeast to return to Denver. Along the way, we will pass by flood boulders on the three higher terrace surfaces. Flood boulders occur on terraces at least as far downstream as Parkdale (Scott, 1984) about 130 km downvalley from this stop.

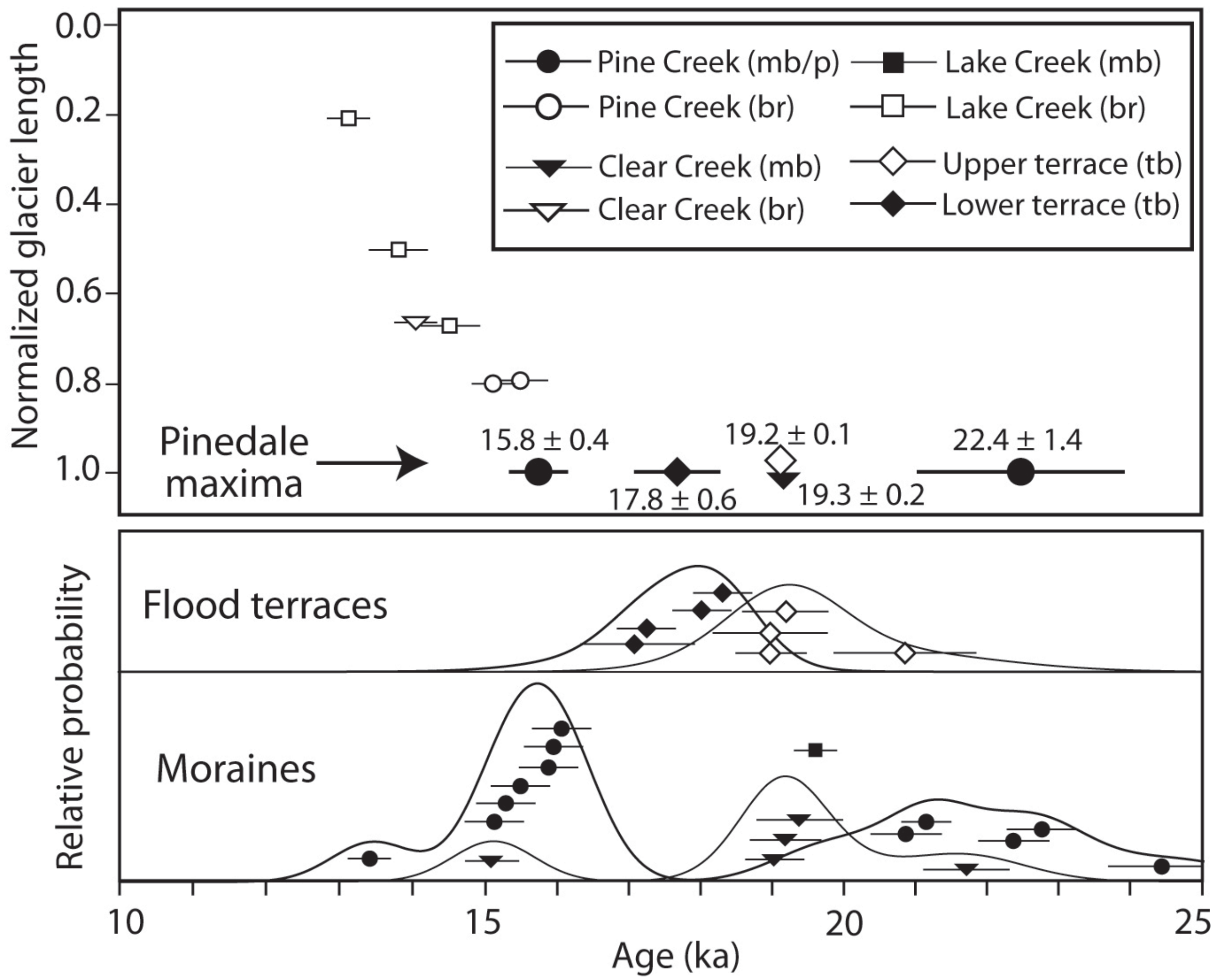


Figure 32. All ^{10}Be ages from the Arkansas River valley. ^{10}Be ages from moraines and flood terraces are plotted individually along with the summed probability distribution of each geomorphic feature (bottom panel). Glacial culminations are shown in the top panel along with the relative location of ^{10}Be ages from glacially polished bedrock within each valley.

Once back on HWY 24, we will proceed south through Buena Vista (7.3 miles to the south) to HWY 285 (42.7 miles), which will take us back to Denver.

END OF FIELD TRIP

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