Geology of the upper Arkansas River valley, Colorado - a field mapper's perspective.

Fieldtrip guide for the Colorado Scientific Society

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Leaders: Karl Kellogg (USGS, Emeritus), Cal Ruleman (USGS), and Scott Minor (USGS)

Cell numbers: (303) 518-2233 (Karl); (303) 859-091 (Cal); (303) 478-8430 (Scott)

The upper Arkansas River valley comprises a northern portion of the Rio Grande rift, a north-trending zone of crustal extension that extends from Mexico northward into northern Colorado. Since the Late Cretaceous, and continuing until the late Oligocene, the region has experienced multiple phases of plutonism and volcanism. We will visit some of the extensive pre-rift (33-39 Ma) volcanic rocks, which include several major ignimbrites. The initiation of rifting at about 30 Ma (Oligocene), associated with the onset of regional extension, resulted in the early formation of the present basin configuration and general west tilting. Rifting was also associated with a change from calc-alkaline (mostly andesitic to dacitic) volcanics to bi-modal rhyolite-basalt assemblages, including the rhyolite flows of the 30 Ma Nathrop volcanics.

Extensive basin-fill deposits of the lower Pliocene? to Miocene Dry Union Formation will be examined, and we'll have a lively discussion of whether we should extend the age of the Dry Union to as young as middle Pleistocene, an idea that involves Cal's new regional neotectonic and paleoclimatic model for the region.

The Colorado mineral belt traverses the northern part of the valley and contains the world-class ore deposits of the Leadville mining district (mostly bearing Pb, Zn, Ag, and some Au and Cu); we will have a brief discussion of the geology of the district.

The glacial history has been recently enhanced by abundant cosmogenic dating of multiple generations of glacial till and associated outwash deposits, revealing the rapidity of the Last Glacial Maximum's (Pinedale's) demise as well as the overall timing of the Pleistocene glacial record. We will visit large glacial outburst flood deposits associated with the Last Glacial Maximum damming of the Upper Arkansas Valley by the Clear Creek glacial system, a late Pleistocene (~22 ka) glacial complex that repeatedly dammed the Arkansas River creating monstrous, catastrophic breakout floods, carrying truck-sized boulders downstream. We will also visit a location where the Lava Creek B ash (~640 ka) from the Yellowstone caldera was captured within early glacial outwash deposits, providing a constraint for the onset of Pleistocene glaciations and its associated effects on the landscape. We have worked out the detailed terrace stratigraphy of the valley, which extends as far back as the early Pleistocene and possibly Pliocene and directly relates to the formation of the downstream Royal Gorge. Ultimately, we will examine the geomorphic, paleoseismic, and geodetic relationships related to basin evolution in light of regional neotectonics and paleoclimatic relationships.

We will depart at 8 a.m. sharp Sept. 15 from the parking lot on the west side of the Federal Center in Lakewood, near the light rail terminal (about 2 blocks south of 6th Avenue just east of Union Blvd.) and spend one night in Buena Vista. Our trip will travel along I-70 to the Copper Mountain exit, and then proceed up Tenmile Creek (State Hwy. 91) to Fremont Pass (Stop 1), which defines the north boundary of the Arkansas River valley. We will try to return by 5:30 p.m. Sunday, Sept. 16.

Note: The map on page 15 shows the field trip stops.

Day 1. Fremont Pass to Buena Vista.

Stop 1. Fremont Pass. Overview of the valley and a brief discussion of the Climax molybdenum deposit.

Although the Climax Mine is not in the Arkansas River Valley (it is in Lake County, however), we thought it appropriate for a brief discussion. This summary follows that was given by Bookstrom (1990), McCalpin and others (2012), and Cappa and Bartos (2007). The mine, operated over the years by various owners, is currently operated by Freeport-McMoran Copper and Gold and is reported to be the largest, highest grade and lowest cost molybdenum ore body in the world. The deposit was discovered in 1879 by Charles Senter during the Leadville silver boom, although he did not know at the time what the silvery gray mineral, molybdenite (MoS_2), was. It had no commercial value until it was discovered to dramatically harden steel. Production began in 1915 and by World War I, Climax was producing 90 percent of the world's molybdenum. By 1959, Climax was the world's largest underground mine and by 1972, large-scale open-pit mining began, largely due to block caving of the underground workings. Between 1915 and 1987, Climax produced over 421 million tons of ore yielding almost a million tons of moly. The tremendous amount of tailings filled the upper regions of Tenmile Creek, thereby burying the entire Kokomo mining district. The price of moly plummeted as the many porphyry copper mines produced moly as a byproduct and the mine was shut down in 1995. It reopened in 2012 as the price once again climbed.

The deposit is composed of a stockwork of molybdenite in small veinlets associated with multiple high-silica granite porphyry stocks and rhyolite dikes containing elevated amounts of Fl, Mo, Sn, W, U, Th, and Ni. The ore occurs in three separate dome-shaped shells, each of which is associated with a separate granitic intrusion, which collectively form an intrusive complex emplaced into mostly Proterozoic schists and gneiss during a 9-million year period from 33 to 24 million years ago. The stockwork is associated with pervasive deposition of secondary potassium feldspar and is just above a zone of massive "flooding" of silica-rich fluids that produced much of the secondary quartz. A western, unmined part of the ore body is dropped down several thousand meters along the Mosquito Fault, a major Oligocene and Miocene Rio Grande Rift structure.

From the pass. Proceed to the town of Leadville and turn left on 5th Avenue. Drive about 1 mile to a spot surrounded by mine tailings (south of Fryer Hill Mines).

Stop 2. Leadville mining district.

The world-class ore deposits of the Leadville district have been extensively studied (for example, Emmons and others, 1886; Emmons and Irving, 1907; Emmons and others, 1927; Behre, 1953; Tweto, 1968; Thompson and Arehart, 1990; Wallace, 1993.). The colorful mining history of the Leadville District is described by Cappa and Bartos (2007). Primary sulfide deposits formed at about 39.6±1.7 Ma (Thompson and Arehart, 1990), at a depth of ~5 km by ascending hydrothermal fluids that replaced Paleozoic carbonate rocks (mostly dolomitic Leadville Limestone (or dolostone), but also to a lesser extent Dyer Dolomite and Manitou Dolomite) with blanket-like (manto) deposits of massive sulfides (mostly argentiferous galena and sphalerite). The first economic discovery in late 1859 was placer gold at the mouth of California Gulch (Henderson, 1926), about 4 km southwest of Leadville, which caused an almost immediate gold rush to the district (Parker, 1974). Placer gold deposits in the upper part of California Gulch were some of the richest in Colorado; however, by 1874 these deposits had been extensively mined and there was little placer mining left in the gulch (Henderson, 1926). A popular but questionable story is that blue-black sand that choked sluice boxes in the Leadville district and plagued the early placer miners was later identified as argentiferous lead carbonates. Analyses of these carbonates, conducted in about 1875, indicated that they contained about 1.4 kg of silver per metric ton (about 40 troy ounces per short ton; Scott, 2004). The first lode deposit of silver-bearing lead carbonate was located in 1874. The rich ore bodies on Fryer Hill, just east of Leadville, were discovered in 1878. A number of rich mines also surround the Breece Hill intrusive center (composed mostly of Johnson Gulch Porphyry) about 3 km east of Leadville (Cappa and Bartos, 2007). The mines at Fryer Hill, Breece Hill, and Carbonate Hill produced a considerable amount of silver. Early smelters used charcoal, but smelting of ore really got underway after the first railroads arrived in 1880, carrying coal.

The early lode mining exploited oxidized ores (lead-zinc carbonates) until the deeper sulfide ores were encountered. The primary control on mineralization was an impervious layer, such as shale, quartzite, or porphyry, overlying a carbonate rock. Mineralizing fluids ascended along faults, which are also locally mineralized. The source for the fluids is apparently a deep, unexposed ~40 Ma magma body. Exposed porphyries are essentially unmineralized, although many of them, particularly the Johnson Gulch Porphyry underlying mostly Breece Hill, were emplaced at approximately the same time as the mineralization.

In all, the Leadville mining district produced more than 1.8 billion dollars' worth of silver, lead, zinc, gold, and copper (at then-current prices; Cappa and Bartos, 2007); the first three metals named accounted for 95 percent of the value. Manganese, bismuth, and sulfuric acid from pyrite were also produced as byproducts. Mining continued more or less continually until the last operating mine, the Black Cloud Mine operated by Asarco Grupo Mexico, closed in 1999.

Return to Highway 24 and drive about 3 miles to a view to the north across the railroad tracks of the Malta Gravel, one of the youngest exposures of the Dry Union Formation.

Stop 3. Malta Gravel.

The type-section of the Malta Gravel, exposed here, underlies the oldest middle Pleistocene (<780–190? ka) pre-Bull Lake outwash gravel and is the most prominent, highest (and therefore oldest) surface grading across the Leadville area. Originally identified by Capps (1909), Tweto (1961) constrained the deposits to include all basin fill underlying the high terrace gravels associated with the onset of glaciofluvial process, later subdivided into pre-Bull lake (>190 ka), Bull Lake (~190-120 ka) and Pinedale (~31-12 ka) glaciations (see refs. in Kellogg and others, 2017). These overlying terrace gravel deposits consist of subrounded to rounded cobble to boulder gravel composed mostly of Mesoproterozoic metamorphic and igneous lithologies.

This exposure along the railroad tracks reveals reddish brown, weathered and oxidized, angular to subrounded pebble to cobble gravels in a sandy silt and clay matrix. The deposit is poorly sorted and bedding is massive to weakly stratified, with silty sand and sandy pebble lenses. Provenance of the deposit is locally sourced from the East Fork Arkansas River and Mount Zion region to the northeast and consists of Mesoproterozoic metamorphic and igneous rocks, competent Paleozoic Sawatch Quartzite, and Leadville Limestone (dolomite), and a mix of late Tertiary porphyry lithologies.

The deposit is strikingly similar in its oxidized pedogenic development and weathering to exposures of basin fill further to the south within the Miocene and lower Paleocene? Dry Union Formation (Kellogg and others, 2017), and we interpret this to be a deposit in the upper part of the Dry Union.

Drive several miles south to stop just before the bridge over Arkansas River.

Stop 4. Mt. Massive Lakes landslide deposit.

The hummocky topography to the south is a large and relatively young (late Pleistocene and possibly as young as Holocene) landslide deposit, mobilized when glacial runoff provided considerably more water than is now available. Numerous closed depressions have created many lakes within the deposit.

Continue about 11 miles, past the small town of Granite, to Clear Lake turnoff (right turn). We're in the basement high that divides the northern and southern basins of the Arkansas River valley graben. Drive about 0.6 miles west to the parking area.

Stop 5. Clear Creek moraine and "wham out"

Here, we have a good view of the Pinedale moraine that formed during the last glacial maximum (about 31-12 ka). Compared to those formed during previous glaciations, moraines of Pinedale age have sharper crests and more hummocky morphology, and the contained boulders are less weathered. ^{10}Be surface-exposure dating on boulders in this moraine ranges from 22.4±1.4 ka to 19.5±1.8 ka (Brugger, 2007, Briner, 2009, Mason and Ruleman, 2011; Young and others, 2013). These studies also show that the Pinedale glaciers were receding from about 14.4 ± 0.8 ka to 12.6±1.2 ka, with deglaciation rates of about 1 meter per year (Mason and Ruleman, 2011). The Clear Creek glacier flowed across the Arkansas River into the eastern wall, creating an ice dam that impounded the river, creating a lake that periodically broke through the dam catastrophically (Lee, 2010). Three major floods are recognized: two during the Pinedale glaciation and an older one whose age is constrained by the age of the Lava Creek B ash (639±2 ka; Lanphere and others, 2002) which locally overlies the flood deposit. Flood deposits associated with the Bull Lake glaciation (130-70 ka; Lisieki and Raymo, 2005, Pierce, 2004) are not recognized here. The preserved steep-walled embayment east of the river ("whamout" – Keenan Lee's term) was carved during maximum flow during breakouts, plucking out huge boulders and carrying them downstream. The estimates of peak discharge are between 21,000 and 65,000 cubic meters per second (Brugger and others, 2011). Compare this to the record 1994 flood on the Mississippi River of about 28,000 cubic meters per second.

Return to Hwy. 24 and drive south 5 miles to Railroad Bridge (left turn). Follow dirt road on east side of river (County Road 371) 3.2 miles to small dirt road on left; drive up several 100 yards.

Stop 6. Lava Creek B ash deposit.

At this drainage, cut into middle Pleistocene basin-margin slopewash and colluvium, the Lava Creek B ash is beautifully exposed in a highly oxidized, deeply weathered sandy pebble gravel. Originally identified by Glenn Scott (1975), the ash came from a caldera eruption within or near Yellowstone National Park, ca. 631 ka (Matthews and others, 2015). Iron concretions are readily visible throughout the exposure, indicating significant time for weathering, and possible sand dikes (if they're real; you decide) suggest seismogenic disruption of the deposit from events associated with the Sawatch fault zone to the west. We interpret this section to represent a period of regional semi-arid to arid climate and associated weathering.

This unit overlies fluvial boulders (some exceeding >2.7x5.0x8.2 m in size; Kellogg and others, 2017; Lee, 2010), interpreted to represent early glacial outburst floods from the north. If so, then the age for the onset of glaciation is greater than ~630 ka. However, compare this to recent work within the Wind River Range (Dahms, 2005) that concluded pre-Bull Lake glaciation began concurrent with the marine oxygen isotope stage 12 (~478-424 ka) (Lisiecki and Raymo, 2005).

Further to the south, Scott (1975) shows Lava Creek B ash at the top of a terrace underlain by fan and/or glacial outwash deposits (our unit Qg3). Before departure of this site, climb up above this outcrop and look to the south at Turtle and Elephant Rocks (our next stop), where the pre-Bull Lake pathway of the Upper Arkansas River began to incise granitic bedrock.

Continue south on Road 371 about a mile to view large flood boulders.

Stop 7. Large flood boulders.

We'll examine some enormous flood boulders associated with the youngest Pinedale flood. These are not, by any means, the largest in the valley. The largest flood boulder yet observed, from the older of the two recognized Pinedale floods, is on a bench about a mile north of here and measures 14 m long. What was the mechanism by which these boulders were transported? Some have hypothesized they were ice rafted, but more likely they were bed-load boulders, attesting to the tremendous energy associated with the glacial breakouts.

Continue south past Elephant Rock, a prominent landmark, and just after the old train tunnels, turn left County Road 375 and drive about a mile to Turtle Rock CG; drive to west end.

Stop 8. Turtle Rock Camp Ground

Following the Arkansas River to the south, the river goes through an incised bedrock canyon. Based on projections from the Lava Creek B exposure at Stop 6, we are looking at <640 ka of incision, and most likely less than ~200 ky of incision in this short canyon.

We can view two of the most prominent Mesoproterozoic granites in the region here - the Langhoff Gulch Granite and the Elephant Rock Granite. The former is a gray, medium-grained, massive to weakly flow-foliated mostly monzogranite, while the latter is a gray to pinkish-gray coarse-grained monzogranite with microcline phenocrysts as long as 3 cm. Both are about 1,435 Ma old (ages indistinguishable), although the Langhoff Granite intruded and partially assimilated the Elephant Rock Granite, indicating that it is slightly younger.

As we come into Buena Vista, note the large granite outcrop next to the main intersection of town. It may look like bedrock, but it's the top of an enormous flood boulder.

Day 2. Start in Buena Vista

Drive south about 6 miles south on Highway 24 to Fisherman's Bridge. Turn left. Go about 0.8 miles on County Road 301 to County Road 300. Turn right. Proceed about 2 miles to base of Sugarloaf Mountain.

Stop 9. Nathrop topaz rhyolite

The Nathrop topaz rhyolite represents an abrupt change at about 30 Ma from the pre-rift, Andean-style plate-margin calc-alkaline (mostly andesitic to dacitic) magmatism to an anorogenic, alkaline, chemically evolved bimodal (basalt and rhyolite) suite (Lipman and Mehnert, 1975; Shannon, 1988). Light-gray, light pinkish-gray, and purplish-gray, strongly flow-layered alkalic rhyolite flow forms a prominent outcrop at Sugarloaf and Ruby Mountains (Keller and others, 2004). Layering is moderately to strongly folded. The rhyolite consists of small, sparse phenocrysts of smoky quartz, sanidine, plagioclase, and rare biotite in a dense microcrystalline groundmass; accessory minerals are apatite, zircon, magnetite, garnet, and topaz. Locally, it contains spherules as much as 10 cm in diameter as well as lithophysae as much as 6 cm in diameter. Lithophysal cavities locally contain euhedral crystals of spessartine garnet and topaz, some of gem quality (Voynick, 1994). The rhyolite contains a basal glassy vitrophyre. Three $^{40}Ar/^{39}Ar$ corrected ages on sanidine at Ruby Mountain are 30.47±0.08 29.25±0.21Ma, and 30.74±0.08 Ma (McIntosh and Chapin, 2004), indicating early Oligocene age. Similarity of age and composition (for example, presence of accessory topaz) with the evolved Mount Antero leucogranites suite of the Sawatch Range suggests a genetic connection (Shannon, 1988). On nearby Bald Mountain, similar rhyolite is estimated to be about 150 m thick and was deposited within a paleovalley that also contains remnants of older Wall Mountain Tuff and Tallahassee Creek Conglomerate (Keller and others, 2004).

Return to Highway 24 and proceed south about 2 miles to the turnoff to Mt. Princeton Hot Springs and Chalk Creek (right turn on County Road 162). Drive to hot springs parking lot on right at intersection with Road 321.

Stop 10. View into Chalk Creek.

Chalk Creek is named for the chalk-like appearance of the cliffs above the creek due to zeolitic and argillic alteration of the Mount Princeton granite, induced by the thermal activity in the area. There have been a number of proposals to exploit the geothermal resources for electrical power production (e.g., Fred Henderson of Mt/ Princeton Geothermal, LLC), but so far the geothermal resources have only been exploited recreationally.

Drive a few 100 yards up Rd 321 to service road for the Frontier Ranch (run by Young Life). Drive up about a mile and park before some ranch buildings (one under construction).

Stop 11. Large fault surface in the Mount Princeton granite at Frontier Ranch.

The entire upper Arkansas River valley is generally west-tilted along a series of down-to-the-east normal faults, including the large Sawatch Range fault zone along the east base of the range. In the Frontier Ranch, we will visit a spectacular normal fault surface exposed in the Mount Princeton granite that contains fault gouge and slickenlines.

The Mount Princeton granite is a major unit in the Sawatch Range and several facies of the granite were described by Shannon (1988), although the predominant facies, exposed here (where it is not brecciated), is an equigranular, medium- to coarse-grained biotite-hornblende monzogranite. Note the prominent honey-colored sphene crystals. The age of the Mount Princeton granite is somewhat controversial. Mills and Coleman (2013) report $^{206}Pb/^{238}U$ zircon ages of 35.80±0.10 Ma and 35.37±0.10 Ma, while McCalpin and Shannon (2005) report an average of 21 K/Ar and fission-track ages of 36.6±0.4 Ma (table 2). McIntosh and Chapin (2004), averaging nine $^{40}Ar/^{39}Ar$ age determinations, report a corrected weighted mean age of 34.75±0.21 Ma. The older U-Pb zircon age is compatible with the age for the 37-Ma Wall Mountain Tuff, which Epis and Chapin (1974) first suggested is the outflow tuff from a long-eroded caldera in the Mount Princeton region. One possible explanation for the slightly younger dates for rocks of the Mount Princeton batholith as compared to the Wall Mountain Tuff is that the 34.8 Ma emplacement of the Mount Aetna caldera, which was superimposed in the Mount Princeton batholith (and which is the source for the Badger Creek Tuff), caused slight argon loss in the rocks of the Mount Princeton batholith and reset the $^{40}Ar/^{39}Ar$ age (Shannon and McCalpin, 2006). Another factor is that the Mount Princeton batholith cooled much more slowly than the Wall Mountain Tuff and therefore might be expected to have a younger isotopic age.

Return to Chalk Creek Road (County 162) and go about ½ mile east to County Rd. 270; turn right. Drive about 3 miles past sharp right turn and stop by gravel pit.

Stop 12. Pre Bull Lake outwash gravel

This deposit (our Qg2) consists of boulder and cobble gravel deposited downslope from pre-Bull Lake till. The gravel is overlain by fine-grained sediment that elsewhere (see Stop 13) contains a water-lain ash interpreted by Izett and others (1988) as the 760 ka Bishop ash (Sarna-Wojcicki and others, 2000); the Bishop ash erupted from the Long Valley caldera in eastern California. Crowley and others (2007) more recently obtained a $^{206}Pb/^{238}U$ age of 767.1 ± 0.9 ka for the Bishop ash.

Continue about 3 miles to Hwy 285 and drive south about one mile south to near pipe over highway.

Stop 13. Bishop Tuff in highway roadcut.

An ash bed, interpreted as the Bishop ash by Izett and others, 1988, occurs in fine-grained sediment at the top of the roadcut that overlies gravel that Scott and others (1975 originally called Nebraskan (?) alluvium; our unit Qg2.

Continue south about 3.5 miles to Hecla Junction; turn left and drive about one mile to just past a mine dump on the right.

Stop 14. Browns Canyon horst and Browns Canyon Formation

The Browns Canyon horst is a rift-related structure that formed concurrently with basin formation and is bounded by normal faults. The rocks of the Browns Canyon Formation sits atop Proterozoic rocks (mostly foliated Elephant Rock Granite) that comprise most of the exposed rocks of the horst. The upper Eocene Wall Mountain Tuff abuts the southwestern side of the horst along the bounding normal fault (see Stop 16 for a description of the Wall Mountain Tuff). The Browns Canyon Formation consist of well-indurated and silicified claystone, siltstone, sandstone, and conglomerate. According to Van Alstine and Cox (1969), the formation locally contains plant fossils similar to late Oligocene fossils preserved in the Creed Formation in the San Juan Mountains, which suggests, if correct, that the formation may represent the basal basin fill that underlies the Dry Union Formation.

However, the geometry suggests an alternative explanation. If Van Alstine and Cox are correct, then the Browns Canyon Formation had to have been emplaced AFTER most movement of the bounding normal faults occurred, which seems unlikely if the formation represents early basin fill. Otherwise, Wall Mountain Tuff most likely would be preserved beneath the Browns Canyon formation, which it is not. We suggest, therefore, that the fossil evidence may be misleading and that the Browns Canyon Formation predates the Wall Mountain Tuff, which was entirely eroded from the footwall block (horst), leaving just scraps of the earlier Browns Canyon Formation overlying the Proterozoic granite. The strong silicification and induration also suggests to us a considerably older age than the Dry Union Formation. Clearly, further study is needed.

The Browns Canyon Formation locally contains fluorite mineralization, which was better developed as fine-grained deposits along the normal faults on the west side of the horst, which juxtapose granite on the northeast with downthrown Wall Mountain tuff (we may visit this relationship nearby). In all, about 130,000 short tons of fluorspar concentrate were mined, largely from the Colorado-American Mine, starting in 1923 and continuing until 1949 (Van Alstine and Cox, 1969).

Return to Hwy 285 and continue south for 2 stops near and within the “Poncha” mountain block that examine features highlighted by recent mapping by USGS geologists Scott Minor, Chris Fridrich, Jonathan Caine, and Cal Ruleman. (The next two stops were added by Scott Minor after the original field guide was created, so are not indicated on the location map)

Stop 14B. Poncha Creek Canyon: Views of Proterozoic basement core and range-front fault of Poncha mountain block

Poncha Mountain Block: Introduction

The main physiographic expressions of the late Oligocene-Quaternary Rio Grande rift in the region of our field trip are two NNW-trending, northward-narrowing basins: the San Luis and Upper Arkansas River Basins. These two basins are separated by a WNW-trending mountain block of Proterozoic and Tertiary rocks that forms the northern tip of the Sangre de Cristo Range.

A change in dip direction of the major range- (basin-) bounding rift normal faults coincides with this distinctive “Poncha” mountain block, from west-dipping faults adjacent the Sangre de Cristo Range (San Luis Basin) to east-dipping faults adjacent the Sawatch Range (Upper Arkansas River Basin). The two Poncha field-trip stops are located in the Poncha mountain block area to observe exposures of: (1) Proterozoic rocks forming the core of the Poncha block that host ductile and brittle structures of various ages; and (2) faulted Miocene to Pleistocene sediments on the northern flank of the block that represent rift-basin fill deposits. Emphasis is on the tectonic history of the Poncha block, including observations that address a hypothesis put forward by earlier investigators that the Poncha block is related to a rift extensional accommodation zone. Prior to our recent U.S. Geological Survey research efforts, little was actually known about the rock types and geometry, kinematics, age, and evolution of structures in this block, or their role in accommodating or transferring rift extension between the basins to the north and south. In this segment of the field trip, we present geologic and structural data that provide new constraints on the tectonic evolution of the northern part of the Poncha block.
Although previously mapped as metamorphosed volcaniclastic rocks, the Proterozoic rocks exposed in the road cut here consist of arkosic metaquartzite interlayered with porphyroblastic sillmanite-muscovite schist. In some places sillmanite porphyroblasts are fist sized. These rocks are intruded by layer parallel, meter- thick amphibolite "sills" that can be seen throughout Poncha Canyon. Poncha Canyon appears to have formed when the axial-planar hinge zone of a possibly Proterozoic or Laramide antiform was "breached" by Poncha Creek. Both quartzite "beds" and amphibolite "sills" are coaxially folded on either side of the canyon
The E-W-striking “Poncha” range-front fault that crosses the canyon just north of the stop location (visible from stop) juxtaposes Miocene basin-fill alluvial sediments of the Dry Union Formation on the north against the Proterozoic rocks on the south. The Poncha fault exhibits dextral-normal oblique slip striae at several localities along the range front, which locally overprint reverse-slip striae, suggesting that at least some fault segments have extensionally reactivated Laramide structures. The Proterozoic rocks were highly faulted prior to extensional development of the Poncha fault, with evidence of a broad range of orientations consistent with both Laramide contractional and rift extensional deformation. Cross-cutting, younger rift faults in the Proterozoic rocks appear to be widely distributed and exhibit strong oblique slip consistent with what is documented in nearby Tertiary and Quaternary sediments. Faults in the Proterozoic rocks locally increase in intensity as the “Poncha" fault is approached from the south. There also appears to have been long-lived (but pre-Miocene?), fault-related hydrothermal alteration with local chloritic alteration overprinted by weak argillic and hematitic alteration adjacent to the range front but also at discrete sites throughout the canyon. Diffuse hydrothermal argillite and hematite, and discrete fluorspar veins are most pronounced at the Poncha Mine located near the southern strand of the range-front fault system about 1 km east-southeast of stop.

Take US highway 285 north back into Poncha Springs and then take US highway 50 east to Salida. In town turn south onto county road 110 and head approximately 1 mile up winding grade to sharp left bend in road and park in turn out for Stop 16.

Stop 14C. Poncha range-front piedmont: Faulted Dry Union Fm. basin-fill deposits

The winding road leading up to Stop 16 passes through basin-fill deposits of the Miocene to Pliocene Dry Union. West- to southwest-tilted alluvial and lacustrine gravels, sands, silts, and

mudstones forming the local Dry Union section underlie the moderately elevated, dissected piedmont located between the Poncha range-front fault on the south and the modern floodplain of the South Arkansas River on the north. Similarly tilted Dry Union sediments and underlying Oligocene intermediate volcanic rocks unconformably overlie Proterozoic rocks along the west and southwest flanks of the Poncha block. Gravels throughout the exposed Dry Union section contain subrounded clasts of mixed volcanic and Proterozoic basement rocks and locally include subangular clasts and house-sized landslide blocks of Paleozoic sedimentary rocks. Present in some areas, however, are monomictic intervals dominated by Proterozoic, volcanic, or Paleozoic clasts. Generally speaking, the entire exposed Dry Union sequence flanking the Poncha mountain block can be viewed as a crude unroofing sequence with respect to the regional stratigraphic section preserved in the nearby mountain ranges. That is, Tertiary volcanic clasts are most abundant or dominant in the lower part of the sequence, Paleozoic sedimentary clasts are locally dominant in the middle part, and Proterozoic basement clasts are dominant in the upper part. Elevated, gently tilted to flat lying, coarse gravels and sedimentary breccias deposits are present on the downthrown sides of the Poncha fault and, west of the Poncha block, the southern Sawatch fault. These sediments, which are mapped as older alluvial deposits and diamicton, uncomformably overlie tilted Dry Union strata in the area. An example of such deposits is exposed a short walk up the road from Stop 16. These deposits contain clasts of almost exclusively subangular, locally derived clasts of Proterozoic rocks; the deposits presumably record a major phase of uplift and erosion of the Poncha mountain block.
Several small- to moderate-displacement faults cut Dry Union strata and, to a lesser degree, overlying alluvial deposits in the range-front piedmont area. One such fault, juxtaposing Dry Union and older alluvial deposits, is nicely exposed at Stop 16. Other smaller faults cutting Dry Union strata are also exposed in nearby road cuts. We collected numerous kinematic measurements (strike, dip, rake, slip sense, and separation) of faults cutting Dry Union strata along the north flank of the mountain block. The measurements indicate that late Neogene and Pleistocene faulting in the piedmont area was characterized normal-oblique slip compatible with coeval normal-dextral movement on the Poncha fault. Faults that cut Pleistocene and late Neogene sediments exhibit kinematic similarities, suggesting that much or all of the faulting is young, or that the strain field has remained relatively unchanged since as early as the middle Miocene.

Return to Highway 285 and drive north to Johnson Village; turn east on Hwy 285 (also Hwy 24). Drive about 5 miles to a roadside stop.

Stop 15. View of Triad Ridge

The units underlying Triad Ridge occupy the pre-rift Trout Creek paleovalley, now topographically inverted along the crest of Triad Ridge, here underlain mostly by the 210-m-thick tuff of Triad Ridge (former tuff of Castle Rock Gulch of Wallace and Keller, 2003, renamed by McIntosh and Chapin, 2004). The ridge is topped by the 36.61±0.10 Ma (McIntosh and Chapin, 2004) capping andesite of Triad Ridge. The tuff of Triad Ridge has an upper and lower member. The $^{40}Ar/^{39}Ar$ age for the lower member is 37.96±0.22 Ma on both biotite and hornblende; the upper tuff of Triad Ridge has a $^{40}Ar/^{39}Ar$ age on sanidine of 36.69±0.32 Ma (McIntosh and Chapin, 2004), which is slightly but significantly younger than the Wall Mountain Tuff (37.25±0.10 Ma (Zimmerer and McIntosh, 2012), implying that the Wall Mountain Tuff may be stratigraphically interlayered with the tuff of Triad Ridge. The Tuff of Triad Ridge underlies Wall Mountain Tuff just to the east of here (next stop), indicating that the tuff of Triad Ridge at that locality is the lower member. Source of tuff of Triad Ridge is unknown (a caldera buried under the Arkansas Valley fill??).

Continue about another mile to County road 307; turn right. Drive about 2 miles and turn right at north end of the prominent "Castles," which are formed in Wall Mountain Tuff.

Stop 16. Wall Mountain Tuff.

The Wall Mountain tuff is a major ignimbrite that Epis and Chapin (1974) first suggested originated from a deeply eroded caldera that once overlay the Mount Princeton granite, although the age of the Wall Mountain tuff is significantly older than that for the Mount Princeton batholith (see discussion of ages at today's Stop 11). The tuff is mostly light brownish gray, crystal rich, moderately to densely welded rhyolitic ash-flow tuff with abundant phenocrysts of sanidine and plagioclase and sparse biotite. The tuff is a simple cooling unit with common eutaxitic texture and locally developed laminar flow foliation. It occupies the Trout Creek paleovalley and overlies the tuff of Triad Ridge (lower member?) The mean $^{40}Ar/^{39}Ar$ sanidine age from 5 samples is 37.25±0.10 Ma (Zimmerer and McIntosh, 2012). Elsewhere, it contains a black, densely welded basal vitrophyre as thick as 6 m (Van Alstine, 1974).

Return to County road 307 and continue northeast until it intersects Highway 285; turn right. Although we won't stop here, note the prominent outcrop of Early Ordovician Manitou Limestone overlying Mesoproterozoic Elephant Rock Granite (the Late Cambrian Sawatch Sandstone was either never deposited or was eroded before deposition of the Manitou Limestone), in turn overlain by Middle Ordovician Harding Sandstone and overlying, dark, cliff-forming Late and Middle Ordovician Fremont Dolomite. Continue over Trout Creek Pass for a brief stop to view the Buffalo Peaks to the north.

Stop 17. View of the Buffalo Peaks

The Buffalo Peaks consist of a layered sequence of late Eocene andesite flows and volcanic breccia, air-fall and ash-flow dacitic tuffs, and lahars (volcanic mudflows). An upper tuff unit has a $^{40}Ar/^{39}Ar$ date on hornblende of 38.18±32 Ma (McIntosh and Chapin, 2004). A small andesite body southwest of the Buffalo Peaks is a possible feeder source for the andesite of the peaks (Kellogg and others (2017).

Continue on Hwy 285 back to Denver.

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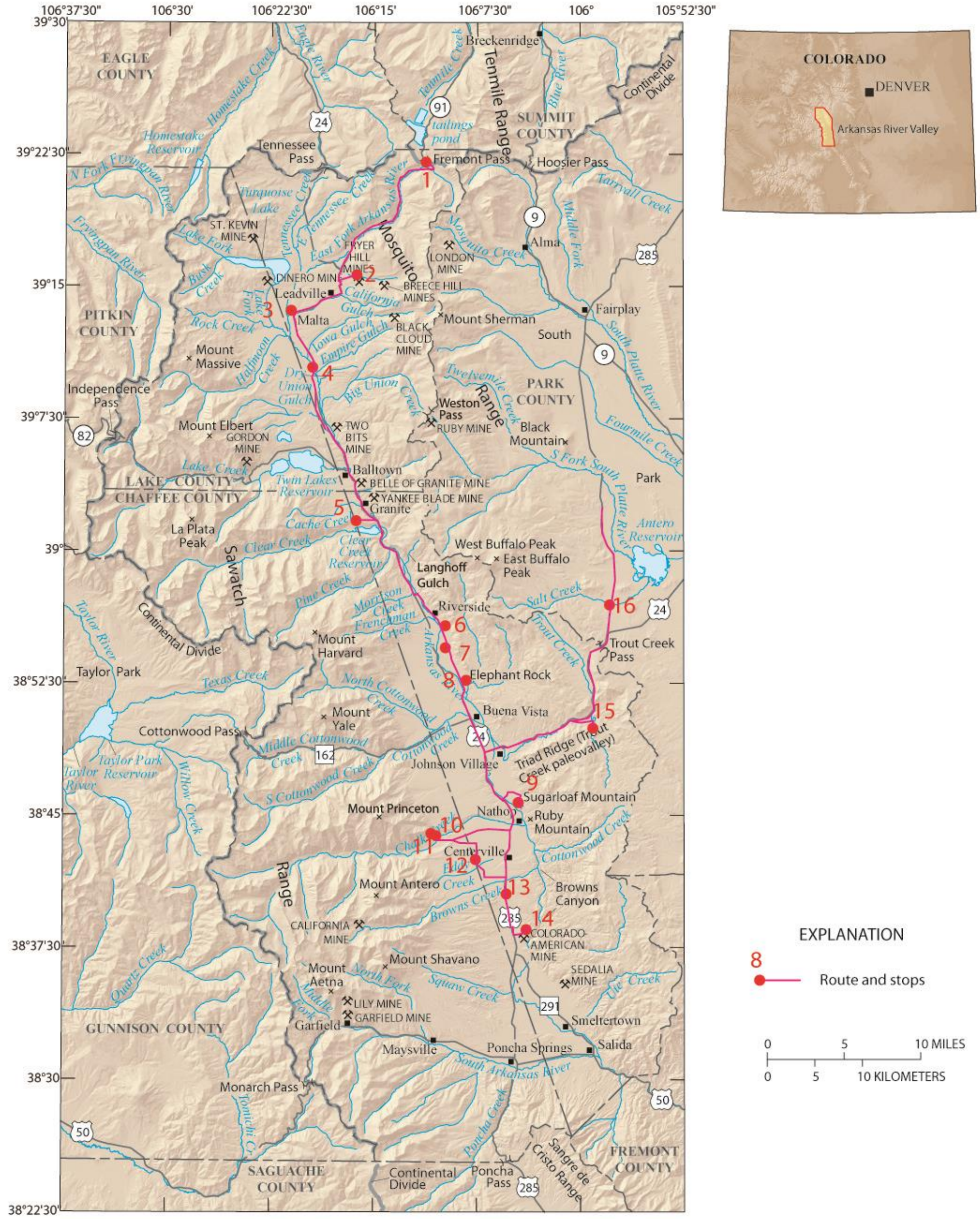

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