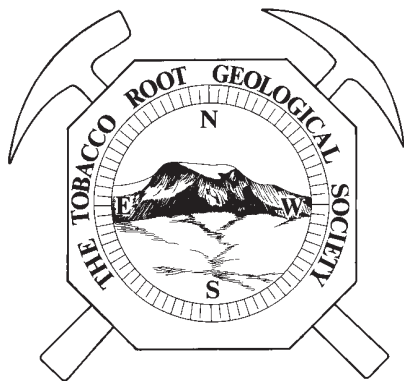




NORTHWEST GEOLOGY

The Journal of The Tobacco Root Geological Society

Volume 54, July 2025

50th Annual Field Conference

Bannack, Montana

July 24–27, 2025



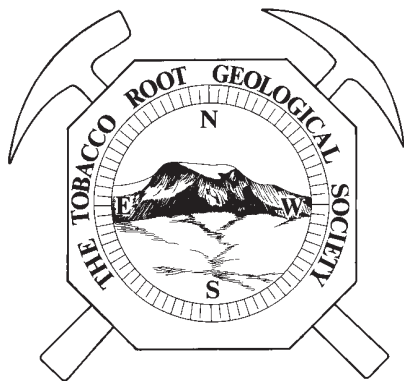
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Edited by: Colleen Elliott and Alan English



The Tobacco Root Geological Society, Inc.

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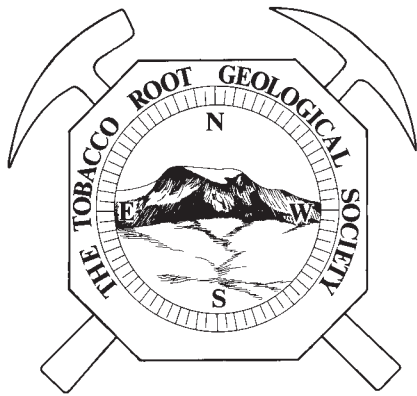
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*Awarded for distinguished achievement
in the study of the geology of the
Northern Rocky Mountains*

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2004: Dean Kleinkopf*
2009: Betty Skipp
2010: Jim Sears
2011: John Childs
2012: J. Michael O'Neill
2013: Paul Karl Link
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2011: Rob Thomas
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2012: Mitch Reynolds
2013: Reed Lewis
2015: Don Winston*
2020: Ted Antonioli, Larry Smith,
Mike Stickney



TRGS 2025 SCHOLARSHIP AWARD WINNERS

- Travis Fisher (Foster), University of Nevada
Taylor Mining District project
- Phoebe Fletcher (Harrison), University of Wyoming
Distribution of rare earth element (REE) enrichment in the Halleck Creek deposit of the Red Mountain Pluton, southeastern Wyoming
- J. Bel Liscomb (TRGS), Montana State University
Geology of the Elkhorn Mountain volcanic field, Montana: Understanding the onset of late Cretaceous magmatism, Wyoming
- Megan Melgren (TRGS), Kansas State University
Geochemistry of off-axis Snake River Plain basalts and implications for the areal extent of Yellowstone hotspot volcanism
- Olson Hunter (TRGS), Stanford University
Montana's window into the Cambrian explosion: Environmental controls of Cambrian soft bodied preservation
- Samantha Malavarca (Skipp), Colorado State University
Magmatism and early rift deformation in the northern Rio Grande Rift, Cotton Creek, Colorado
- Shayla Triantafyllou (Kleinkopf), Colorado State University
Large wood induced pool scour and implications for river restoration
- Liv Wheeler (TRGS), Montana State University
Tuff love: The cyclic relationship between Yellowstone's Mount Jackson Rhyolite series and Lava Creek Tuff



MEMORIAL

Hugh Dresser, 1930–2023

by Phyllis Hargrave

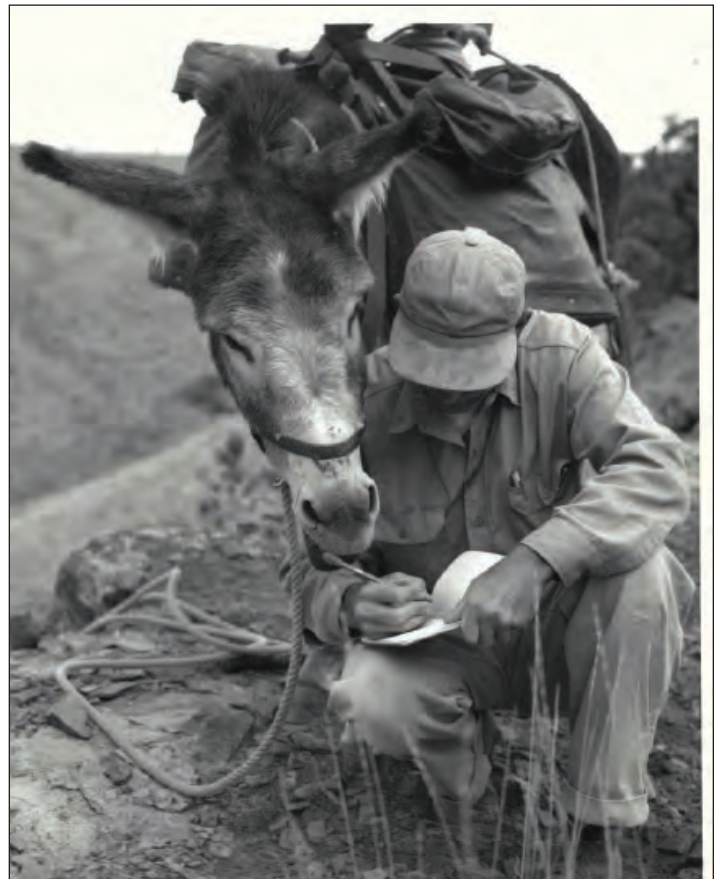


On July 14, 2023, Hugh Dresser passed away, after packing a whole lot of living into his 93 years. To say that he was a renaissance man would be truncating all his interests and accomplishments. Hugh's early education included a BS and MS from the University of Cincinnati and a Ph.D. from the University of Wyoming. He and his soul-mate Joyce moved from Cincinnati to Wyoming, where he worked for Carter Oil Company (now Exxon). His family included their children Kristi and Doug, grandkids, and many cats and dogs through the years. Junior, his donkey, was his trusted field partner while mapping the Muddy Sandstone looking for petroleum reservoirs. In 1965, he joined the faculty at Montana Tech in Butte. Somewhere along the way, he became known as "Doc Rock," because students considered him hard as a rock. Take it as a label of derision or respect, it fit him. He had his standards. With his steely blue eyes, he would simply ask students,

"What do you see there?" He taught many of his classes in 3-D, projecting two slides as stereo images viewed with polarized glasses. There again, love it or hate it, students got to use their minds with this unique way of seeing geology in the classroom. Years later, former students would admit they learned many things in Hugh's classes that helped them in their careers.

His love of geology and photography led him to buying a two-seater Cessna 150 (sort of the VW Bug of planes) to visualize geology, especially structures, from above. Hugh generously took students and faculty, at no charge, for flights around Montana. He also instilled the importance of boots on the ground geology—following contacts, structures, and lithologies was a requirement. Hugh once said, "Well, I lose a lot of students after field trips"—meaning they dropped out and never came back. For years, he had an all-day Sunday field course visiting a variety of sites around southwest Montana. One of the great things about this immersion in geology was that incoming students became oriented to new and representative places they could explore again and again on their own. Trips to Sandy Hollow, Dougherty Mountain, Lost Creek, and the Jefferson Valley were famous for their long and intense days.

Hugh had a vast collection of stereo-paired slides and photographs, with hand-drawn interpretations of structure and lithology, that he used to print a series of pictorial guides on many general subjects such as volcanoes, rivers and streams, pingos, landslides, folds, pediments, faults.... the list goes on. He was not constrained to just studying structural geology, and was intrigued by volcanics and intrusives as well. Belt rocks did not deter him and volcanics let his imagination roam.





Hugh was always eager to explore new terranes and was a willing field partner. His approach to geology was to carefully trace contacts, collect structural and lithologic data, and then come up with a theory, as opposed to following a preconceived model. From a personal perspective, Hugh advised myself and my husband Dan Truckle for our Master's degrees. One Saturday, he and I set out to explore the Lowland Creek volcanics in Browns Gulch, west of Butte, promising his Joyce that we would be back early. About 6:00 pm, we rolled in to Butte and I asked Joyce, "Are we in trouble?" She said, "I've been married to Hugh a long time. Dinner is on the table. Sit down and eat."

It would be a disservice not to mention his skills as a photographer (often printing his own photographs and having students print air photos to use in class), world traveler, stock analyst, carpenter, cook, winemaker (okay, not so much), and animal lover (after retiring, he volunteered at the local animal shelter in Butte, and fed the ravens west of Butte daily). Hugh was a person to respect and hopefully imbued just tiny bit of his life into ours as geologists.



TRGS HAMMER AWARD

Mike Stickney

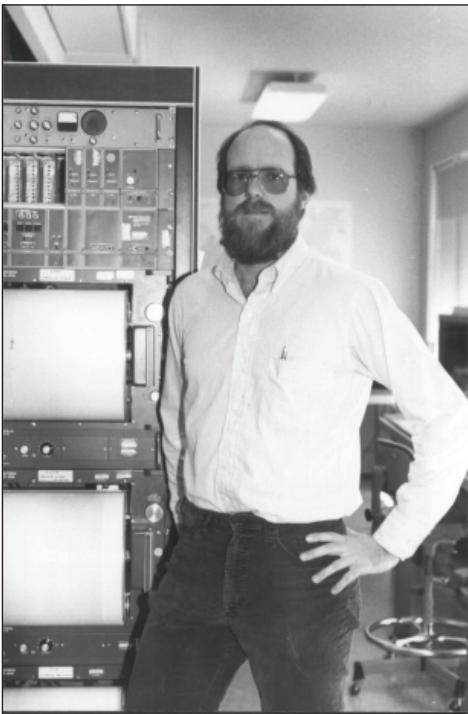


TRGS is pleased to honor Mike Stickney with the 2025 Hammer Award. Mike earned his B.S. in Geology from the University of Montana in 1978, and in short order, his M.S. in Geology from the University of Montana in 1980. He received his Ph.D in Earth Sciences (honorary), from Montana State University in 2004.

Mike has been involved with seismic monitoring since 1973, first as a student at the University of Montana, later as a visiting scientist at the U.S. Geological Survey in Menlo Park, CA, and then with the MBMG beginning in 1980. He established and operates the Montana Regional Seismograph Network and uses the

data it collects to produce a regional earthquake catalog. In addition to seismological studies, Mike has mapped Quaternary faults and related geology, participated in paleoseismological studies, and provides information on seismic hazards in Montana to the public and State and Federal agencies.

Mike has been an enthusiastic supporter of TRGS for many years, serving as President for 9 years (2007–2015), and on the Board of Directors for 10 years (2016–Present). Over his career, Mike has been a longtime contributor to numerous professional publications, with over 60 publications in his quiver, including his many papers in Northwest Geology. As if he had time for anything more, Mike is also an avid downhill and cross-country skier, often sighted on the slopes of Discovery Ski area near Philipsburg.



PAPERS



RECALLING THE FIRST TRGS FIELD CONFERENCE

Richard I. Gibson

Geologist, Butte, Montana

<https://richardigibson.substack.com/>

“And to the young who have no pain, who have not yet kept watch on man’s mortality, nature is a joy responding to their own, haunting them like a passion.”
— G.M.Trevelyan

It took Bob and Marian Lankston and me more than 2 years to plan the first TRGS Field Conference. We were naïve and inexperienced, but with the confidence of youth, together with no idea what hurdles we’d have to cross, we just did whatever seemed to need doing. And the first field conference was, I think, a success. This paper is a photographic memory of that meeting in 1976. A report on the origins of the Society, together with the start of the 1976 meeting, was published in the 1993 Guidebook, and is reprinted here as well, with a few additions.

As has usually been the case since, the first conference was dominated by field trips, five of which are documented in the Guidebook, plus one more. Except for the 1991 guidebook, the 1976 volume, at 165 pages, was the thickest TRGS produced until 2003.

Four of the field trips were based on the seminal PhD work that had recently been completed by the trip leaders, Hanley (1975, Structure and petrology of the northwestern Tobacco Root Mountains, Madison County, Montana: Indiana University), Garihan (1973, Geology and talc deposits of the central Ruby Range, Madison County, Montana: Pennsylvania State University), Hess (1967, Geology of Pre-Beltian rocks in the central and southern Tobacco Root Mountains, with reference to superposed effects of the Laramide-age Tobacco Root Batholith: Indiana University), and Schmidt (1975, An analysis of folding and faulting in the northern Tobacco Root Mountains, southwestern Montana: Indiana University).

Although make and model of vehicles, hair lines, and clothing styles change, many things about geology field trips—lunch stops, field lectures, and so on—are largely timeless, at least over a 50-year timespan. The photos are by Gibson except as noted.



Steve Henderson, Grace Koehler (1948–2015), and Steve Koehler at the registration desk in the Bozeman Holiday Inn. Both Steves are Charter Members of the TRGS, and Steve Koehler contributed a paper to the 1976 Guidebook. Koehler was Vice President, then President of the TRGS from 1978 to 1980; he lives in Grangeville, Idaho. Steve Henderson’s career was with Oxford College, Georgia, and in 2025 he lives in Covington, Georgia. TRGS signage has not changed much in 50 years: the handmade sign simply says, “Tobacco Root Geological Society Registration.”

THE VERY FIRST TRGS FIELD TRIP—WITH HANLEY—INTO THE TOBACCO ROOTS

The first conference started with a 3-day preconference backpacking field trip, led by TRGS Charter Member Tom Hanley into the Archean–Paleoproterozoic rocks of the Tobacco Root Mountains, in the Curly Lake area. We actually engaged a packer, Hubert Piper, a friend of Bob Lankston's family since the 1930s. Hubert's gigantic skillet afforded a cooking space for elegant meals. Marian recalls that Hubert's friend with a mule, scheduled to carry a lot of the supplies, broke his leg. Undaunted, Hubert suggested that "we could pack in everything on our backs," even though it took one round trip by a core group before the trip technically began, plus a second trip with the other attendees. I carried in the champagne we used to celebrate the U.S. bicentennial; the empty bottles seemed just as heavy going out.

Tom Hanley retired in 2006 after a 31-year career teaching at Columbus State University, and in 2025 still lives in Columbus, Georgia.



Tom Hanley arm-waving to John Brady, Grace and Steve Koehler, Dave Hess, and Pete Norbeck.



Steve Koehler playing around the sign on the South Boulder Road, with the addition "TRGS Ahead!"



Champagne and "Oly" in the Curly Lake refrigerator, July 4, 1976.



Tom Hanley and Steve Koehler examining a mafic dike cutting quartzofeldspathic gneiss.



Hubert Piper, Dick Gibson, and Bob Lankston take a break on the Curly Lake hike-in. Photo by Marian Lankston.



Bob Lankston, Dick Gibson, Tom Hanley, Steve and Grace Koehler (left to right), along South Boulder Road before the start of the Curly Lake hike. Photo by Marian Lankston.



High country in the Tobacco Roots, Grace and Steve Koehler for scale.

SPECIMEN CREEK PETRIFIED FOREST TRIP

We were sufficiently organized to charter a bus for the trip to the Specimen Creek (Gallatin) petrified forest in Yellowstone, led by John Montagne (1920–2008), Bob Chadwick (1929–1985), and Lanny Fisk (1944–2018). The bus allowed us to haul a full-fledged portable chalkboard for on-the-road discussions, although we also used the traditional display space on the side of the vehicle.

Bob Chadwick served the Society as Vice President in 1985 until his death that year. Lanny Fisk, a Charter Member, was Vice President, Secretary, and a Board member over the time from 1979 to 1991. John Montagne co-founded the geology department at Montana State (with Charles Bradley) and worked there until his retirement in 1982.



The bus as backboard.



Bob Chadwick and John Montagne explain Gallatin Canyon geology. Photo by Marian Lankston.



Through the reverse-graded lahar conglomerate.



Lunch at Specimen Creek trailhead.



Lanny Fisk.



Petrified tree, Gallatin Petrified Forest.



GARIHAN TRIP

Jack Garihan's field trip to the Precambrian of the central Ruby Range and its talc deposits was based on his thesis work. In 2025, he observed, "Looking at those basic 1976 descriptions reminds me now what I thought I knew then. Well, any thesis is really just a 'progress report.' Detailed first-pass geologic maps stand as the basis for meaningful subsequent research." Every mapper knows this!

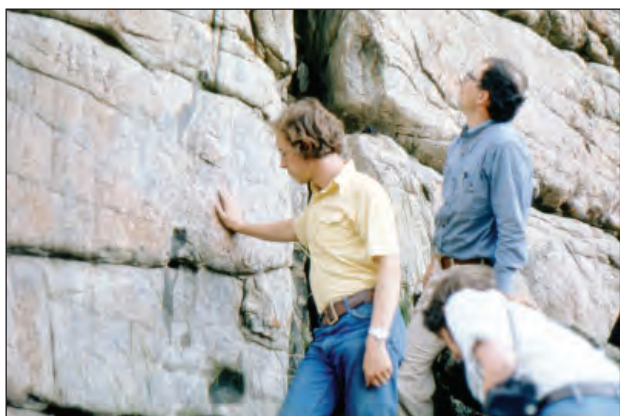
Jack taught at the Indiana University Geologic Field Station from 1974 to 1990, and in 2025 is an Emeritus Professor at Furman University, Greenville, South Carolina.



Jack Garihan discusses Ruby Range geology.



Charter Member Joe Liteheiser with the caravan on the Ruby Range trip. Joe went on to serve as the Secretary of the Seismological Society of America from 1987 to 2007. Photo by Marian Lankston.



John Brady and Tom Hanley examine the outcrop.



Standard TRGS field trip activity.



At the Treasure Chest Talc Mine.



Another lecture on the outcrop.

HESS TRIP

Dave Hess led a trip to the Norris–McAllister area on the east side of the Tobacco Root Mountains, to examine both the Precambrian metamorphic rocks and the much younger igneous rocks of the Tobacco Root Batholith and Tertiary volcanics.

Dave retired in 2000 after a long career at Western Illinois University. In 2025 he lives in Macomb, Illinois.



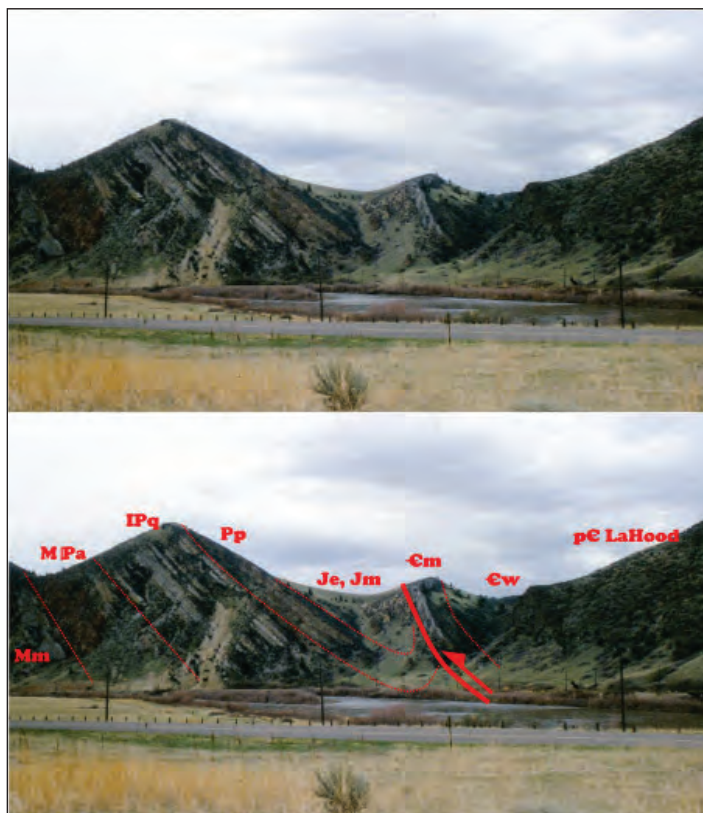
Multiply deformed Precambrian rocks in the outcrop near the crest of Norris Hill, a stop on the Hess trip. Modern photo by Dick Gibson.

HAWLEY–SCHMIDT TRIP

Dave Hawley (1914–1985) and Chris Schmidt led a trip to examine outcrops of the Precambrian Belt Supergroup La Hood Formation, especially in Jefferson Canyon, combining Hawley’s expertise on turbidites and similar rocks with Schmidt’s tectonic and structural analysis of the northern Tobacco Roots.

Perhaps the most iconic feature visited was the Cave Fault exposed on the west side of the Jefferson River south of La Hood Park, near Cardwell, where the La Hood, Cambrian Wolsey, and Cambrian Meagher Formations are thrust onto the overturned limb of a syncline cored in the Jurassic Morrison Formation. The Flathead is absent there as a consequence of nondeposition on Middle Cambrian islands of La Hood strata.

Hawley’s career included work in the geology department at Hamilton College, Clinton, New York, from 1954 to 1980, with several summers at the Indiana Geologic Field Station teaching and conducting research. Schmidt, long associated with Western Michigan University, served as faculty at IUGFS various years between 1975 and 1985 and has a home now in the South Boulder Valley.



Cave Fault, Jefferson Canyon. 2019 photo and annotation by Dick Gibson.



SUTTNER-STRAW TRIP

There's no road log in the Guidebook, but an extra field trip was led by IUGFS professors Lee Suttner and Tom Straw. While to an extent it was intended as a reminiscence for former students, for newcomers it also gave a great overview of the geology in the northern Tobacco Roots. That geology and the camaraderie of the IU Field Course excited me and Bob and Marian and led directly to the creation of the TRGS.

The Suttner-Straw trip visited classic locations where IUGFS students learn and map, including the South Boulder section where exposures of the Archean-Paleoproterozoic crystalline rocks to the Cretaceous Elkhorn Mountains Volcanics are visible and accessible, and represent the basement-involved "thick-skinned" deformation style of the Laramide-Sevier Orogeny. Only Ordovician, Silurian, and Triassic strata are missing.

We also visited the complex structure along today's Montana Highway 2 (referred to by IUGFS as the "Highway 10 problem" for decades after old US 10 had been designated Montana Highway 2), between Lewis and Clark Caverns and Sappington Junction, where students try to decipher folded and faulted strata, mostly in the section above the Mississippian Mission Canyon. There, the structure reflects the "thin-skinned" deformation dominated by thrust faults and their consequences. It was the completeness of the stratigraphic section, together with the contrasting styles of deformation, that led Charles Deiss in the late 1940s to press Indiana University to establish its geologic field station on 60 acres at the confluence of Carmichael Creek and the South Boulder River, erected in 1949 on land that cost \$5 per acre.

In 2025, Lee Suttner is retired in Bloomington after a long career with Indiana University, and Tom Straw lives in New Harmony, Indiana, after working for decades at Western Michigan University.



Lee Suttner points northwest along the Carmichael Fault. Indiana University Geologic Field Station in the background.



Lee Suttner (pointing) and Tom Straw (with charts) in the Lodgepole Formation on Carmichael Mountain. Charter Member Marv Miller looking into the camera at left center. Marv was TRGS Vice President (1976–1977) and second President (1978). His career included teaching at the IU Field Station in 1963–1967. He is now in Butte, retired from the Montana Bureau of Mines and Geology, where his 45-year career included a stint as Assistant Director.



Lee Suttner and Tom Straw explain the fold-thrust relationships in the Highway 2 map area.



In Jefferson Canyon, La Hood Formation in the background.

TRGS BANQUET

The culminating “banquet” in 1976, now traditional for the TRGS, was a dinner at the Indiana University Geologic Field Station, the literal origin of the Society. IUGFS Director and TRGS Charter Member Judson Mead (1917–2010) welcomed about 50 TRGS attendees following the Suttner–Straw trip.

Among the 65 students in the IU field course when the TRGS was there was Tekla Harms, a student at Bryn Mawr who is now at Amherst College, with a long career studying the geology and tectonics of southwestern Montana, among many other things. Another student that summer was Mike Stickney from the University of Montana. About 4 years later, Mike began his career with the Montana Bureau of Mines and Geology, where he is still the seismologist in 2025. Mike was TRGS President from 2007 to 2015, shepherding nine Field Conferences, and has served on the Board of Directors since 2016. Sixty-five students, plus faculty and staff and researchers, meant that the dining area of the lodge at the Field Station was full, so the TRGS attendees had a “second supper” after the students had finished.



Jud Mead welcomes TRGS to a cocktail party in the classroom at the Indiana University Geologic Field Station. Jane Mead and Gene Hinton at right.



Tom Hanley, Chris Schmidt, Steve Koehler at dinner.



Charter Members Steve Koehler and Tom Hanley share smiles after dinner.



Charter Members Vicki Miller (1945–2016) and Marv Miller visit with Gene Hinton (1922–2005), longtime resident manager at the Indiana Field Station.



The TRGS group took up six or seven tables for dinner at the IUGFS.



Charter Members Tom Hendrix (1933–2023; at left, serving) and Judson Mead. Hendrix contributed a paper to the 1976 Guidebook and led a TRGS field trip to Sandy Hollow in 2007.



TRGS President Dick Gibson was a little excited to be back at the Field Station and couldn't get enough pictures. Photo by Marian Lankston.



ORIGINS OF THE TOBACCO ROOT GEOLOGICAL SOCIETY

by Richard I. Gibson (with contributions from Bob & Marian Lankston)

"I am in love with Montana." John Steinbeck wrote that in *Travels with Charley*, but it nicely sums up why the Tobacco Root Geological Society was formed. Many if not most of the students who took Indiana University's course, G429, *Field Geology in the Rocky Mountains*, based at the IU Field Station in the Tobacco Roots, went back to Indiana with an intense desire to return to Montana. Bob Lankston and Marian Millen, field course alumni from 1968, and I (1969) were but three of many geology majors who filled many winter hours in the Geology Club Room, on the second floor of the Geology Building in Bloomington, with reminiscences of Montana. We were much more concerned about reunions in Montana than about jobs, and by some time in 1970 the date and place of a "Grand Reunion" were set: July 4, 1976, in Ennis. The Bicentennial date would be unforgettable, and in those days students in the field course typically went to the Ennis Rodeo on the Fourth of July. There was nothing particularly serious about all this—just the dreams of a bunch of idealistic '60s students.

The summer of 1969 at the Field Station cemented friendships between Bob and Marian and me that had begun in the previous school years. Bob Lankston and Marian Millen were field assistants that summer for Gregg Vane, who was working on a geophysics Master's thesis based at the Field Station. I joined them for the remainder of the summer after I finished the field course. Bob and Marian were married the following fall, and my friendship with them both has continued to the present.



Gregg Vane, Marian Millen, and Bob Lankston at the "Crater of the Clods," the result of an explosion during a seismic survey of the Jefferson Basin, near Waterloo, Montana, August 1969.

Bob and Marian moved to Missoula in 1971, where Bob pursued his Ph.D. in geophysics. I went to Davis, California, in 1972, to analyze kidney stones. A critical event in the history of the TRGS was Bob and Marian's inviting me to Missoula for Thanksgiving 1973. I got there despite blizzards in Nevada, and after a traditional Thanksgiving dinner, the conversation turned (not surprisingly) to the dream of the "Grand Reunion." This time there was a difference—we decided to DO IT. In my mind, the TRGS came into being, at least conceptually, that November day in 1973.



Bob and Marian preparing to rappel in Blackfoot Canyon near Missoula, Thanksgiving 1973.

Originally, the TRGS was conceived literally as a party. We figured we'd invite all the guesstimated 1,500 alumni of the Indiana field course, and we assumed (with youthful optimism) that at least 300 would show up. Our grandiose plans focused on renting the Ennis Grange Hall (or whatever we could find) and reserving enough rooms to house all these partygoers. We learned quickly that you couldn't possibly house 300 visitors in Ennis (population then about 600), even if you counted the 60-odd campsites in the area. The good news was that we also began to realize that 50 might be a more rational estimate for attendance—and that included the 3 of us and 12 friends who swore they would show up.

Over the winter of 1973–1974, the idea that it might be more fun (and maybe even easier) to form a professional society, rather than sponsor a one-shot party, developed and grew. Bob and Marian paid a reciprocal visit to me in California, and we drove our blue and green (respectively) Volkswagen Beetles to Death Valley for a geo-vacation in March 1974.

After a night of sand-blasting storms at Stovepipe Wells Campground, a scorpion under Marian's sleeping bag, and a Styrofoam cooler punched out by two somewhat inebriated male geologists, we retired to the Oasis Lounge at the Furnace Creek Inn for a few cold drinks. Throughout this Death Valley vacation, we'd been casually discussing plans for 1976, but the meeting in the Oasis Lounge March



Campsite for the formative TRGS meeting in 1974.

21, 1974, was effectively the formative meeting of the TRGS. Marian took notes, so it must have been official. We celebrated after the meeting with a bottle of Mateus Rosé, and in 2025, I still had the cork from that bottle.



Although faded, the hand-written inscription says “w/Lankstons, Mar 21-74.”

We had been calling our proposed group the “Indiana University G429 Alumni Association,” which is a mouthful to say and has no good acronym. So a primary goal of the Oasis Lounge meeting was to settle on a new name. Dozens of ideas were entertained: Pacific Northwest Geological Society, Southwest Montana Association of Geologists, Indiana Geologists of Montana, Upper Missouri River Geologists, Ennis Geological Society, Northwestern Association of Geologists, Northwest US Geological Society, and all possible variations. None had hooked us all, until Marian said, “Tobacco Root Geological Society.” I think we all, almost immediately, liked it; it was easy to pronounce, and the acronym, TRGS, was distinctive and not a tongue-twister. More importantly, it reflected the “roots” of the Society in the range that holds the Indiana Field Station, yet did not exclude non-Indiana people. We realized from the start that there might be a problem in conveying that the Society would support studies beyond just the Tobacco Roots, so the Society, in its Articles of Incorporation, was subtitled “Promoting the study of the geology of the Northern Rocky Mountain Province.” Before Bob and Marian left Montana

in August 1974 (for Gulf Oil and a 2-year stint in Houston), a first draft of the TRGS constitution had been generated.

At the Lankstons’ urging, I too went to Houston in February 1975, to change my career from kidney stone mineralogy to oil industry aeromagnetics. Being together again allowed us to get to work seriously on the TRGS and plans for 1976. One of the most “taxing” efforts was in obtaining non-profit status for the Society. This introduction to IRS language, sub-paragraphs, and incomprehensibility has actually been beneficial to me over the years, as my own tax picture has become more cloudy, but at the time it was quite intimidating. After we filed the initial forms, I got a call (at work) from the IRS with a lengthy list of questions we hadn’t even considered: Who will own the copyrights? (We were not yet to the point of worrying about publications.) Would interest income be distributed to the shareholders (what shareholders?) or reinvested, or distributed in accordance with Section 501 (c) (3), paragraphs ii to iii? And by the way, what interest income? So far, our income was zero and the total expenses were for a gavel, bought by the Lankstons, and a copy of *Robert’s Rules of Order (Revised)*, paid for by me. Somehow, we passed the scrutiny of the IRS and non-profit status was granted eventually. Legally, the society was incorporated in early 1977, but official non-profit recognition from the IRS did not come until mid-1981.

Initial operating expenses were paid for out of our pockets, but as we got more organized, we solicited Charter Memberships at \$25 each. There were 23 Charter Members, and their dues paid for letterhead stationery, postage, and a few phone calls in connection with the 1976 Field Conference, as well as filing fees for incorporation.

In part to promote charter memberships among our “rich” oil industry friends in Houston and in part to give early Charter Members something fun to do, the very first Tobacco Root Geological Society field trip took place March 22, 1975, in southeastern Texas. Charter Member



TRGS field trip at Blaisdell Point Bar, Brazos River, Texas, March 22, 1975. Charter Member Pat Price at center.



Patricia Price, who was with Gulf Oil (and who, I think, never set foot in Montana in her life), led the group to beach and fluvial sedimentation sites in the Galveston area and along the Brazos River. Other Charter Members who participated in that trip included Stan Anderson, Clyde Cody, Steve Henderson, Ray Rene, Bob and Marian, and me.

We were busily drafting and re-drafting the by-laws for the Society at this time, and most of our “official” meetings related to this. These get-togethers, usually at Bob and Marian’s apartment on Bissonnet Street in southwest Houston, were highlighted by “the chair recognizing the floor,” or the sofa, or wherever the other person happened to be sitting. We found nothing in *Robert’s Rules* that precluded making a business meeting a fun event! For organizational and necessary legal purposes, we designated ourselves the officers of the Society, with me as President, Bob as Vice-President, and Marian as Recording Secretary-Treasurer. In the first election at the 1976 conference, Bob was elected to the vacant position of Corresponding Secretary and Marv Miller was elected Vice-President, while Marian and I were confirmed in our positions. The first field conference pretty much paid for itself and broke even financially.

Marian designed and drafted the original logo in 1975, based on photos of Hollowtop Mountain taken by Marian and by me, and on the layout of a Brunton compass face. The color logo based on this was designed by Brian Collins. The first spin-off logo product was green T-shirts with the logo, but they didn’t appear until 1980. I had them made in Houston.



Original TRGS logo.

Plans were proceeding for the July 1976 meeting, now scheduled for Bozeman because of lack of accommodations in Ennis. We contacted people we knew to write papers and lead field trips. Through them, and others they led us to, the first publication and field conference were born.

We thought we were home free when the Montana Bureau of Mines & Geology agreed to publish the Guidebook, ultimately a 165-page collection of papers and road logs, MBMG Special Publication 73. However, the agreement

stipulated that we would do all the editing. As it turned out, all three of us reviewed and edited all of the papers (pretty much a new experience), and Marian redrafted a number of the original submitted figures and retyped some of the manuscripts. We were, not surprisingly, somewhat late in getting copy to the Bureau, and consequently, the Bureau was somewhat late in getting the books printed and delivered to us. In fact, the publication was delivered to me at the Bozeman Holiday Inn at 2:30 p.m. on the day that the meeting started, with registration (and book distribution) to begin at 3:00 p.m.

That afternoon was made more complicated by the fact that I got lost hiking out of the Curly Creek drainage in the Tobacco Roots, from the pre-meeting hiking/camping field trip led by Tom Hanley. I was 5 hours late to Bozeman, arriving 20 minutes ahead of the Guidebooks. Buster LaMoure, a University of Montana and I.U. field station alumnus, and Chauncey Avrel, an octogenarian friend of Bob and Marian’s from Missoula, who happened to be in the lobby, were designated as receptionists while I took a shower and dug up the registration packets and got the hotel organized. That day also happened to be the only time in the history of the Indiana Field Station that a student was lost overnight.



Commemorative mug commissioned by Marian, produced by Clays in Calico pottery in Cardwell, using clay from the Grayson Shale. We think at most five of these were made.

After that, the 1976 Field Conference went off basically without a hitch, if you discount a few humorous typos in the Guidebook, such as the “erosion surfaces, which may be traced southward to Corres Pounding,” presumably a village somewhere south of Ennis. There were about 60 registrants for the first meeting, and, like the 1993 meeting, it was marked by a reception and dinner at the Indiana University Geological Field Station. The membership decided at the 1976 meeting to pursue incorporation, which was accomplished in about a year.

Membership in the TRGS in June 1975 was 38, derived from 14 states. By October 1975, the society had 50 members in 20 states, and in January 1976, membership had increased to 65, representing 26 states. After the 1976 Field Conference, membership had grown to almost 100.

Today, you’re a member if you attend the meeting, although certainly some people pay the annual dues when they can’t attend.

Bob and Marian Lankston and I were actively involved in the Society until 1983; Marian’s involvement was a couple years longer, and by 1986 the Board consisted of Liz Brenner, Lanny Fisk, and Dick Berg. I returned as a board member in 2001–2013 and as a guidebook co-editor and layout guy for 2004–2013. Over the years, meetings have been held in many different locations in Montana, as well as in Washington, Idaho, and British Columbia. The TRGS has given more than \$120,000 in scholarships. The Society and its membership and leaders have evolved in very positive ways, and the TRGS remains one of the things in our lives of which we are most proud.

In 2025, Dick Gibson lives in Butte, Montana, and Bob and Marian Lankston live in the Seattle area.



Dick, Marian, and Bob at the 2007 conference. Photo by Mike Stickney.





Robert W. Lankston

Graham, Washington

ABSTRACT

A serendipitous observation in 1984 by a group of geophysics students during a summer field course suggested that one small talc exposure on the natural surface near the then-embryonic Regal Mine near Dillon, in southwestern Montana, had a lower seismic velocity than the surrounding country rocks. A year later, data from seismic uphole measurements on a bench in the mine seemed to confirm the surface seismic observation at depths well below the level of mechanical weathering. Mechanical weathering was apparently not a controlling factor in the velocity anomaly observed at the surface, though it probably had some influence. The interpretation of the surface observation and the later modeling of the uphole data invited the conclusion that seismic methods could be applied to talc exploration, possibly even as a direct detection technique. By the mid-1990s, seismic surveys were being conducted for talc exploration in Australia, apparently based on subsurface velocity relationships similar to those interpreted from the Regal Mine experiments a decade earlier.

INTRODUCTION

The thread of this story begins in 1967. Talc may have been key to the start of my life in the geosciences, and popped up randomly several times through the next 30+ years of my career in geophysics. In 1967, I was an undergraduate hoping to become a geology major. However, I had to get past the rocks and minerals course first, and talc was one of the few minerals that I could routinely identify. I squeaked by the required course with a C, thereby opening the door to taking geoscience courses that were more suited to my computational interests.

Talc tiptoed into my life again one summer day in 1971. William (Bill) Cordua, who would later become a TRGS charter member, was mapping in the southern Tobacco Root Range. On that day, he and I traveled from Ennis, MT, Bill's base of operations that summer, to the Yellowstone talc mine south of Ennis. I suppose that we were given some sort of tour of the mine, but I do not remember much from that outing.

A few more years went by. Marian Lankston, Richard (Dick) Gibson, and I laid the groundwork for the TRGS in early 1974, and we began planning the organization's first field conference, which would be held in 1976. John (Jack) Garihan offered to lead a field trip during the 1976 program, and we accepted his offer. Marian, Dick, and I were on Jack's field trip that crossed the Ruby Mountains from west to east, with a stop at the Treasure Chest talc mine (fig. 1).

THE GEOPHYSICAL FIELD EXERCISE CONNECTION

During the 9 years after the TRGS was formed, I worked in big oil, had a post-doctoral university appointment, and did contract geophysical work along with Marian. In the fall of 1983, I was a new faculty member in the geology department at the University of Arkansas. I was charged with building a program in geophysics at an institution where such a focused program had not previously existed. One aspect of the new program that I visualized would be an annual summertime geophysical field course. I had high hopes for the summer program, which I dubbed the Geophysical Field Exercise. I anticipated that the field geophysics course would tie in various ways to the geology department's summer geologic mapping, "field camp" course, which operated from the campus of the then Western Montana College in Dillon, MT. My hopes for the course were realized to one degree or another through the five summers of the program. However, with the price of oil steadily declining from a peak in 1980 to a low in 1986, and with geology student enrollments following that trend with a lag of 2 or 3 years (Lankston, 2024), the geophysical field course ultimately died from the lack of student interest and funding. We had five good field seasons, nevertheless; four in the Dillon area and one near Whitehall, MT.

The first thing that I needed for the inaugural course in 1984 was a location for the fieldwork. The location had to be small enough for me to be responsive to the students as they worked, yet big enough for groups of two or three students not to be in each other's way as they operated the various geophysical instruments. Marian and I had known Elizabeth (Liz) Brenner since 1977. Liz, of course, was a long-time TRGS member and leader. In the 1980s, Liz lived in Dillon, and I asked her whether she knew of some spot near Dillon that we could use for the Geophysical Field Exercise.

Liz put me in touch with Thomas (Tad) Dale, who was the Mine Superintendent for Pfizer, Inc. In those days, Pfizer, Inc., operated the Treasure complex of talc mines in the Ruby Range and controlled the site of the Regal Mine. I had met Tad in 1982. Tad gave the after-dinner presentation at the TRGS "banquet" at the old hotel in Bannack, MT. After Tad's presentation, a member of the audience inquired about the relative importance of talcum powder in the talc market. Tad's response was, "You can powder a lot of babies' bottoms with fifty pounds of talc." The Pfizer, Inc.,



Figure 1. Jack Garihan (right) describing the talc occurrence at the Treasure Chest mine during the 1976 TRGS Field Conference. As on most field trips, participants show varying degrees of interest in the stop.

mines, of course, produced tons of talc ore every day for a variety of industries other than cosmetics.

Tad suggested that my students could work on the undisturbed ground at what is now the Regal Mine. Actually, the site was called the Regal Mine in 1984, but it was little more than a large sampling trench. No actual mining was underway then. The site had a long history by then, though. The 1961 edition of the Christensen Ranch (MT) 1:24,000-scale topographic quadrangle, which is available through the U.S. Geological Survey historical topographic maps website, shows a mine symbol and the outline of a small pit on the south side of the Sweetwater Road approximately 10.5 mi (17 km) by air generally east of Dillon. The site had apparently been worked at least since the 1940s, with the original mining actually underground (Perry, 1948). Northwest of the small pit or trench was a mostly treeless and sagebrush-free, windswept hilltop with little to no soil cover. That hilltop would be our study area.

My objective for the first offering of the course, which was just 5 days, was to focus on survey design and operation of the instruments, with some basic interpretation in the evenings using a classroom that the college made available to us. I had no intention of attempting to geophysically map any particular geologic feature such as a fault, an igneous intrusion, the water table, etc. Subsequent Geophysical Field Exercise courses, however, were of longer duration

and did focus on specific subsurface targets as my knowledge of the geology of the Dillon area increased and as I began to wonder what we might resolve geophysically that surface mapping alone could not define.

I had no real budget for the 1984 launch of the Geophysical Field Exercise. I planned to do only refraction seismic and total field magnetic profiling during the inaugural field course experiment. I was counting on tutorial help from Marian. I borrowed a 12-channel seismic system and a couple of microcomputers from Marian's company for the course, but I do not recall the source of the magnetometer that we used.

Tad gave us a detailed topographic map of the site based on the mining company's own photogrammetry. With the map as a guide, we staked out a grid for our geophysical observations. While Marian and Liz, who graciously lent her time to the Field Exercise effort that year, led the students in observing the rock types along the grid lines on our first day in the field, I roamed around the site, which I had never seen before, trying to imagine how to make a meaningful experience for the students. I was reminded of a line from the original Indiana Jones movie of a few years earlier in which the hero said, "I'm making this up as I go."

The surface geology on the undisturbed hilltop was mostly dolomitized marble encasing small outcrops of talc. The small talc outcrops that were mapped were likely

splays off of the main ore body, which dipped under the hilltop where we would work. Berg (1979) noted that a diabase dike cut across the west edge of the site, and that was recognized in the initial mapping. I expected the dike to give some sort of magnetic anomaly, which it did. I did not know what to expect from the seismic profiling. The mapping that Marian and Liz led did not extend to the schist mentioned by Berg (1979), which was geologically below the dolomitized marble and south of the area covered by the topographic map.

With a sense of the geology in hand and my *ad hoc* operations plan, we split into groups to collect data. We were on the site for just a few days in early June, but each group had a chance to learn a little about geophysical operations applied to mapping shallow targets and a chance to operate each of the two instrument systems that we had, despite losing one afternoon to a soaking, cold graupel shower (Liz was on site that afternoon and introduced me to the meteorological term graupel).

In the evenings, Marian, the students, and I would spend an hour or so graphing our data and discussing possible interpretations, planning for the next day, and so forth. One such evening discussion revolved around an apparent seismic velocity anomaly that we saw along one of the seismic lines.

THE 1984 SEISMIC ANOMALY

In those days, a relatively new approach to interpreting refraction seismic data was the Generalized Reciprocal Method (GRM; Palmer, 1980). Marian and I were early adopters of the technique. I taught the method in my geophysics courses at the university, and Marian and I gave short courses on the topic. The GRM was a revolutionary technique that depended on three stages of data processing. Each stage required the generation of a graphical display of many curves, the interpretation of which provided input parameters for the following stage. The relatively new personal computer of that era was ideally suited to storing the required volume of input travel time data, making the multitudinous GRM computations, and rendering the graphics, particularly the vertically exaggerated wavefront arcs in the final geologic cross-section. With the desktop computers available to us for the evening work sessions, we used the GRM code that I had developed based

on Palmer's (1980) equations for the display and analysis of our refraction seismic data.

In general, in those days, a minimum of five "shots" was needed for each geophone spread in order to get the volume of data needed for GRM processing (Lankston, 1990). After recording the five shots, the whole spread would be moved ahead along the line, perhaps just one quarter of a spread length, i.e., three geophone intervals when using a 12-channel system. The full details of the field operation are beyond the scope of this narrative, but are relatively easy to demonstrate in the field, and the students caught on quickly to the routine of setting the geophones, hammering at the various shotpoints, picking up the geophones, and moving the cable ahead a short distance to start the process anew. Their youthful enthusiasm was a definite plus, particularly for the task of swinging the sledgehammer at the numerous shotpoints. Since the mid-1980s, field instrumentation has advanced considerably, with the result that the five shots per spread then move ahead scheme is largely obviated. However, the technology around the sledgehammer source has not changed.

Figure 2 shows the travel time data collected along our Line B at the Regal site. The data points in figure 2 are connected in groups of 6 or 12 arrival times depending on whether the source point, i.e., the hammer impact point, was in the middle of a 12-geophone spread or off of the

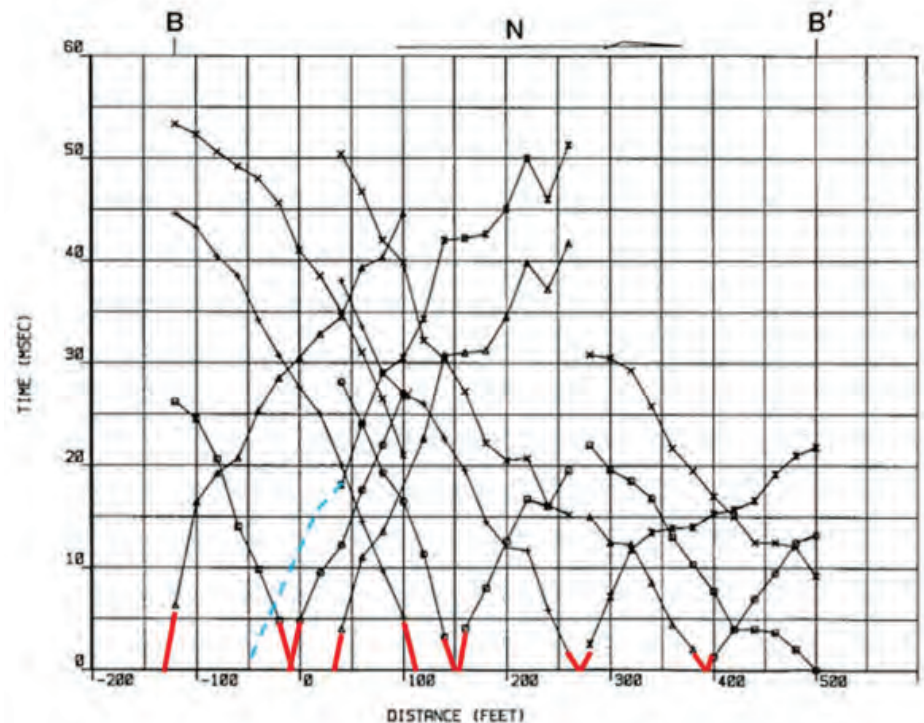


Figure 2. Image of Line B travel time data from a 1984 computer-generated plot. The red lines extend from a shotpoint to the near geophone of the respective spread. In some cases, the shot points are in the middle of a 12-geophone spread, e.g., the shotpoint at 150 ft. The blue dashed line projects from the first geophone of a spread to the approximate shotpoint. The blue line conveys no geologic information. It was added only to illustrate a long off-the-end shotpoint location as mentioned in the text. The red and blue lines were added to the image of the original plot in a graphics editor.



end of a spread, either near or far off-the-end in either the forward or reverse shooting direction.

The first of the three processing stages in a GRM interpretation is called the velocity analysis. The velocity analysis gives the true refractor velocity regardless of any irregularities in depth to the refractor. The velocity analysis can show lateral changes in the velocity along the refractor surface. As you might guess, refractor velocity is the parameter forwarded to the second GRM processing stage.

At the Regal Mine site, we did not anticipate a layered geology with a refracting horizon, such as bedrock, beneath alluvium, a common environment in which refraction seismic profiling is employed, because the bedrock was exposed at the surface. However, the velocity analysis computation scheme could still be applied to the travel time data to determine the velocities of the respective rock types that were exposed at the surface. Figure 3 is the GRM velocity analysis derived from the Line B data (fig. 2) for the rocks exposed at the surface.

A GRM velocity analysis graph (fig.3) plots time on the vertical axis and distance along the seismic line on the horizontal axis. In this coordinate system of time and distance, the inverse of the slope of a straight line through a set of related points indicates the true velocity of the rocks in the refracting horizon along that stretch of the survey line. In figure 3, I have drawn three straight-line segments that pass through groups of time–distance points that seem to be linearly related. One segment is short, and one might

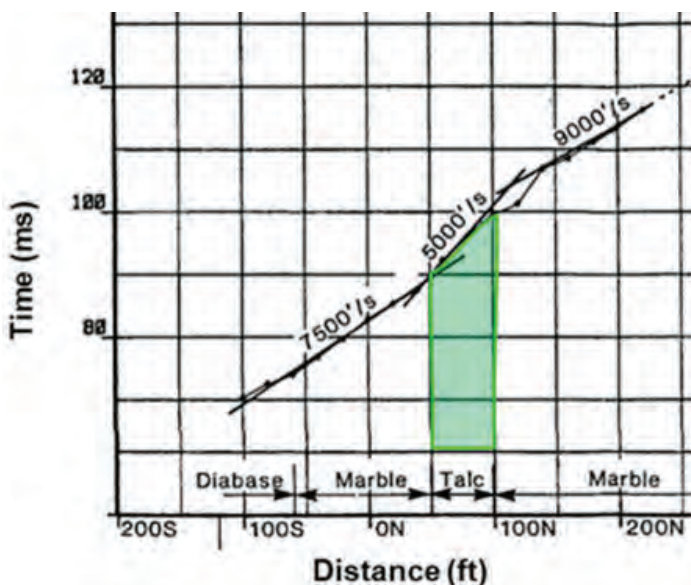


Figure 3. Image of the GRM velocity analysis for Line B from a 1984 computer-generated plot. The straight-line segments through the various points were drawn by hand. The time axis in this sort of graph is strictly for scaling in determining slopes of line segments, hence velocities, along the survey line. The 5,000 ft/s (1,525 m/s) zone is discussed in the text. The green highlight was added for this illustration in a graphics editor. The green shading and the labeled talc arrow are based on mapping, while the velocity analysis graph might suggest that the low velocity zone extends from 40 N to 120 N, i.e., a range of 80 ft (24 m).

dismiss it as being defined by insufficient data. However, that short 50 to 80 ft (15.25 to 24 m) segment along the seismic line with the velocity lower than the adjacent rocks just happened to be across a talc splay that had been identified during the surface mapping on day 1. Line B had not been placed such that it would cross the talc exposure. Line B was just one of the regularly spaced grid lines that was drawn on our basemap and staked on the ground. The discovery of the velocity anomaly was truly serendipitous.

I was a little surprised that the calculated velocities for “solid” rock were so low (fig. 3). From published values in geophysics textbooks, e.g. Dobrin and Savit (1988), I had expected the metamorphic rocks to exhibit velocities well above 10,000 ft/s (3,000 m/s). As a working hypothesis, I assumed that the lower than anticipated velocities were the result of weathering.

THE 1985 FOLLOW-UP

Intrigued by that correlation between a small talc outcrop and the velocity slowdown in the velocity analysis plot (fig. 3), I contacted Tad Dale with a research idea sometime after the 1984 field season. Contacting in those days was mainly by U.S. mail supplemented by the occasional land-line telephone call. Email was not yet a thing, and cellular unlimited talk and text was also yet to be available. Working through a plan for conducting the research during the next field season took some time.

I noted the curious seismic velocity–rock type correlation to Tad and allowed that the observation could simply be the result of mechanical weathering of the talc splay. I focused on mechanical weathering based on a statement published by Berg (1979) that said that talc was chemically stable at the surface. If so, any freeze–thaw cracks that contributed to the observed velocity decrease would die out with depth. If we could measure the velocities of the talc and the country rocks at depth, we could learn whether the talc inherently exhibited a seismic velocity slower than that of the surrounding rocks. My question to Tad was, could the mining company drill a hole through the Regal Mine ore body so that we could measure uphole seismic travel times from which to calculate the velocities in the unweathered subsurface?

The Geophysical Field Exercise did not return to the Regal Mine site in 1985, but moved out of the Ruby Range foothills into the Beaverhead Valley, continuing with the philosophy of multi-method data acquisition and interpretation. Part of the 1985 program related to a magnetic anomaly a few miles south of Beaverhead Rock (Lankston and Lankston, 1985). The source of that anomaly might be a diabase dike of the sort that is exposed in the vicinity of the Regal Mine but buried beneath Tertiary basin-fill. Through the years, the Geophysical Field Exercise collected a large volume of data in the Beaverhead Valley and in the vicin-

ity of the intersection of the Beaverhead and Blacktail Deer valleys, e.g., Hecker (1989) and Ahlert (1989).

James Buice was a new graduate student at Arkansas in 1984. He participated in the 1985 Geophysical Field Exercise, then went on to do thesis research near the Regal Mine. Some of the research continued efforts that were begun by the Geophysical Field Exercise students in 1984. No expansion of the Regal pit had occurred between the 1984 and 1985 field seasons. Through his position at Pfizer, Inc., Tad arranged for the boring that I had requested. The hole was drilled from the first bench, which was approximately 15 ft (5 m) below the natural surface on which the 1984 students had worked. The bench was developed on dolomitized marble. The talc dipped under the bench. The rule of thumb orientation of the talc body was a northward dip at approximately 45 degrees. During drilling, James collected cuttings for later determination of the top and bottom of the talc body. In his thesis, James referenced the uncertainty in determining the contacts as a result of the lag time in the samples being blown back to the surface (Buice, 1989).

The borehole geophone that we had (fig. 4) worked best in cased water wells. I had used the tool in Arkansas in water-monitoring wells in another avenue of my research in shallow target seismic methods. The rugosity of the open hole that the blast hole drill left us at the Regal Mine, however, caused some problems during data acquisition, mainly in knowing the exact depth of the sensor in the hole and in getting a good mechanical connection between the tool and the borehole wall. Of course, those uncertainties propagated into the interpretation of the data. Buice (1989) also referenced this uncertainty.

The Buice (1989) thesis reported on the integration of data from the 1984 program with the new data: specifically, in the present context, the correlation of the rock types interpreted from the drill cuttings with the subsurface interval velocities that were calculated from the uphole travel time measurements. I suppose that theses and dissertations always leave loose ends, i.e., subtle or explicit invitations for follow-up research. Sometime after James finished his analyses, I decided to take an alternate approach to relating the uphole times to the interpreted geology, and I built a raypath modeling program to calculate expected uphole times. Working down from the surface, adjusting velocities model layer by model layer, I constructed a subsurface model that yielded travel times (blue circles in figure 5A) that compared reasonably well to the 1985 travel time observations. That blue model (fig. 5B) reinforced the notion that the seismic velocity of the massive talc body was, in fact, less than the velocities of the overlying dolomitized marble and the underlying schist. You might ask, was my blue model in figure 5B correct? No. However, it did illustrate a velocity decrease from dolomitized marble to talc at

the interpreted dolomitized marble–talc contact at a depth of 65 ft (20 m).

The code for the modeling scheme to which the blue velocity–thickness layers depicted in figure 5B were submitted was very simple. I set the seismic signal path length to be the hypotenuse of a right triangle, i.e., a straight line. One leg of that right triangle was measured from the shot-point on the surface along strike to the top of the boring, and the other leg was assumed to be the vertical distance from the top of the boring, i.e., the surface of the bench, to the nominal subsurface recording point (appendix).



Figure 4. Deploying the downhole geophone. The black hose contained the geophone signal wires and the pneumatic tube for inflating the packing bladder, the black bulge at the base of the tool. The “drawworks” was the student or me manually handling the white polyethylene rope, which stretched under the weight of the tool and the hose, leading to uncertainty in the exact depth at which a reading was made. Photo by James Buice.



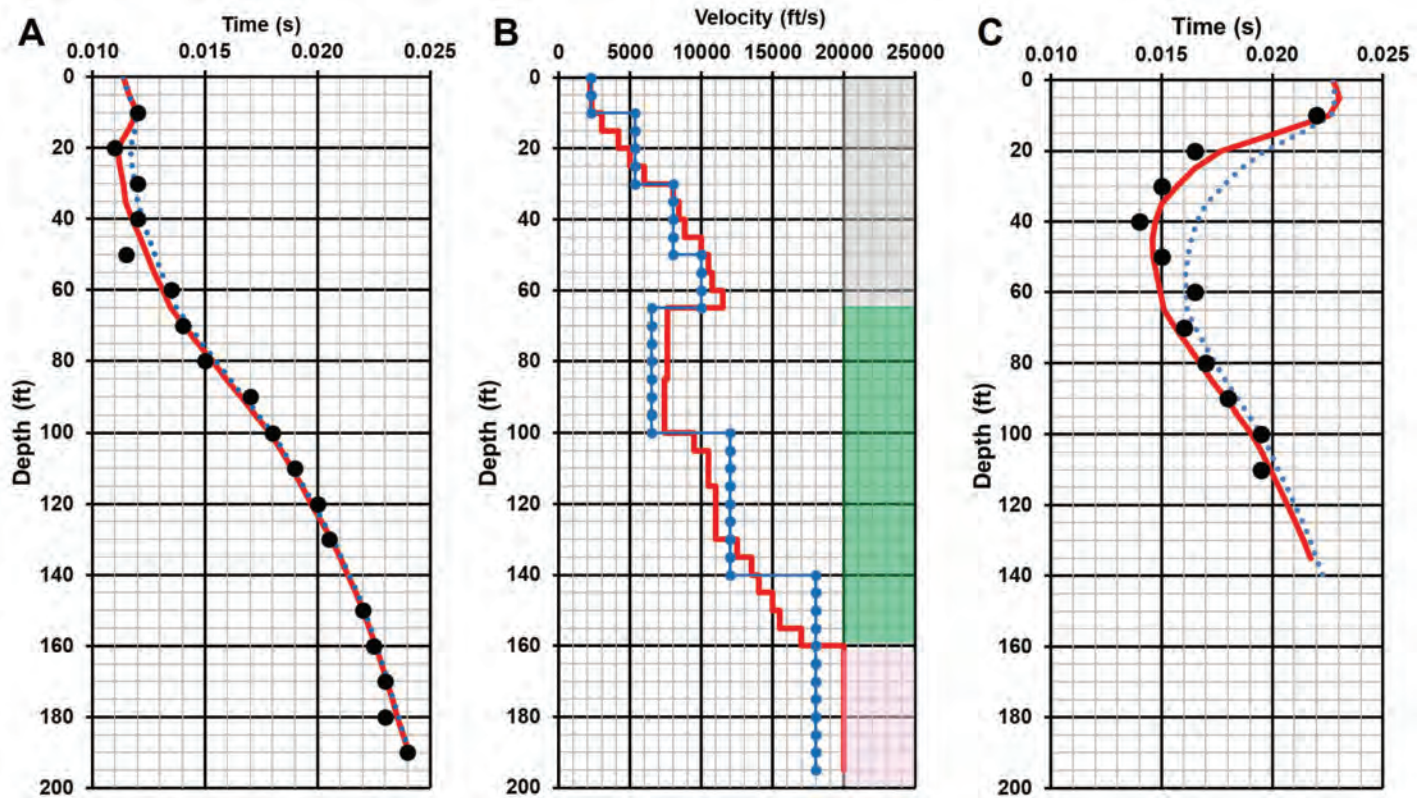


Figure 5. Raypath models and modeling results. The black dots representing the observed first break times in A and C are large to suggest the uncertainty in picking the uphole times (left–right dimension) and the uncertainty in depth to the sensor (vertical dimension). The offset from the boring was 25 ft (7.6 m) in A and 50 ft (15.25 m) in C, generally along strike. The two models in B were developed based on the comparison of observed and predicted travel times in A. The same model was then used to generate the curves for the offset in C. B shows the velocity–depth curves for the respective models. The blue model (B) was first generated around 1986 using the hypotenuse raypath assumption. The talc zone interpreted from cuttings (Buice, 1989) is indicated in green in B. Gray and pink in B are the interpreted dolomitized marble and schist, respectively. Details of the red and blue curves/symbols are discussed in the text.

That geometrical simplification worked tolerably well for comparing the observed uphole times to the modeled times in figure 5A, where the shotpoint was just 25 ft (7.6 m) from the boring. However, the blue circles do not track the observed times nearly as well where the offset between the boring and the shotpoint was 50 ft (15.25 m; fig. 5C).

At some point, I set aside time to code a new mathematical model in which refraction at each model boundary was considered (appendix). The physically more correct algorithm made use of the two parametric equations for calculating (1) a source to geophone offset distance and/or (2) a two-way reflection travel time given a specific value for the raypath parameter, which is usually denoted as p (eqs. 11-37 and 11-38 in Dobrin and Savit, 1988). The parametric equations had to be modified slightly because the raypath geometry for the uphole measurements was just half of the total path for which the parametric equations were derived.

The red curves in figures 5A and 5C are the results of the newer modeling scheme, and the results track the observed values better than those from the simpler hypotenuse path model. Part of the difference in the predicted times from the two modeling schemes is the effect of my subjective decisions as to the significance of certain observed

travel times in the shallow section in the uphole data from the 25 ft (7.6 m) offset during the two different stages of modeling. Decisions such as those in modeling are always open to discussion. Nevertheless, accounting for refraction at boundaries made a significant difference in the result for the 50 ft (15.25 m) offset case (fig. 5C), clearly justifying the extra effort.

The question above might be raised again. Does the red model in figure 5B correctly portray the subsurface geology? Still, no. The first issue is in the definition of the layer thicknesses. I set them to be 5 ft (1.5 m). I chose the 5 ft (1.5 m) interval such that a layer boundary would occur at the reported top of the talc, i.e., 65 ft (20 m; Buice, 1989). These are mathematical layers. Actual geologic contacts could be at any subsurface depth, but such detail was not available.

Perhaps more concerning is that the mathematical models (fig. 5B) assume uniform, horizontal layers, i.e., the one-dimensional assumption even though the site was known to be nominally two-dimensional. In a one-dimensional case, the seismic velocity can vary in only one dimension, i.e., only depth. The velocity of a given horizontal layer cannot change laterally. The one-dimensional logic was the foundation of the 1985 experiment, i.e., how



the velocity varied with depth. With the one-dimensional assumption, the position of the shotpoint on the surface of the bench could be at any point with respect to the boring, but the ray path equations for modeling would remain the same.

Knowing that the talc body dipped, if the source were placed in the dip direction with respect to the boring, the experiment could have been considered to be two-dimensional, with the seismic rays traveling in the vertical plane defined by the vertical line of the boring and the source point on the surface, i.e., the basic Euclidian rule for definition of a plane. Derivations of travel time equations for many two-dimensional geologic settings are given in geophysics textbooks, though generally assuming the source and geophone(s) are on the surface, not with the detectors in a borehole. One objective of a test with the source displaced in the dip direction with respect to the boring would be to identify reflections from the top and/or the bottom of the talc body in addition to gaining additional or corroborative velocity information using the one-dimensional modeling. The reflection aspect of such an experiment would be called a vertical seismic profile (VSP).

James and I actually did make several uphole surveys with the source down dip from the boring. I think that this was an *ad hoc* experiment given the availability of the boring and space on the bench plus access to the recording equipment, which was, as in 1984, provided by Marian's company. I really should note that Marian was on the site also. She generally had a camera in hand, but she also helped with the flow of the work, e.g., by keeping the rope and cabling pictured in figure 4 untangled.

Buice (1989) tabulated expected travel times for rays reflected from only the top of the talc body and for only one source to boring offset. The tabulated reflection times indicated that the times decrease from the time when the borehole geophone was at the 10 ft (3 m) depth to the time when the geophone was at the 60 ft (18.3 m) depth, i.e., near the interpreted top of the talc in the boring. This might seem counterintuitive, but the concept is correct. The trend of the predicted arrivals crosses the range of the first break wave train in the seismic traces, and the reflected signals could not be identified in the field record that was analyzed. This, of course, was discouraging. However, that does not say that the top of the talc is not a reflecting surface in this case. It indicates that our experiment of opportunity might have benefited from some planning. Perhaps a better experiment to test for reflections, specifically, would have been traditional surface geophone spreads with surface shotpoints.

In the uphole surveying with the source point placed along strike, the seismic rays of interest likely did not travel exclusively in the vertical Euclidian plane defined by the boring and the source point, the case illustrated in the ap-

pendix. Though the geology could be considered to be two-dimensional in the small area in which we were working, the seismic rays likely did not travel in that idealized plane. I have never attempted to develop code nor to find code on the internet that would account for the geometry of a dipping layer, sensors in a vertical boring, and a source point on the surface along strike.

Concerns of two- and three-dimensionality notwithstanding, a one-dimensional model can still be illustrative. The two velocity models (fig. 5B) show an unanticipated gradual increase in the velocity from the surface of the bench to the top of the talc body. With the top of the borehole being well below what I thought would be the freeze-thaw zone, I was surprised to see the modeled velocity steadily increasing to a depth of 65 ft (20 m), the Buice (1969) interpreted top of the massive talc (fig. 5B). I have interpreted that increase in velocity with depth to be indicating a decrease in the fracturing of the dolomitized marble with depth. That fracturing, I believe, was the result of the blasting that was done to develop the bench on which we worked, i.e., it was not a deep weathering phenomenon. To emphasize the interpretation of decreasing fracturing, the later model (red stairsteps in fig. 5B) shows many thin, 5 ft (1.5 m) layers, with each successively deeper layer being slightly faster than the overlying layer.

I felt that the degree of fracturing in the dolomitized marble in the upper section of the models could explain the gradual velocity increase with depth. Figure 5B shows the same sort of stair-stepping velocity increases with depth below 100 ft (30 m). This is a mathematical result. Figure 5B illustrates the non-uniqueness of the modeling process. I offer no geological explanation for the stair-stepping increase in the velocity, model layer by model layer (red curve in figure 5B), in the deeper section. Perhaps the blockier shape of the older (blue) model in figure 5B is geologically better.

At the contact between the overlying dolomitized marble and the top of the talc (65 ft in fig. 5), both stages of modeling showed a sharp decrease in velocity, from approximately 11,000 ft/s to approximately 7,000 ft/s (approximately 3,350 m/s to 2,100 m/s), a contrast magnitude similar to the surface observation in 1984 (fig. 3). However, the velocities were still slower than I anticipated from textbook tabulations.

At that point in the mid to late 1980s, I harbored a thought that seismic surveying might be useful in talc exploration, but I was soon to be gone from the university and from my research interests in shallow target geophysics. My new employer was a large petroleum company again.





Figure 6. Tourist's view of the Regal Mine looking generally northwest from the Sweetwater Road on a summer afternoon in 1998. Using an aerial view from Google Earth dated 1995, the light-colored highwall would appear to be at least 700 ft (210 m) across. The dark area to the left likely trends at a right angle to the light-colored highwall, generally toward the camera. The darkness is partly shadow from the afternoon sun and may include the diabase dike mentioned in the text. The tree on the skyline of this view is among the few trees on the "mostly treeless...hilltop" that is referenced in the text. The tree is clearly imaged in the 1995 Google Earth aerial view.

BACK TO THE THREAD

In 1998, our son was in Boy Scouts working on the Geology Merit Badge. Marian and I decided that our summer family road trip that year would be to Montana, where our son could complete some requirements for the merit badge. One badge requirement was to collect a certain number of geologic samples. Marian and I confidently stated that we knew of a site near Dillon where we could pick up talc from the middle of the county road. What a surprise we had as we approached Dillon from the north and looked eastward across the Beaverhead Valley. We saw a white spoils pile that had not been there in the mid-1980s. Nevertheless, we drove up the Sweetwater Road toward the busy mine (fig. 6), and we collected a handful of talc chips from the ditch alongside the road (fig. 7).

Four more years slipped by. In 2002, I was still working for the oil company. I did fluid overpressure risk analysis for deep-water oil wells in those days, a long way from hard rock talc exploration and the broader sphere of shallow target geophysics where I had practiced for many years. However, in my 12 years in big oil after leaving the university, I had learned a lot about seismic methods, particularly the seismic data processing, and about wireline logging employed by the petroleum industry.

Those geophysical tools might have been developed for hydrocarbon exploration, but under it all, they were just tools for studying the earth. For example, nothing said that a sonic well log gave velocities of sandstone and shale only. Actually, the sonic log was developed to determine porosity of sandstone. Porosity, though, is not measured directly. Porosity is derived using a weighed by volume equation from the downhole sonic travel time measurements. The downhole measurements are just fine-scale refraction

seismic experiments. The downhole parameter that is actually measured in sonic well logging is travel time between the source and the sensor. The travel time is called transit time in the well logging industry, and it is proportional to $1/\text{velocity}$. The measured transit times of the sonic log are inverted and scaled to account for the spacing between the source and detector on the tool to yield rock velocities at sub-foot or sub-meter intervals in the subsurface. The technique can work for any rock type, not just clastics.

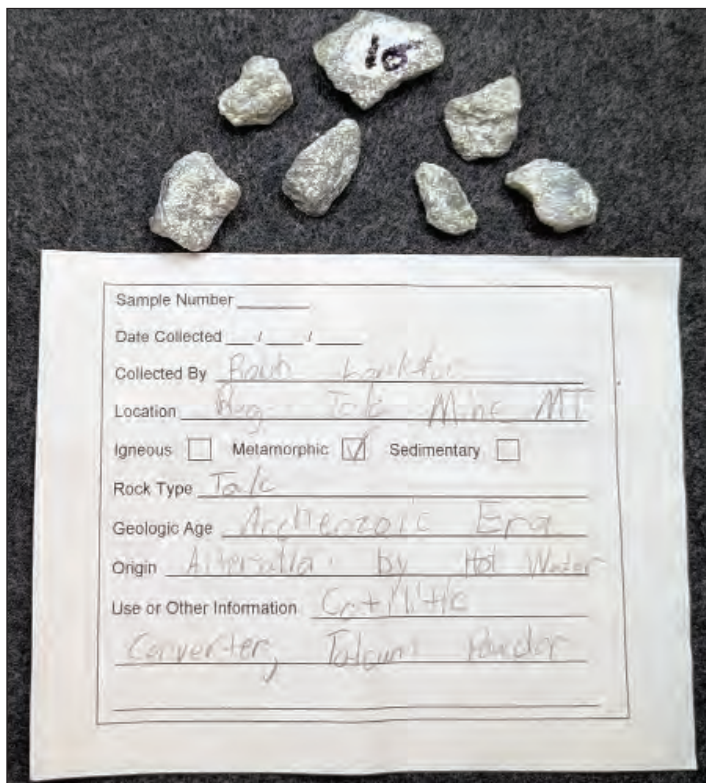


Figure 7. "Talc" samples collected near the Regal Mine in 1998.

The Leading Edge, a monthly journal published by the Society of Exploration Geophysicists, showed up on my desk one day that fall, and it had an article on the use of seismic surveying in mapping the Three Springs talc body in Western Australia (Urosevic and others, 2002). Of course, I was drawn to the article. Other than the non-traditional geologic target in the study, the seismic technology expressed in the article did not seem to be unusual to me at that point in my career. The authors stated that they had measured the velocities of rock samples in the laboratory and had done geophysical borehole logging at their site prior to recording their seismic line in 1993. The reported geophysical logging would have yielded subsurface velocities directly at sub-meter vertical intervals, though the paper did not show their logs. By contrast, the 1985 uphole survey at the Regal Mine yielded subsurface velocities indirectly and at a much coarser vertical interval through modeling.

Urosevic and others (2002) concluded that “seismic methods could be a valuable exploration tool for talc exploration in this area.” Of course, an obvious caveat is at the end of the quotation, i.e., “in this area.” Understanding the physical properties of the rocks at a site is necessary for any geophysical exploration, not just for seismic methods. The Urosevic and others (2002) study complemented my notion of 15+ years earlier and Buice’s (1989) conclusions as to the potential utility of seismic methods in mapping talc occurrences.

Urosevic and others (2002) did not give the seismic velocities that they had measured in the laboratory or had determined from logging. In doing research for this narrative, I discovered another paper by the Australian researchers in which they published their laboratory measured velocities for samples from the Three Springs site (Vella and others, 2003). The velocity of dolomite was reported as 6,818 m/s (22,370 ft/s). Two talc velocities, 4,348 and 4,535 m/s (14,260 and 14,880 ft/s) depending on the grade of the talc, were tabulated. Those reported velocities were more in the range that I was expecting to see in the modeling during the 1980s. The reported velocity difference between dolomite and talc in the Vella and others (2003) study is approximately 2,500 m/s (8,200 ft/s).

The maximum modeled value (red curve in fig. 5B) for the dolomitized marble above the talc was 3,505 m/s (11,500 ft/s). The modeled talc velocity below that contact was 2,290 m/s (7,500 ft/s). The difference between the velocities at the 65 ft (20 m) depth in the model (fig. 5B) is approximately 1,220 m/s (4,000 ft/s), approximately half of the difference in the Australia case, but a difference in the same sense, nevertheless.

CONCLUSIONS

As I mentioned in the opening of this narrative, talc entered my life at unexpected times. Observing that talc exhibited a lower seismic velocity than its surrounding metamorphic rocks was, itself, unexpected. The lower seismic velocity of massive talc compared to the surrounding rocks was independently observed in studies at another site half a planet away. Urosevic and others (2002) and Vella and others (2003) added to the limited literature on the use of seismic methods to map talc occurrences, and complemented my notion from the 1980s.

Site-specific differences in geophysical properties, e.g., my velocities versus those from the Australian researchers, are bound to be observed in measurements regardless of the physical property being studied. Further, field versus laboratory methods may have differences. In the case of talc, though, the key observation is that massive talc does seem to exhibit a seismic velocity lower than its adjacent host rocks.

ACKNOWLEDGMENTS

Marian Lankston contributed to this story in many ways through the years. I particularly appreciate Marian’s curation of photographs and artwork from the era of events in this narrative, resources that I drew upon for this story, and I appreciate her maintenance of a certain Boy Scout’s rock and mineral collection alongside her own collection. I am grateful to John Childs, a geologist well known to TRGS members, for his suggestions for improving an early draft of this manuscript. Tad Dale, now retired after a long career in the minerals industry in western Montana, coordinated Pfizer, Inc., resources for the 1984 and 1985 fieldwork mentioned herein and, more recently, helped to refresh my memory of certain details in this story. Finally, I am grateful to longtime acquaintance, TRGS contributor, and acknowledged talc expert, Dick Berg, for his review of the “final” manuscript and for his suggestions for polish to this technical memoir.

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APPENDIX

In figure A1, the column sketches the boring as the vertical line on the left, with the geophone points indicated by blue circles. The geophone points were nominally at intervals of 10 ft (3 m). The yellow starburst is the source point on the surface, in this case, 25 ft (7.6 m) from the boring. The horizontal and vertical scales of the sketch are basically the same. The layer boundaries for the blue model in figure 5 are indicated by the horizontal lines. The layer

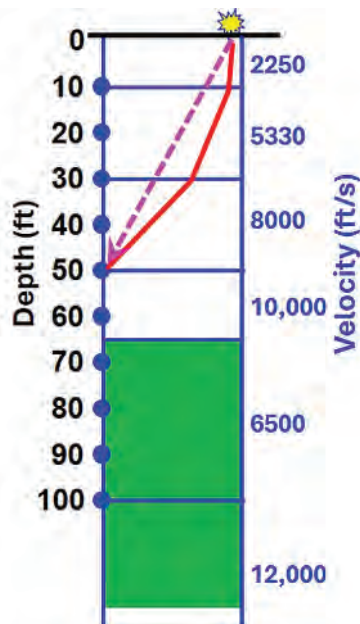


Figure A1.

velocities for the model are on the right. The green area is the upper part of the interpreted talc body (fig. 5B) from Buice (1989). The dashed magenta line shows the assumed hypotenuse raypath for the ray that traveled from the source to the geophone at the 50 ft (15.25 m) depth. Of course, an analogous straight line could be drawn from the source point to each of the geophone points. The path length is trivially calculated using the Pythagorean theorem, and the diagonal path lengths through each layer are proportional to the respective layer thicknesses.

In the later modeling, refraction at the layer boundaries was taken into account using the parametric equations referenced in the text. The red line shows, conceptually, how the raypath was bent at each velocity boundary for the ray that traveled from the source to the geophone at the 50 ft (15.25 m) depth. For the near-surface geophone points, the raypath lengths were significantly different between the two assumptions for the path geometry, particularly for the 50 ft (15.25 m) offset case (fig. 5C). However, the path length difference diminished with depth to the geophone as the refractions at the respective boundaries decreased and the path approached the straight line, i.e., the hypotenuse, geometry.



INTEGRATED INVESTIGATIONS OF CRETACEOUS MAGMATIC-HYDROTHERMAL SYSTEMS LEVERAGING ONGOING EARTH MRI GEOPHYSICAL ACQUISITIONS ACROSS THE IDAHO–MONTANA PORPHYRY BELT

Snir Attia,¹ Benjamin P. Magnin,¹ Eric D. Anderson,¹ Julia A. McIntosh,¹ Kyle A. Eastman,² Jarred L. Zimmerman,² Theresa M. Schwartz,³ Jonathan A. Funk,¹ Karen Lund,¹ George Brimhall,⁴ Neil P. Griffis,¹ April D. Allen Langhans,¹ Cameron M. Mercer,¹ Michael A. Cosca,¹ and Celestine N. Mercer¹

¹U.S. Geological Survey, Geology, Geophysics, and Geochemistry Science Center, Denver, Colorado

²Montana Bureau of Mines and Geology, Montana Technological University, Butte, Montana

³U.S. Geological Survey, Geologic and Environmental Change Science Center, Denver, Colorado

⁴Clementine Exploration LLC, Wise River, Montana

ABSTRACT

The northern Rocky Mountains contain many significant Cretaceous and younger mineral deposits that originated before, during, and after the Paleogene transition from regional contraction to extension. To better understand mineral systems throughout the region, high-resolution airborne radiometric and magnetic surveys are being flown over a broad area corresponding to the Idaho–Montana porphyry belt as part of the U.S. Geological Survey Earth Mapping Resources Initiative. In southwestern Montana, these continuing efforts have centered on the Boulder batholith and surrounding areas, a region of historic metal production from Late Cretaceous to Paleogene deposits. Our ongoing project aims to utilize these new, high-resolution geophysical data to better characterize and understand the geologic and metallogenic framework of these deposits through integrated field, geophysical, geochemical, and geochronologic investigations. We highlight our preliminary work that leverages airborne geophysical data, on-the-ground geophysical measurements, geologic mapping, and numerical methods to elucidate the relationships between magmatism and metallogenesis in several localities surrounding the Boulder batholith. The selected study areas in southwestern Montana include the Elkhorn mining district, Clementine prospect in the Big Hole Canyon pluton area, Winston district, and Bison Creek area of the Rimini district, demonstrating the broad applicability of these new geophysical data for both geologic and metallogenic framework studies and exploration from regional to local scales.

INTRODUCTION

The U.S. Geological Survey (USGS) Earth Mapping Resources Initiative (Earth MRI) aims to improve the Nation's understanding of critical mineral resources (Day, 2020; Hammarstrom and others, 2023). To help address this goal, high-resolution airborne geophysical surveys are being flown over a broad area in the northern Rocky Mountains, corresponding to the Idaho–Montana porphyry belt (Rostad, 1978; Chadwick, 1981; Taylor and others, 2007), to better understand mineral systems throughout the region.

The northern portion of the U.S. Rocky Mountains, north of a broadly defined boundary loosely coincident with the Snake River Plain, contains many significant magmatic-hydrothermal mineral deposits that originated before, during, and after the Paleogene transition from regional contraction to extension (fig. 1; Armstrong and Ward, 1991; Burchfiel and others, 1992; Scarberry and others, 2021). In southwestern Montana, continuing geophysical data acquisitions have centered on the informally named Boulder batholith and surrounding areas, a region of historic metal production from Late Cretaceous to Paleogene porphyry, epithermal vein, breccia, and skarn deposits (Gammons and others, 2020).

The broader northern Rocky Mountains region is underlain by Archean sialic cratonic domains connected by Paleoproterozoic intracratonic mobile belts, and subsequently modified by both Mesoproterozoic intracontinental rifting and the Neoproterozoic–Paleozoic rifting of Rodinia (Sims and others, 2004, 2005; Lund and others, 2015). Overlying sedimentary (and subsidiary volcanic) strata record multiple phases of basin evolution including: (1) Mesoproterozoic rift-related opening of the Belt basin; (2) Cambrian–Devonian continental margin development in response to Rodinian rifting; (3) Mississippian–Triassic retroarc basin development associated with renewed tectonism along the western Laurentian margin, as well as the uplift and exhumation of the Ancestral Rocky Mountains in the Laurentian interior; (4) Jurassic–Paleogene foreland basin development associated with Cordilleran orogenesis; (5) Paleogene exhumation partly associated with post-orogenic extension; and (6) Neogene Basin and Range-associated and Yellowstone hotspot-associated extension (see Coney and Harms, 1984; Bond and others, 1985; Burchfiel and others, 1992; DeCelles, 2004; Yonkee and Weil, 2015; Blakey and Ranney, 2018; Schwartz and others, 2019; Chapman and others, 2021; among others). Archean–Paleoproterozoic basement rocks and Mesoproterozoic through Cretaceous strata variably compose structural culminations of the Sevier fold-thrust belt and adjacent Laramide intraforeland uplifts (Vuke and others, 2007), which host Late Cretaceous–Pa-



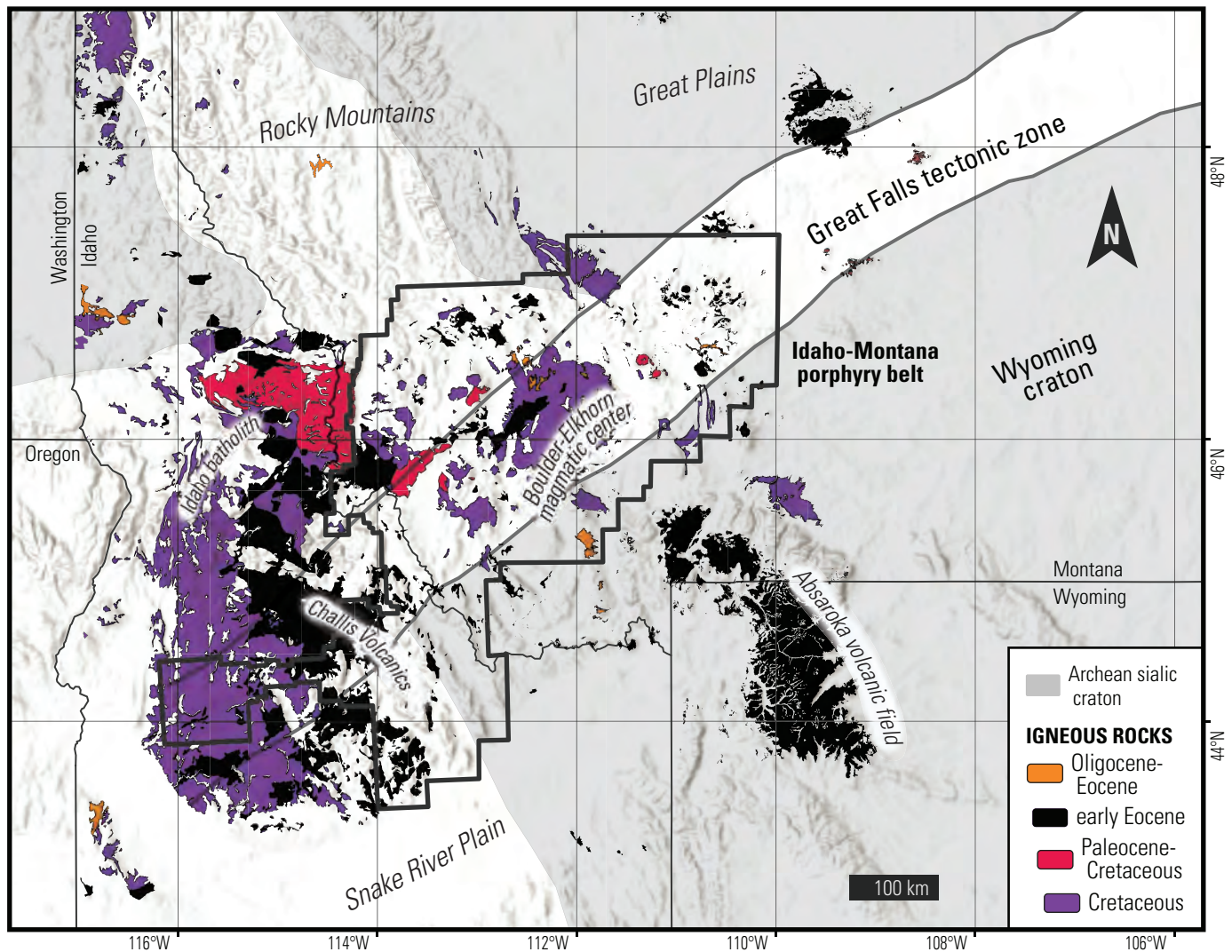


Figure 1. Cretaceous–Paleogene igneous rocks of the northern Rocky Mountains. Geology adapted from the State Geologic Map Compilation (Horton and others, 2017) following Love and Christiansen (1985), Vuke and others (2007), and Lewis and others (2012), modified from Gaynor and others (2025). Idaho–Montana porphyry belt of Taylor and others (2007). Basement domains of Lund and others (2015). Topography in all figures from U.S. Geological Survey 3D Elevation Program. Datum: WGS84.

leocene magmatic rocks associated with prodigious mineral deposits (Gammons and others, 2020).

Late Cretaceous–Paleocene magmatism and orogenesis extended from the informally named Idaho batholith at the western edge of Laurentia to the Little Rocky Mountains at the western margin of the Great Plains (Scarberry and others, 2021). Late Cretaceous–Paleocene magmatism appears preferentially associated with areas underlain by Paleoproterozoic intracratonic mobile belts, particularly the Great Falls tectonic zone (fig. 1; O'Neill and Lopez, 1985; Sims and others, 2004; du Bray and others, 2012). Porphyry-related mineral deposits of Cretaceous, Paleocene, and Eocene ages are also clustered along a northeast-striking Idaho–Montana porphyry belt (Taylor and others 2007), further suggestive of basement control on syn- to post-orogenic petro- and metallo-genesis (Lund and others, 2023). Initiation of extensive Eocene volcanism in the Challis–Kamloops–Absaroka belt was broadly contemporaneous with the relatively rapid transition from contraction to

extension in the northern Rocky Mountains (Armstrong and Ward, 1991; Dostal and Jutras, 2022). In the late Eocene–Oligocene, more limited bimodal magmatism locally associated with topaz rhyolites occurred along with continued extension and intermontane landscape development, including the Helena, Elliston, Virginia City, and Hog Heaven volcanic fields in Montana and the Salmon Creek and Potlatch volcanic fields in Idaho (Chadwick, 1978; Christiansen and others, 1986; Lewis and others, 2005; Scarberry and others, 2023).

The new Earth MRI geophysical data acquisitions provide opportunities to examine the complex history of tectonism, magmatism, and hydrothermal activity that generated widespread and varied mineral deposits across the northern Rocky Mountains. Our ongoing efforts aim to build on these high-resolution geophysical data to better characterize and understand the geologic and metallogenic framework of these deposits through integrated field, geophysical, geochemical, and geochronologic investigations. We highlight



preliminary work at four selected study areas surrounding the Boulder batholith that represent Cretaceous–Paleogene magmatic-hydrothermal systems associated with porphyry deposits and are at different stages of exploration and extraction (fig. 2). The highlighted examples include: (1) radiometric signatures of alteration around the Turnley Ridge porphyry deposit in the Elkhorn mining district (fig. 3); (2) magnetic structure of the informally named Big Hole Canyon pluton and nearby intrusions related to hydrothermal alteration at the Clementine prospect, which occurs above a hypothesized buried porphyry system (fig. 4); (3) preliminary geophysical interpretation of mineral systems associated with felsic porphyry stocks in the Winston district (fig. 5); and (4) 3D magnetic inversion of an exceptionally high-amplitude magnetic anomaly in the Bison Creek area, which is buried underneath Eocene volcanic rocks near the Rimini mining district of the northwestern Boulder Mountains (fig. 6). These examples illustrate the utility of Earth MRI geophysical data acquisitions for further examination of known mineral systems, providing insight into identified prospective areas, identifying new areas of interest for further study, and generating new hypotheses and questions.

METHODS

Geophysical Data

Helicopter-borne magnetic and radiometric surveys were flown by Sander Geophysics Limited over the Boulder batholith and surrounding areas (Allen Langhans and others, 2024; Anderson and others, 2025a). The surveys consist of east–west flight lines with 200 m spacing and north–south control lines with 2,000 m spacing at a nominal terrain clearance of 100 m. Magnetic data were acquired using a magnetometer mounted in a stinger, and radiometric data were collected with a fully calibrated gamma-ray spectrometer stowed on the aircraft. Both magnetic and radiometric data were processed using industry standard methods (Luyendyk, 1997; Minty and others, 1997). To ground-truth the airborne survey data, measurements of magnetic susceptibility and concentrations of potassium, thorium, and uranium were collected on rock outcrops throughout the region (Anderson and others, 2025b). For magnetic susceptibility, an SM-30 meter was used to make a series of 16 measurements on representative rock types at each outcrop to ensure robust characterization of the materials present at the surface. A total of 3,888 individual magnetic susceptibility measurements have been collected across the Boulder batholith area. Gamma ray measurements were made using a Radiation Solutions RS-230 BGO Super-SPEC Handheld Gamma-Ray Spectrometer (GRS), which averages measurements over a 2-min interval to provide concentrations of potassium (weight %), thorium (ppm), and uranium (ppm). A total of 199 measurements have been made on altered and unaltered plutonic, volcanic, and sedimentary rocks across the Boulder batholith region.

Processed magnetic data are presented as the anomalous magnetic field (AMF), which represents the measured total magnetic intensity minus the regional magnetic field. The data were gridded at a 40 m cell size using minimum curvature gridding and a histogram equalization color distribution. A reduced-to-pole transformation using a magnetic inclination of 69.6° and declination of 12.2° was applied to the AMF grid, which corrects for the inclination of the Earth's magnetic field and ideally relocates magnetic anomalies over their causative sources (Baranov and Naudy, 1964).

Processed radiometric data, which represent the top 60 cm of materials at the surface, are reported in percent concentrations of potassium and ppm of equivalent thorium (eTh) and equivalent uranium (eU). Airborne radiometric measurements collected at heights greater than 350 m above topography exhibited low signal and are not included in the final maps. Measurements for each element were gridded at 40 m cell size to produce elemental maps. The airborne and GRS measurements were similarly processed using ratios that may indicate areas of potassium enrichment from hydrothermal alteration. The ratios include potassium to thorium (K/Th) and potassium to uranium (K/U). In addition, a potassium enhancement (K enhancement) map was generated by combining radioelement concentrations and ratios into a ternary image displaying potassium in red, potassium to thorium in green, and potassium to uranium in blue.

3D Magnetic Inversion

To constrain the 3D geometry of subsurface magnetic sources in the Big Hole Canyon pluton and Bison Creek anomaly, AMF data were inverted using Mag3D software (Li and Oldenburg, 1996). The inversion program constructs a subsurface model of piecewise constant magnetic susceptibility that reproduces the observed AMF data. The inversion is iteratively solved by minimizing a multicomponent object function, which usually specifies a desired smoothness and comparison to a reference model, until the observed data are reproduced to within a specified noise level. A depth-weighting function is also applied that distributes magnetic sources at depth instead of concentrating them near the surface. This inversion method assumes that magnetic remanence and self-demagnetization (e.g., rocks with magnetic susceptibilities >0.1 SI) are minimal in modeled sources, which is likely the case for most magnetic intrusions within the study areas (Hanna and others, 1994).

However, the Bison Creek anomaly has a strong, inclined dipolar character even after the reduction-to-pole transformation, indicating significant remanence and/or self-demagnetization. In such cases where these assumptions are invalid, traditional inversion methods cannot be applied without knowledge of the orientation of the source's magnetization vector (i.e., induced + remanent



+ effects of self-demagnetization). An alternate inversion method using the amplitude of the anomalous magnetic field, which is not dependent on the orientation of the magnetization vector, was used for the Bison Creek anomaly (Li and others, 2010). Although this method accounts for the unknown effects of remanence and self-demagnetization, the solved distribution of magnetic material in the resulting model is usually more diffuse and of higher magnitude than in reality (Li and others, 2010).

Subsurface magnetic structure of the Big Hole Canyon pluton and Bison Creek anomaly was modeled within defined 17.6 km by 15.9 km and 7.2 km by 7.2 km areas, respectively. A 3D mesh was created for each location using a 100 m by 100 m xy (i.e., horizontal) cell size for Big Hole Canyon and a 50 m xy cell size for Bison Creek. Cell thicknesses (i.e., z-axis) are one-half the xy cell size down to one-fourth the model length and are equal to the xy cell size down to one-half the model length. Each model was expanded by three progressively larger cells in the $\pm x$, $\pm y$ and $-z$ directions to reduce edge effects. Observed AMF and magnetic amplitude data were decimated so that each surface cell corresponds with two datapoints along flight lines. The magnetic observation height was calculated by adding the helicopter's laser altimeter to a 10 m digital elevation model. The same digital elevation model was used to constrain topography.

Near-surface geologic models were created for the two study areas to constrain the inversion, based on 1:100,000-scale geologic maps, topography, estimated unit thicknesses, and magnetic susceptibility measurements (table 1). These models help to produce geologically realistic results by using geologic mapping to specify the rock types and magnetic susceptibilities of the shallowest portions of the subsurface, 300 m deep for Big Hole Canyon pluton area and 75 m deep at Bison Creek, below which the magnetic inversion is not directly constrained. Multiple near-surface geologic models were produced to test different hypotheses for the magnetic source of the Bison Creek anomaly. The simplified, near-surface geologic models are constructed with vertical lithologic contacts, which is equivalent to assuming no significant drift in location of contacts over the shallow modeled depths. Magnetic susceptibility measurements from the Big Hole Canyon pluton area were incorporated from Hespenheide (2003) in addition to data from Anderson and others (2025b). Unit thicknesses are based on geologic mapping by McDonald and others (2012). Maximum, minimum, and average magnetic susceptibility values were assigned to each unit in the model (table 1). Outside the near-surface model from the Big Hole Canyon pluton area, a constant maximum (0.2 SI), minimum (0 SI), and reference susceptibility (0.001 SI) were assigned. Considering the high magnitude of the Bison Creek magnetic

Table 1. Magnetic susceptibility and thickness values used for near-surface geologic models.

Unit	Thickness (m)	Lower Bound (SI)	Upper Bound (SI)	Reference (SI)	<i>n</i>
Cenozoic surficial deposits	75 (25)*	0	0	0	0
Eocene rhyolite	75	0.00112	0.00436	0.002749	64
Paleogene basalt	75	0.00132	0.00744	0.002945	16
Eocene Lowland Creek Volcanics	75	0.000122	0.000759	0.000249	32
Cretaceous intermediate intrusive rocks	300	0.000714	0.0853	0.001	0
Cretaceous Big Hole Canyon pluton					
Northern phase	300	0.000718	0.0582	0.0371	21
Southern phase	300	0.0185	0.132	0.068	36
Cretaceous aplite	75	0	0.001	0.0005	0
Cretaceous Butte Granite	300 (75)	0.0118 (0.000042)	0.0294 (0.0348)	0.021896 (0.0139)	48 (144)
Cretaceous diorite	75	0.00564	0.0661	0.029296	16
Cretaceous Moose Creek pluton	300	0.000714	0.0088	0.005262	48
Cretaceous Elkhorn Mountains Volcanics					
Middle member (informal)	75	0.000139	0.0698	0.010998	176
Lower member (informal)	75	0.000034	0.0582	0.005247	176
Phanerozoic sedimentary rocks	300	0	0.000603	0.000105	287
Mesoproterozoic sedimentary rocks	300	0.000002	0.00496	0.000209	80

*Numbers in parentheses represent values for the Bison Creek area near-surface geologic model (75 m maximum depth) that differ from the Clementine prospect (300 m maximum depth).



anomaly, a maximum susceptibility of 2 SI was assigned, with the minimum and reference susceptibilities kept as 0 SI and 0.001 SI, respectively. These models act as an upper and lower bound, as well as a reference and initial model to constrain the inversion results.

SELECTED STUDY AREAS

Elkhorn District Porphyry Related Alteration

The Elkhorn mining district, in the Elkhorn Mountains along the eastern margin of the Boulder batholith (fig. 2), consists of a Cu-Mo porphyry system, gold and iron skarns, tourmaline breccia pipes, and carbonate replacement deposits. The district is known for past production of precious (Au, Ag) and base metals (Pb, Zn, Cu) from the skarn and carbonate replacement deposits (Everson and others, 2005). The area also preserves a complex geologic framework of

Mesoproterozoic to Cretaceous strata that have undergone multiple phases of deformation and Late Cretaceous volcanic and plutonic rocks that include ca. 89–74 Ma compositionally variable intrusions (fig. 3A; Klepper and others, 1957; Lund and Aleinikoff, 2022). A porphyry deposit is centered on and related to the ca. 89–85 Ma informally named Turnley Ridge stock (Steeffel and Atkinson, 1984; Everson and others, 2005; Lund and Aleinikoff, 2022). Previous alteration mapping shows a K-feldspar-biotite core zone surrounded by quartz-sericite and propylitic assemblages, with alteration boundaries offset sinistrally along the Boulaway fault (fig. 3B; Steeffel and Atkinson 1984). Skarn and carbonate replacement deposits found near the margins of diorite and granodiorite intrusions are most pronounced around the ca. 80–78 Ma informally named Black Butte and Cemetery Ridge stocks (Everson and others,

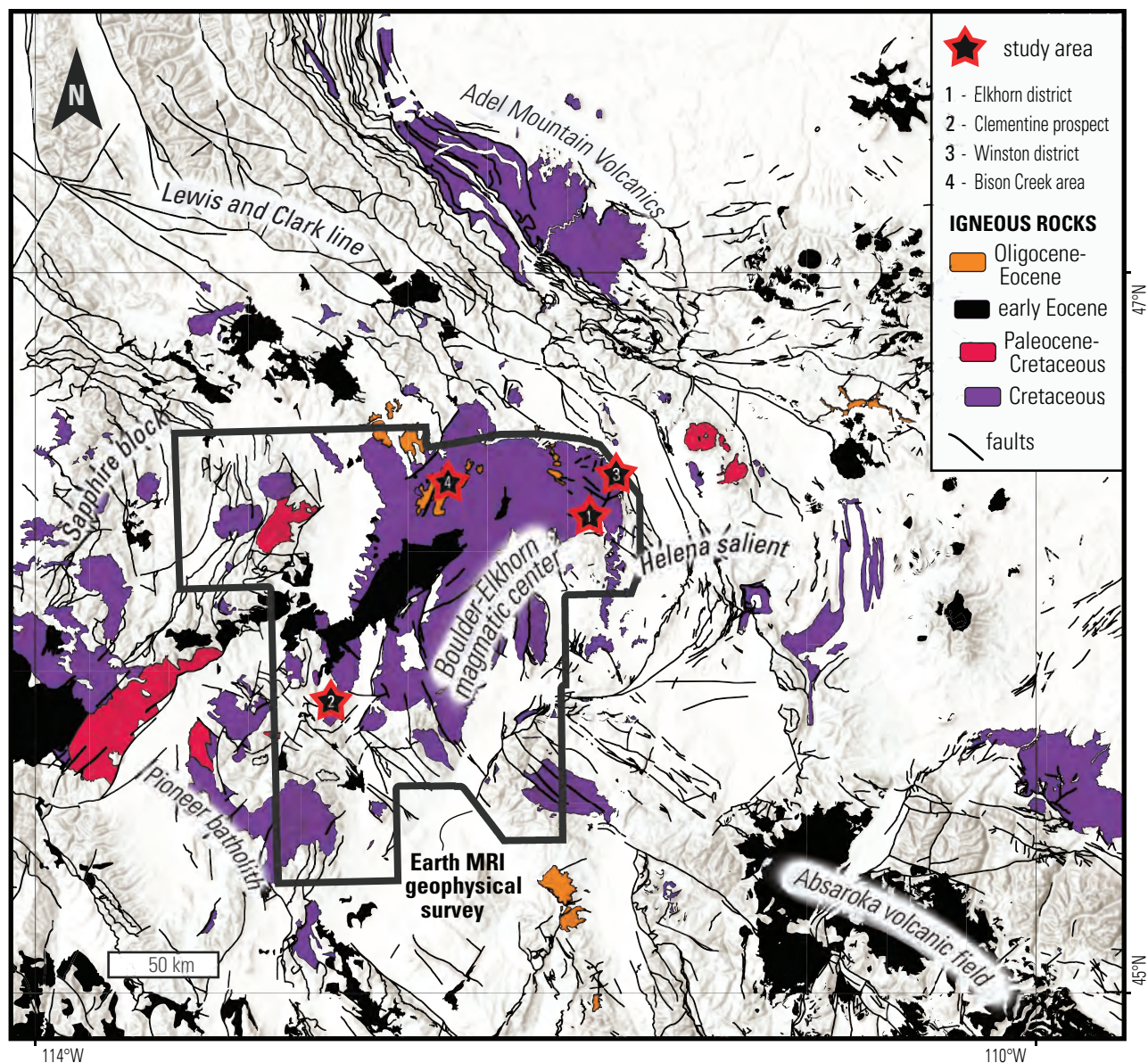


Figure 2. Selected study areas within the Helena salient and Earth MRI airborne geophysical survey area (Allen Langhans and others, 2024; Anderson and others, 2025a). Geology adapted from the State Geologic Map Compilation (Horton and others, 2017) following Love and Christiansen (1985), Vuke and others (2007), and Lewis and others (2012). Datum: WGS84.



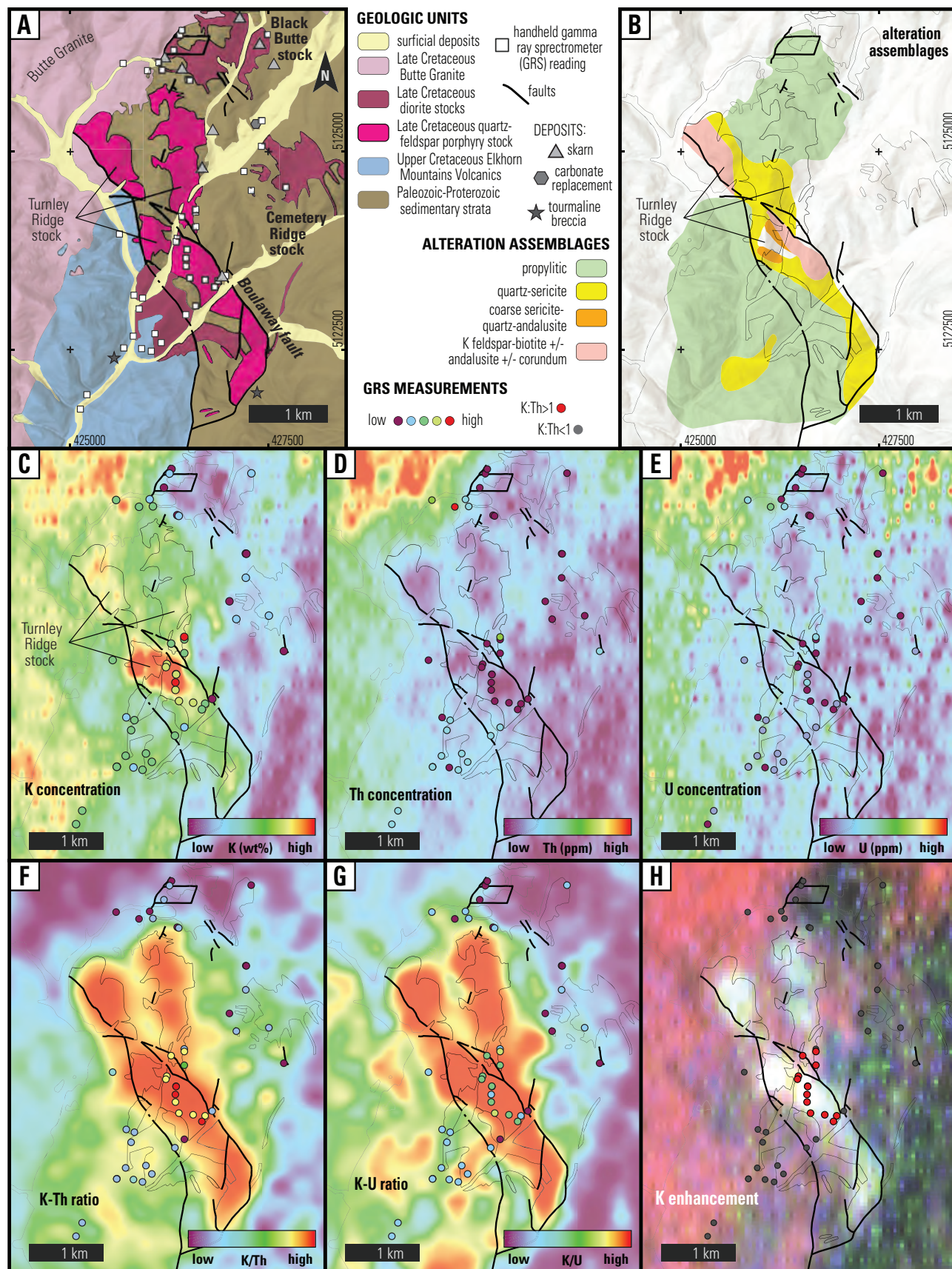


Figure 3. Geology, alteration, and radiometric data across the Elkhorn mining district and Turnley Ridge porphyry deposit, modified from Anderson and others (2025c). (A) Simplified geology from Reynolds and Brandt (2006), datum: WGS84 zone 12N, same extent for all panels. (B) Alteration assemblages related to the Turnley Ridge porphyry mineral system from Steefel and Atkinson (1984) with outlines of mapped intrusions for reference. Earth MRI airborne radiometric results with corresponding handheld gamma ray spectrometer (GRS) point measurements shown in panels C–G (Allen Langhans and others, 2024; Anderson and others, 2025a,b). (H) Airborne radiometric potassium enhancement map with K/Th from handheld GRS data.

2005). Sulfur isotope data suggest that the tourmaline breccia pipes and carbonate replacement deposits are unrelated to the Turnley Ridge porphyry deposit (Brown and others, 2019).

Given that the distinct alteration assemblages of porphyry mineral systems can be expected to display varying concentrations of the radioelements (Dickson and Scott, 1997; Hoshcke, 2008; Sillitoe, 2010), we further investigate how previously mapped alteration assemblages in the Elkhorn district are expressed in maps of radioelement concentrations, ratios of potassium to thorium and uranium, and data enhancement maps as well as handheld GRS measurements (figs. 3C–3H). The airborne radioelement concentration maps and corresponding radioelement concentrations from handheld GRS measurements exhibit broadly similar values and trends across the Elkhorn district. Both airborne radiometric and handheld GRS data show elevated potassium values in the Butte Granite (following Lund and others, 2002) and Turnley Ridge stock, with lower concentrations over the Black Butte and Cemetery Ridge diorite stocks (fig. 3C). Conversely, both thorium and uranium concentrations are lower over the Elkhorn district and higher over the Butte Granite (figs. 3D, 3E). Sedimentary rocks show relatively low values in all three radioelement maps, whereas Elkhorn Mountains Volcanics exposed to the southwest show both high and low values. The radiometric data do not show any clear signatures corresponding to skarn mineralization in the Elkhorn district, which may be due to the limited area of skarn exposure at the surface.

K/Th, K/U, and K enhancement maps clearly highlight the core of the Turnley Ridge porphyry system. Handheld GRS measurements from mapped areas of K-feldspar-biotite and quartz-sericite alteration assemblages show elevated K/Th (fig. 3F), but only slightly elevated K/U (fig. 3G). Conversely, in the airborne radiometric data, potassium-uranium ratios are greater across these alteration assemblages than potassium-thorium ratios. However, comparison with the surrounding area shows that potassium enrichment relative to thorium is still more pronounced within the core of the porphyry system. The potassium enhancement map (fig. 3H) shows in white where all three bands (K ppm, K/Th, K/U) have elevated values, again highlighting the mapped K-feldspar-biotite and quartz-sericite alteration assemblages. In contrast, the radiometric data do not show consistent trends over the areas of propylitic alteration. A simple cutoff of 1.0 applied to the K–Th ratios of GRS measurements appears to neatly highlight the intensely altered central zones of the porphyry system (fig. 3H).

The above observations can be interpreted in the context of the expected radioelement concentrations associated with mineral and alteration zonation across porphyry mineral systems (e.g., Davis and Guilbert, 1973; Irvine and

Smith, 1990). K-rich minerals expected in potassic alteration zones at the core of porphyry systems include K-feldspar [KAlSi_3O_8] and biotite [$\text{K}(\text{Mg},\text{Fe})_3\text{AlSi}_3\text{O}_{10}(\text{F},\text{OH})_2$], whereas sericite [$\text{KAl}_2(\text{AlSi}_3\text{O}_{10})(\text{OH})_2$] and illite [$(\text{K},\text{H}_3\text{O})(\text{Al},\text{Mg},\text{Fe})_2(\text{Si},\text{Al})_4\text{O}_{10}[(\text{OH})_2,(\text{H}_2\text{O})]$] are expected in the sericitic and chlorite-sericite alteration zones (Sillitoe, 2010). In the advanced argillic alteration zone of the distal portions of porphyry systems, K-rich minerals may include alunite [$\text{KAl}_3(\text{SO}_4)_2(\text{OH})_6$]. However, radioelement concentrations can depend on silicification whereby moderate amounts of silica addition decrease potassium and retain thorium, but massive silicification can result in very low concentrations of all three radioelements (Dickson and Scott, 1997). In the nearby Butte porphyry deposit, for example, alteration envelopes of the potassic series, which include sericitic alteration, characteristically contain added K on an absolute volume basis (Reed and Dilles, 2020). Thus, it is likely that elevated K–Th or K–U values would highlight areas of K enrichment within a porphyry deposit exhumed to depths below the argillic alteration zone, as observed around the Turnley Ridge stock. Furthermore, the correspondence of the airborne radiometric data with handheld GRS measurements observed in the Elkhorn district suggests that handheld GRS may be a useful screening tool even in the absence of high-resolution airborne surveys.

Clementine Prospect Magnetic Structure

The Clementine prospect, located ~10 km northwest of Divide, MT, has received little attention for mineral exploration due to rugged topography, dense vegetation, and limited road access (fig. 2). The prospect is located at the broad intersection of the Big Hole Canyon pluton to the south and Butte Granite to the north (fig. 4A), where recent mapping has revealed evidence of hydrothermal alteration and local Cu–W–As–Sb breccia vein gossan (Brimhall, 2018). The area is dominated by intermediate to felsic Cretaceous plutons, including the ca. 86–74 Ma Big Hole Canyon pluton and the ca. 77–74 Ma Butte Granite of the Boulder batholith, which intrude Proterozoic to Mesozoic carbonate and siliciclastic strata (fig. 4A; Zinter and others, 1983; du Bray and others, 2012; McDonald and others, 2012). The main alteration assemblage at the prospect consists of actinolite-titanite-calcite orbs and disseminated chalcopyrite (Brimhall and Fanning, 2019), similar to alteration documented in skarn at the Carr Fork area, Bingham, UT (Atkinson and Einaudi, 1978). These alteration zones are distal to breccia vein gossan and are interpreted to represent the alteration halo of a buried porphyry copper system emplaced into sedimentary rocks (Brimhall, 2023). U–Pb analyses of hydrothermal titanite provide ca. 73–71 Ma interpreted mineralization ages, younger than both the Butte Granite and Big Hole Canyon pluton (Brimhall and Fanning, 2019). Thus, neither pluton is the source of metals nor fluids in the mineral system and instead, mineralization



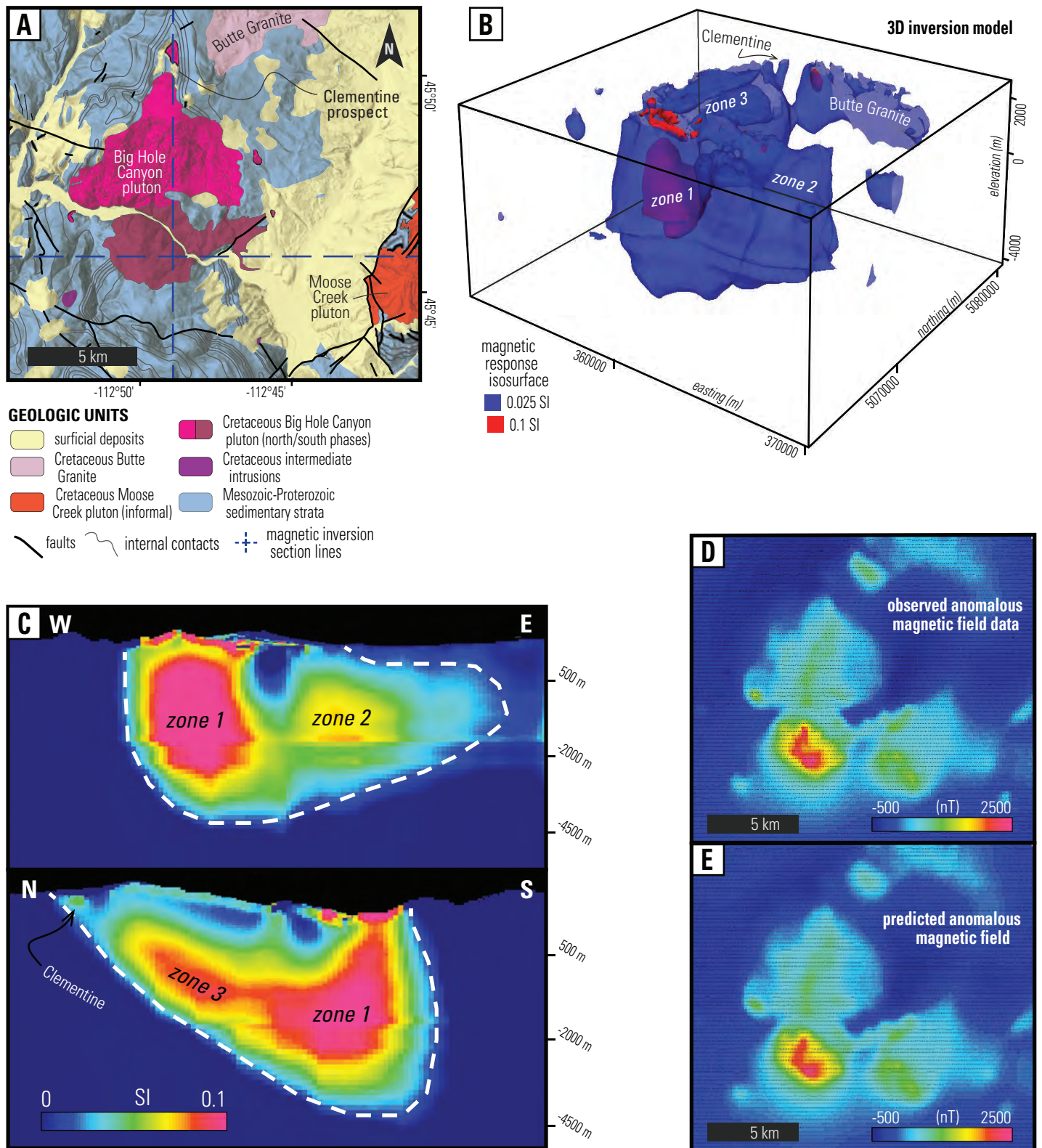


Figure 4. Geology, aeromagnetic survey data, and modeling results of the Clementine prospect and Big Hole Canyon pluton, modified from Magnin and others (2025). (A) Geologic map of Clementine prospect used for near-surface magnetic susceptibility model. Internal contacts illustrate structure of sedimentary units. Geology from Hesperheide (2003), McDonald and others (2012), and Brimhall and Fanning (2019). Datum: WGS84. (B) Preferred magnetic susceptibility model from geophysical inversion, with horizontal extent same as map in A. (C) Predicted magnetic susceptibility along east–west and north–south cross-sections shown on map in A. (D) Observed anomalous magnetic field data (Anderson and others, 2025a) compared to (E) synthetic signal predicted by preferred 3D inversion model, extent same as A.

is hypothesized to be related to a younger intrusion inferred at depth (Brimhall, 2023).

To further investigate the architecture of intrusions associated with alteration at the Clementine prospect and shed light on the hypothesized source intrusion at depth, aeromagnetic data were combined with surface observations and geologic context to model the subsurface magnetic structure of the area (fig. 4). Magnetic susceptibility models were obtained through a series of inversions with differing data misfit values (i.e., how closely predictions match the observed data). The chosen preferred model reproduced all observed anomalies, displayed relatively low data misfit, and yielded results consistent with known geology (fig. 4B). Vertical slices of the preferred 3D model show three zones of high susceptibility (figs. 4B, 4C). Zone 1 is a subvertical body that underlies the southern portion of the Big Hole Canyon pluton (figs. 4A, 4C), has the highest susceptibility values in the model, and corresponds with the most magnetite-rich rocks at the surface (Hespenheide, 2003). The second zone of high susceptibility extends to the east under upper Cenozoic unconsolidated sediment and inferred Phanerozoic siliciclastic and carbonate strata at depth (figs. 4A, 4C). This feature corresponds with isolated exposures of granodiorite (McDonald and others, 2012) and is likely buried plutonic rocks that may be related to the Big Hole Canyon pluton. Although the economic potential of this buried feature is unknown, it is spatially associated with Paleozoic and Mesozoic carbonate-bearing units at or near the surface that contain W-bearing skarn prospects (Andersen and others, 2023). The third high-susceptibility zone extends and thins to the north, corresponding with the northern portions of the exposed Big Hole Canyon pluton (figs. 4A, 4C; Hespenheide, 2003). A south-plunging magnetic protrusion extends off the northern margin of the buried Big Hole Canyon pluton, where detailed mapping reveals two small granitic intrusions just north of the Big Hole Canyon pluton along the axis of a north-plunging anticline (Brimhall, 2018). The northern of the two small intrusions is altered, with a U–Pb zircon crystallization age of 77 ± 1 Ma and a U–Pb titanite alteration age of 73 ± 1 Ma (Brimhall, 2023). The southern, undated body intrudes barren hydrothermal breccia and is unaltered, suggesting it was emplaced post-mineralization.

Due to likely magnetite destruction from hydrothermal alteration in the northern intrusion, the southern intrusion is interpreted to represent the shallow magnetic feature in the model. Its modeled subsurface geometry and slightly lower magnetic susceptibility than the rest of the Big Hole Canyon pluton (fig. 4C) could be consistent with an interpretation that the granite is an evolved, late-stage differentiate from the Big Hole Canyon pluton. In this instance, the Big Hole Canyon pluton would have still been cooling at the time of hydrothermal alteration and mineralization at the Clementine prospect, indicating it may have been one of

several potential heat sources. No other high-susceptibility feature that may represent a buried syn-mineralization intrusion is indicated from geophysical modelling. Although not recognized in the model, such an intrusion may still exist if it is not magnetic or any pre-existing magnetite was subsequently destroyed by oxidizing fluids. A further evaluation of potential sources of the fluids and metals at the Clementine prospect will be conducted by integrating these inversion results with stable isotope analyses and geochronology (Magnin and others, 2025).

Winston District Reconnaissance

The Winston mining district in the northeastern Elkhorn Mountains (fig. 2) consists of fault- and fracture-related, base- and precious-metal and Au–Ag–quartz–pyrite veins (Klepper and others, 1971). In this part of the Elkhorn Mountains, gently dipping strata of the Upper Cretaceous Slim Sam Formation and Elkhorn Mountains Volcanics are intruded by Late Cretaceous mafic to intermediate hypabyssal intrusions and quartz monzonite to granite porphyry stocks (fig. 5A; Freeman and others, 1958; Klepper and others, 1971). Faulting postdates most intrusions with some activity continuing into Cenozoic time (Reynolds and Brandt, 2006). Vein-hosted deposits are closely associated with Late Cretaceous felsic stocks and are hosted in the Elkhorn Mountains Volcanics and the intrusions themselves (Klepper and others, 1971). Most past mining activity was focused near porphyry stocks along the NNE-trending Weasel Creek fault (fig. 5A), but the relationship between these felsic stocks and the metal endowment of the epithermal veins has not been studied.

We conducted preliminary reconnaissance to ground-truth geophysical survey data and more closely examined mineralization and deformation in this area of concentrated felsic porphyry stocks. On the western edge of the Winston district, the poorly exposed, informally named Pole Creek stock is mapped as a 3-km-long quartz monzonite porphyry near the base of the Elkhorn Mountains Volcanics and has only been described as similar to the nearby informally named Edna stock (fig. 5A; Klepper and others, 1971). Visiting several small outcrops of the Pole Creek stock, we observed felsic porphyry affected by pervasive sericitic alteration and cut by extensive quartz–feldspar stockwork veins and many small faults and fractures. Handheld magnetic susceptibility measurements show uniformly low values (Anderson and others, 2025b), consistent with expected magnetite destruction from observed sericite alteration (e.g., Gunn and Dentith, 1997; Hoschke, 2008).

These preliminary observations prompted closer examination of the airborne geophysical data acquired over the northern Winston district and comparison of the geophysical patterns around the small porphyry stocks related to mineralization (Allen Langhans and others, 2024). Airborne radiometric data reveal higher potassium concentrations



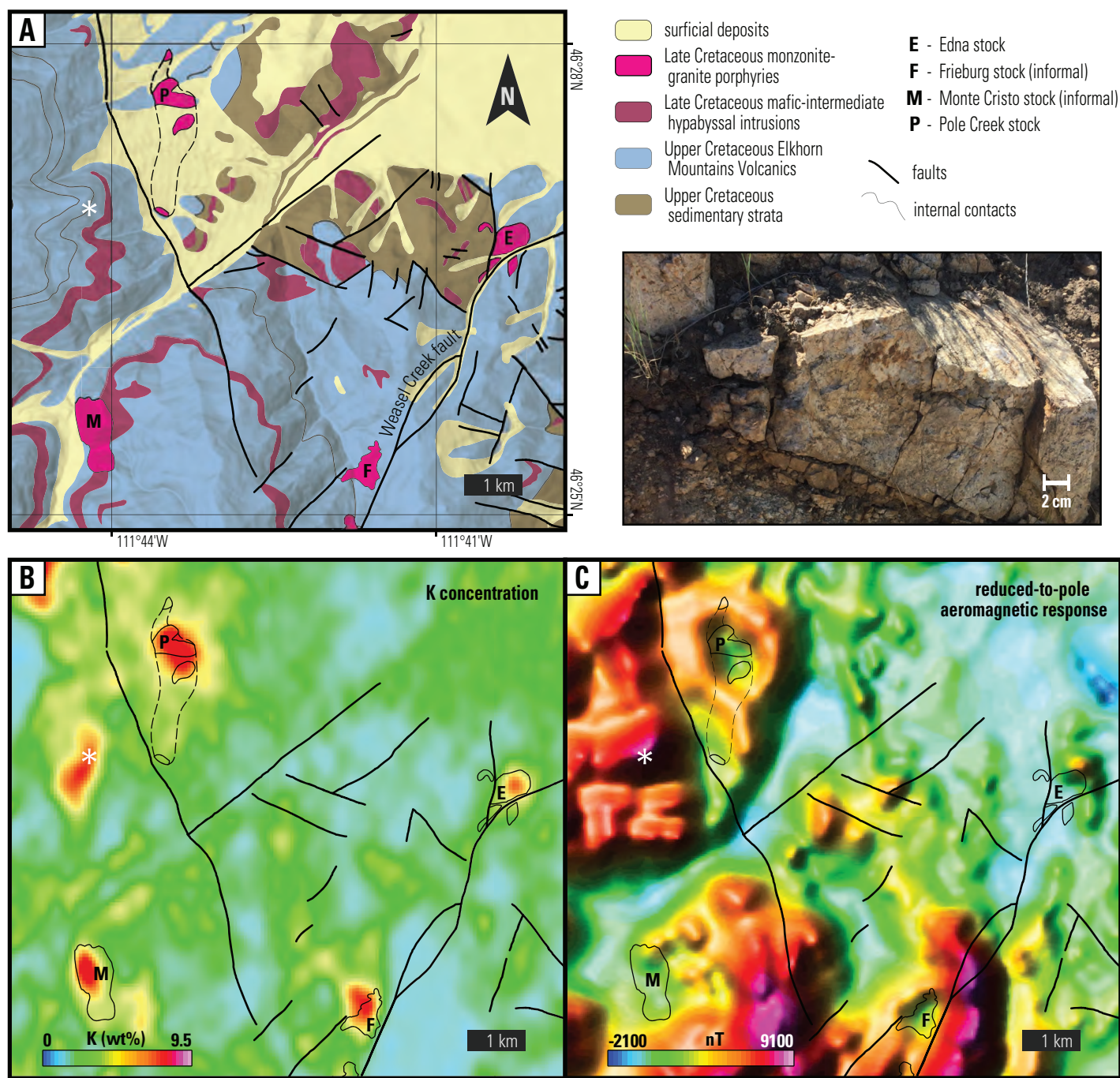


Figure 5. Geology and geophysical survey data of the Winston mining district. (A) Simplified geology of the Winston mining district modified from Reynolds and Brandt (2006); internal contacts of Elkhorn Mountains Volcanics shown to highlight gently dipping strata, datum: WGS84. (B) Potassium concentration and (C) reduced to pole aeromagnetic response maps (Allen Langhans and others, 2024), shown with simplified faults and same extent as A. White star denotes geophysical anomalies discussed in text.

over mapped stocks, with few other consistent signatures in radiometric element concentrations or ratios (fig. 5B). The small stocks also correspond to lows in magnetic response collocated with the areas of elevated potassium and are ringed by regions of elevated magnetic response (fig. 5C). The magnetic lows and potassium enrichment may possibly represent portions of the stocks where magnetite has been destroyed, with the peripheral elevated magnetic response corresponding to secondary growth of magnetite or pyrrhotite at a reduction front. Combined with the observations of quartz–feldspar stockwork, pervasive alteration, and secondary sulfide minerals, this may speculatively indicate a set of small porphyry mineral systems associated with the Winston district felsic stocks. Alternatively, these stocks may simply have high primary potassium concentrations with subsequent magnetite destruction by oxidizing fluids during alteration. A high-potassium anomaly southwest of the Pole Creek stock is collocated with a magnetic high, denoted by a white star in figure 5. Whereas the high potassium concentration corresponds to a specific tuff mapped within the Elkhorn Mountains Volcanics, the magnetic high and others just to the north and south follow steep ridgelines and appear to result from topographic effects on the airborne magnetic results. We plan to target this area for further fieldwork to test these hypotheses and further characterize its magmatic, structural, and metallogenic framework.

Buried Bison Creek Magnetic Anomaly

The Bison Creek area, in the northwestern portion of the Boulder Mountains along the continental divide (fig. 2), consists of Late Cretaceous Butte Granite intruding Upper Cretaceous Elkhorn Mountains Volcanics, both of which were later dissected by extensional faults and partially covered by Eocene Lowland Creek Volcanics as well as Eocene–Oligocene topaz rhyolite tuffs and lavas associated with the Elliston volcanic field (fig. 6A; Ruppel, 1963; Tilling, 1973; Christiansen and others, 1986; Dudas and others, 2010; Scarberry and others, 2018, 2019). Mining districts in the area include the Basin, Clancy, Elliston, and Rimini districts, which consist of Cretaceous and younger polymetallic and epithermal vein deposits and minor porphyry and shear zone related deposits (Elliott, 1996). Mines in these districts produced gold, silver, lead, zinc, and copper with additional known geochemical anomalies of antimony, arsenic, barium, beryllium, bismuth, boron, manganese, molybdenum, tin, and tungsten (Elliott, 1996).

At the boundary of the Basin, Elliston, and Rimini districts is a circular, dipolar magnetic anomaly with the strongest magnetic response (up to ~8,300 nT) of the entire survey area (fig. 6B). The anomaly is coincident with Eocene–Oligocene banded rhyolite flows of the Elliston volcanic field. These volcanic rocks show low magnetic susceptibilities, indicating that a different magnetic source

sits at depth (table 1). The Bison Creek anomaly was previously identified from a USGS aeromagnetic survey collected in the 1990s and was interpreted as a vent complex with a large intrusion at shallow depth (McCafferty and others, 2004). Initial blind magnetic inversions using magnetic amplitude data (fig. 6C), however, indicate that subsurface magnetic material have susceptibility values on the order of 1–2 SI, which approaches the scale of massive magnetite mineralization. Additionally, mafic intrusions elsewhere in the Boulder batholith region do not have such high susceptibilities and typically produce magnetic anomalies on the order of only ~500 nT. Alternative explanations to produce such an anomaly include a Late Cretaceous to late Eocene magnetite-rich iron skarn, kimberlite pipe, or ultramafic intrusion of different character than the rest of the mafic intrusions in the Boulder batholith region (e.g., alkaline mafic intrusions of the Magnet Cove Complex; Amaral and others, 2024). These potential models are all geologically reasonable considering the prolific and diverse skarn mineralization in the region as well as occurrences of both Late Cretaceous to Eocene kimberlite and alkaline magmatism to the west and northeast such as the Eocene kimberlites found in the Missouri Breaks portion of the Central Montana alkalic province and Late Cretaceous alkalic rocks within the Montana–Idaho alkalic belt (Heaman and others, 2004; Duke, 2009; Gammons, 2020; Gammons and others, 2020).

We test the different preliminary hypotheses with 3D magnetic inversion using different geologic reference models. Igneous models (i.e., kimberlite pipe or ultramafic complex) assume that the intrusion crosscuts the underlying Butte Granite and were tested using a near-surface (75 m depth) reference model based on the thickness and magnetic susceptibility values in table 1. All areas outside this model were allowed to vary between 0 and 2 SI, essentially constraining the upper intrusive contact but not the lower. The hydrothermal model (i.e., skarn) assumes that magnetite mineralization is constrained between the Eocene rhyolite at the surface and underlying Butte Granite. A reference model was created using the near-surface susceptibility model previously described, while also requiring Butte Granite and its physical properties at depths below 1,643 m in elevation, corresponding to 500 m below the average elevation of the model area.

Preferred inversion results constrained by both models (figs. 6C, 6D) were selected based on whether they reproduced all the major anomalies in the observed data, had a low data misfit, and were geologically reasonable. For example, models with the lowest data misfit made the entire subsurface highly magnetic and were highly unrealistic. Results from both models have areas of high susceptibility (1–2 SI) in the shallow subsurface, suggesting that there are rocks with a high magnetite content directly beneath exposed rhyolite. This is consistent with the short-wavelength



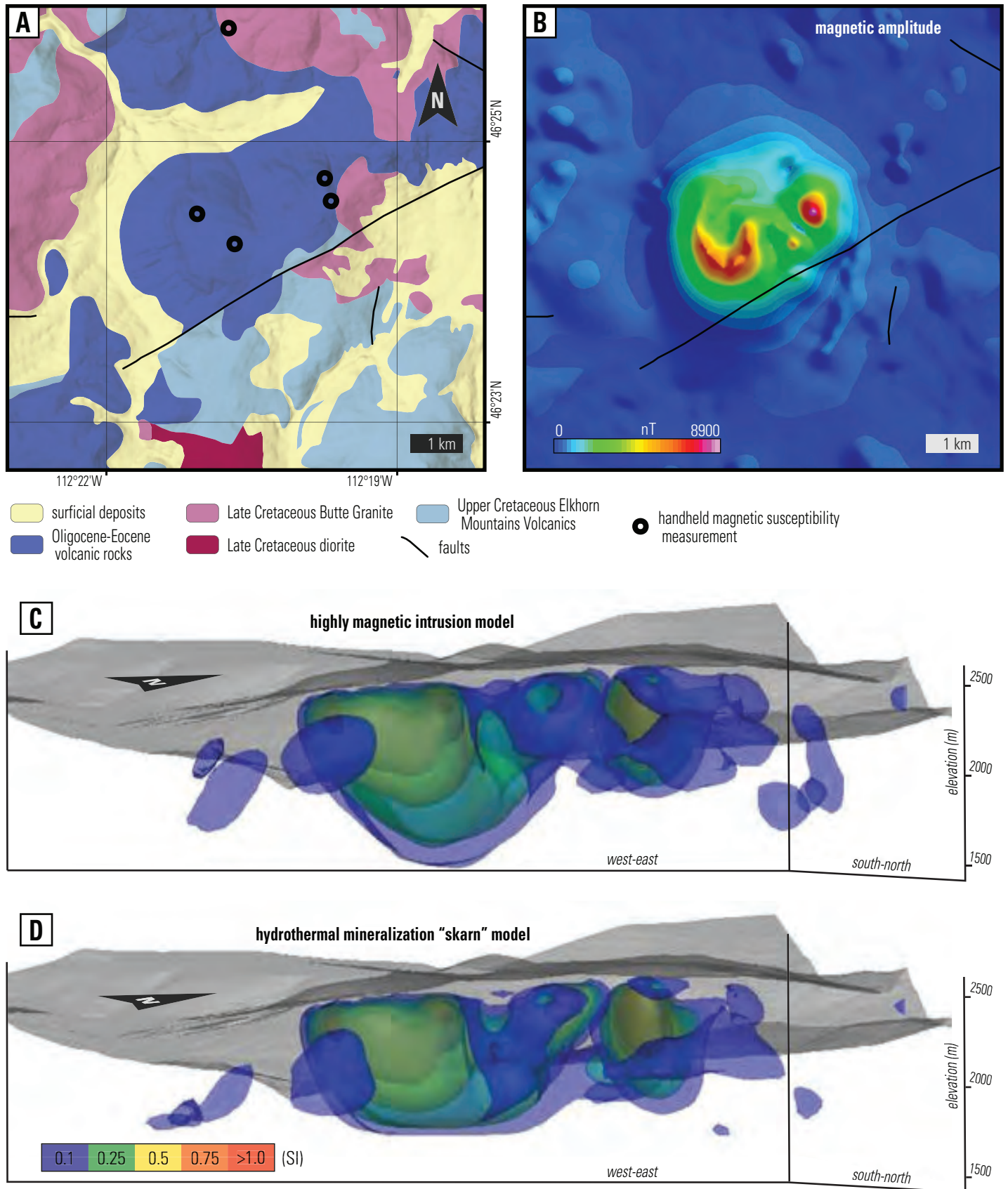


Figure 6. Geology, aeromagnetic survey data, and 3D magnetic inversion results of the Bison Creek magnetic anomaly. (A) Simplified geology of the Bison Creek area above magnetic anomaly, modified from Scarberry and others (2019); datum: WGS84. (B) Magnetic amplitude from aeromagnetic survey data (Allen Langhans and others, 2024), same extent as A. 3D magnetic inversion subsurface models following hypotheses of buried (C) highly magnetic intrusion and (D) concentrated, highly magnetic hydrothermal mineralization zone; extents of model areas same as A.

variations within the magnetic anomaly (fig. 6B), suggesting that the highly magnetic source cannot be deeply buried (c.f., Blakely, 1995). Using a cutoff of 0.1 SI as a guide to the overall magnetic geometry of the subsurface feature, both models suggest a thin structure of <1 km (figs. 6C, 6D). A thin magnetic body perched above the Butte Granite is slightly more consistent with a skarn or hydrothermal model. However, any potential subsurface ultramafic intrusion may be either thin or composed of materials with variable magnetic susceptibility. Future work to further refine the subsurface model will focus on ground gravity data collection, which should distinguish a dense ultramafic intrusion from the low-density rhyolites regardless of its magnetite distribution. Additional exploration for indicator minerals or evidence of hydrothermal alteration coincident with the anomaly in the case of a kimberlite or skarn is also planned. It is likely that rocks with significant magnetite content lie in the shallow subsurface, and these may potentially also host elevated concentrations of critical minerals (e.g., Meinert and others, 2005; Kelley and others, 2017).

SUMMARY

Integrated investigations using Earth MRI airborne geophysical data contribute to a refined understanding of the mineral systems surrounding the Late Cretaceous Boulder batholith. Measurements of magnetic susceptibility and concentrations of potassium, thorium, and uranium using handheld instruments provide ground-truthing of the airborne data. In the Elkhorn district, the radiometric data highlight alteration zones associated with a porphyry copper system, which may be a useful indicator to apply in other areas. At the Clementine prospect, 3D modeling of the magnetic structure of the Big Hole Canyon pluton suggests that potential late stages of its construction postdate mineralization, providing a framework for further hypothesis testing. Felsic intrusions in the Winston district exhibit magnetic lows and elevated K concentrations. Combined with field observations of K-feldspar-quartz stockwork and alteration, this suggests that several kilometer-scale porphyry mineral systems may exist in the area with uncertain relationship to the subsequent fault- and fracture-related vein deposits that have been previously mined. The buried Bison Creek magnetic anomaly, the largest anomaly observed in the two airborne surveys, corresponds to highly magnetic material in the subsurface that may represent either an intrusion or magnetite-rich mineralization. These observations and hypotheses are being further investigated with ongoing geologic, geochemical, geochronologic, and geophysical work.

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THE GRAVITY AND MAGNETIC SIGNATURE OF CRETACEOUS OROGENY—TECTONIC INVERSION OF PRECAMBRIAN RIFTING STRUCTURES IN THE PIONEER MOUNTAINS OF SOUTHWESTERN MONTANA

Kevin Lielke

Independent Geologist, Ennis, Montana

ABSTRACT

Legacy gravity and aeromagnetic data are a valuable, underutilized resource for examining crustal structure and testing proposed models for the tectonic deformation of western North America. In the area centered around the Pioneer Mountains of southwestern Montana, analysis employing gradients of gravity and aeromagnetic data show an enhanced ability to resolve faults, igneous intrusions, and other tectonic features. Based on attribute maps of the gravity and aeromagnetic fields as well as forward modeling of the vertical component ($\partial f/\partial z$) of the gravity field, a general model for the Cretaceous tectonic evolution of the area is proposed. This model involves the tectonic inversion of normal faults formed during late Proterozoic rifting of the North American continent. The influence of Precambrian rifting on the subsequent tectonic evolution of western North America requires serious consideration in future regional deformation models.

INTRODUCTION

During Cretaceous orogeny, the area of the Pioneer Mountains occupied the forward propagating edge of the Sevier fold and thrust belt (Ruppel and Lopez, 1984; Ruppel, 1993). This largely thin-skinned deformation belt was heavily influenced by preexisting tectonic features, most of which have a Precambrian ancestry (Garihan and others, 1983; Harms and Baldwin, 2021; Ruppel and Lopez, 1984; Schmidt and Garihan, 1983). An additional complication arises from the observation that Cretaceous magmatism often employed fault zones as emplacement conduits (Foster and others, 2007; Kalakay and others, 2000, 2001). This convoluted deformation history resulted in complex surface geology that is difficult to disentangle and interpret (fig. 1). Geophysical methods have the potential to help make sense of the clutter of the surface geology. In particular, gravity and aeromagnetic data collected during the middle to late 20th century are an underutilized resource for understanding the geologic history of the region. Processing techniques and computer capabilities have changed dramatically since the time these datasets were first collected, giving new life to old data (Dentith and Mudge, 2018; LaFehr and Nabighian, 2012). This study reinterprets legacy gravity and aeromagnetic data using modern processing and visualization techniques in order to clarify the complex geologic history of the region.

REGIONAL GEOLOGY

The tectonic history of the region can be broken down into five broad divisions. During Archean/Paleoproterozoic time, this area was the site of primarily compressional deformation related to the assembly of the northwestern margin of the North American craton (Harms and Baldwin, 2021, 2023; Mogk and others, 2020). Predominately northwest- and northeast-oriented structures were active during this period of orogenesis and terrane amalgamation (Garihan and others, 1983; Mogk and others, 2020; Sims and others, 2004; Schmidt and Garihan, 1983, Sumner and Erslav, 1988). During late Precambrian and into early Cambrian time, the western edge of Laurasia was subject to widespread extension that culminated in continental break-up and the establishment of a passive margin bordering the open ocean. The details of this event are obscure, but it seems likely that extensional fault systems paralleling the new coastline were involved. Extensional fault segments were probably linked by oblique or perpendicularly oriented transfer faults, as is observed in most ancient and modern rift systems (Druet and others, 2018; Grauch and Ruleman, 2013; Shah and others, 2023). Proterozoic rifting in southwest Montana was influenced by the tectonic framework inherited from Archean compressional deformation, a circumstance also seen in other rift basins (Drenth and others, 2013; Druet and others, 2018).

The passive margin whose beginning is marked by Cambrian transgression lasted throughout the early part of Phanerozoic time (Peterson, 1986; Thomas and Roberts, 2007). However, it was not a single homogenous depositional system. Differential subsidence and relict paleotopography led to differences in sedimentary depositional systems and thicknesses throughout the Paleozoic sedimentary record (Peterson, 1986). These discontinuities continued to manifest themselves during the early stages of Mesozoic compressional tectonics (Cobban and others, 1994; Gibson, 2007; Schwartz and DeCelles, 1988; Vuke, 1984). Continuing into the main phase of Cretaceous orogenesis, preexisting structures guided the emplacement of thrust sheets and contemporary igneous activity and largely determined the structural style of the resulting deformation (Parker and Pearson, 2021; Ruppel and Lopez, 1984; Schmidt and others, 1988; Vasey and others, 2024). Finally, during Cenozoic extension, the location of normal fault systems was in

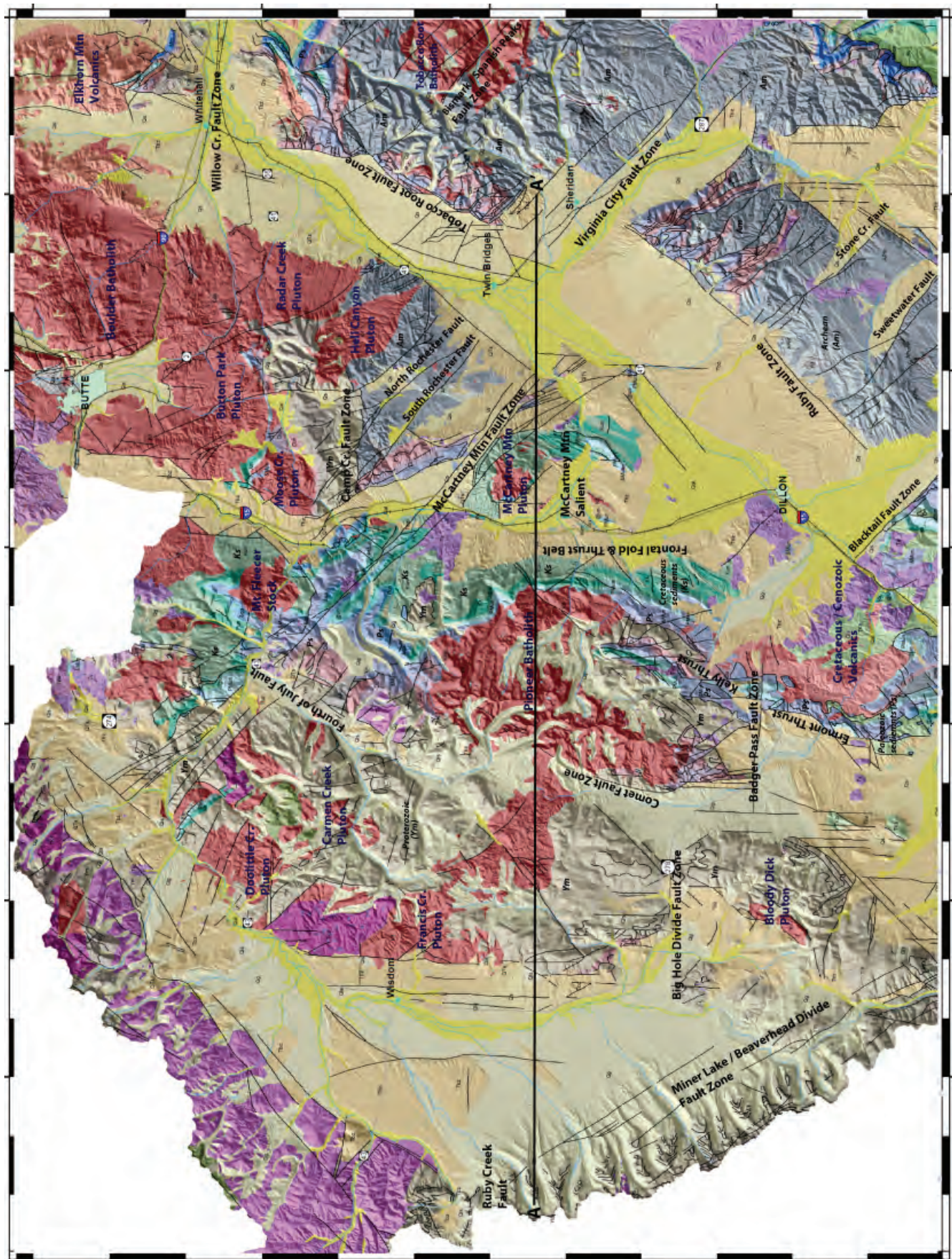


Figure 1. Simplified geologic map of southwestern Montana. Prominent tectonic features and igneous bodies are labeled. Line A to A' corresponds to line of gravity model in figure 6. Modified from Portner (2003).

large measure determined by the location of thrust ramps, high topography, and overthickened crustal roots formed during the prior episode of compressional tectonics (Lielke, 2017; Ruppel, 1993).

DATA AND METHODS

The primary datasets consist of approximately 1,400 surface gravity datapoints (Hanna and others, 1993; Hassemer, 1981; Kaufman and others, 1983; Mankinen and others, 2004; USGS online database, 2024) supplemented by gridded aeromagnetic surveys (USGS online database, 2024). These data were collected primarily by the USGS during the latter part of the twentieth century. Gravity data were originally reduced by the USGS using a standard array of processes including free air, Bouguer and terrain corrections. For this study, gravity residuals were calculated from the USGS Bouguer data by subtracting the regional trend in the data using a weighted least-squares regression, thereby enhancing the upper crustal component of the data (LaFehr and Nabighian, 2012; Lielke, 2023). Since detecting the boundaries of crustal structures is the most important goal of this study, the gradients (or in the language of calculus, the partial derivatives) of the potential field were calculated. Since gravity anomalies record contrasts in mass/density between geologic bodies, sharp discontinuities—steep gradients—in the data likely correspond to abrupt changes in geology (Dentith and Mudge, 2018; LaFehr and Nabighian, 2012). Gravity gradients therefore reflect the edges and shapes of sources rather than just mass distribution (Bell, 1997). This same principle holds true for contrasts in magnetic properties (Bates and others, 2024). Steep gradients in a gravity or magnetic field are therefore “edge finders” that should highlight faults, intrusive contacts, and other geologic features. The first and second partial derivatives of the potential field data were calculated along with several interpretation formulas that are combinations of the various partial derivatives (table 1). Comparisons of the generated image maps with the mapped surface geology were made to determine which of these processing techniques were most useful in illuminating the regional crustal structure.

In addition to the generation of plan view maps, a profile of the gravity vertical derivative ($\partial f/\partial z$) was calculated and a forward model was developed in an attempt to match the calculated $\partial f/\partial z$ curve of the model to the observed data. The vertical derivative was used instead of the raw Bouguer or residual data since the vertical component of the gradient field is more sensitive to the actual position of the boundary between geologic bodies than the gravity field itself (Blakely, 1995; Dentith and Mudge, 2018). Published cross-sections and geologic interpretations were used as the initial foundation of the model along with reported density data from prior studies (Hanna and others, 1993). The model was then modified until a satisfactory fit was established between the observed and calculated gravity gradients within the context of the known surface geology. An error of 5% or less was considered sufficient to consider the model as geologically viable.

INTERPRETATIONS

It was determined that a combination of processing techniques, interpreted in tandem, was most effective in examining crustal structure. Gradient methods due to their sensitivity to contrasts in the potential field proved to be effective “edge finders” for outlining faults and other geologic discontinuities. However, no one method is universally applicable, especially in an area of highly complex geology. Employing a suite of techniques, tailored to the specific study area, gives the best results. However, it should be kept in mind that interpretations of potential field data are non-unique and therefore can only test, not prove, hypotheses.

Aeromagnetic Data

Within the study area, highly magnetic rock bodies fall into two main categories: igneous rocks of Cretaceous to Cenozoic age and metamorphic rocks of Precambrian age (Hanna and others, 1993). The majority of magnetic igneous rocks are Cretaceous plutons, while the most magnetic metamorphic lithologies are either mafic/ultramafic units or metasedimentary banded iron formations (Hanna and others, 1993). However, several factors not present in gravity data complicate the interpretation of magnetic data. The most important is the dipolar nature of magnetism.

Table 1. Mathematical formula for calculating gravity and magnetic attributes.

Method	Equation
Vertical Derivative	$\partial f/\partial z$
Horizontal Derivatives	$\partial f/\partial x \quad \partial f/\partial y$
Horizontal Gradient (HDR)	$((\partial f/\partial x)^2 + (\partial f/\partial y)^2)^{1/2}$
Tilt Angle (TA)	$\tan^{-1} \{ \partial f/\partial z / ((\partial f/\partial x)^2 + (\partial f/\partial y)^2)^{1/2} \}$
Curvature	$((\partial^2 f/\partial x^2 - \partial^2 f/\partial y^2)^2 + (2 * \partial^2 f/\partial x \partial y)^2)^{1/2}$



The Earth's magnetic field is vertical only at the poles; at mid-latitudes, the field is inclined, and the resulting anomaly consists of distinct positive and negative components not necessarily centered above the source of the anomaly (Dentith and Mudge, 2018). It is possible to correct for this effect, a process generally referred to as “reduction to the pole,” providing the measured rock magnetism is primarily induced by the present-day magnetic field (Bates and others, 2024, Blakely, 1995). However, if a significant component of remanent magnetism exists, that is, magnetism caused by “locked” magnetic grains oriented to an earlier configuration of the Earth's magnetic field, then this technique will be less effective (Blakely, 1995; Dentith and Mudge, 2018). Processing techniques employing magnetic gradients are less sensitive to the type of magnetism present and are more effective at delimiting the boundaries of the sources of the magnetic anomalies. Methods employing partial derivatives are, however, more sensitive to high-frequency noise and acquisition effects compared to the raw total field data (Dentith and Mudge, 2018).

In the study area, the presence of remnant magnetism and complex, multi-stage igneous intrusions leads to complicated aeromagnetic signatures as can be seen in the magnetic residual map (fig. 2, top). However, maps of the vertical component ($\partial f/\partial z$), the total horizontal derivative (HDR), and the curvature attribute (table 1) of the magnetic field greatly simplify the resulting anomalies (figs. 2, 3) and serve to highlight the boundaries of the source rock bodies (Saibi and others, 2019). An example can be seen in the anomaly caused by the McCartney Mountain pluton. The magnetic residual is a NE/SW elongated feature consisting of a coupled positive and negative component (fig. 2, top). The vertical derivative, HDR, and curvature responses collapse the original anomaly to a more compact feature centered over the area of surface exposure of the McCartney Mountain pluton. A progressive refinement of the shape and location of the source bodies throughout the study area is in fact evident (fig. 2, 3). Within the study area, a generally good correspondence between the mag-

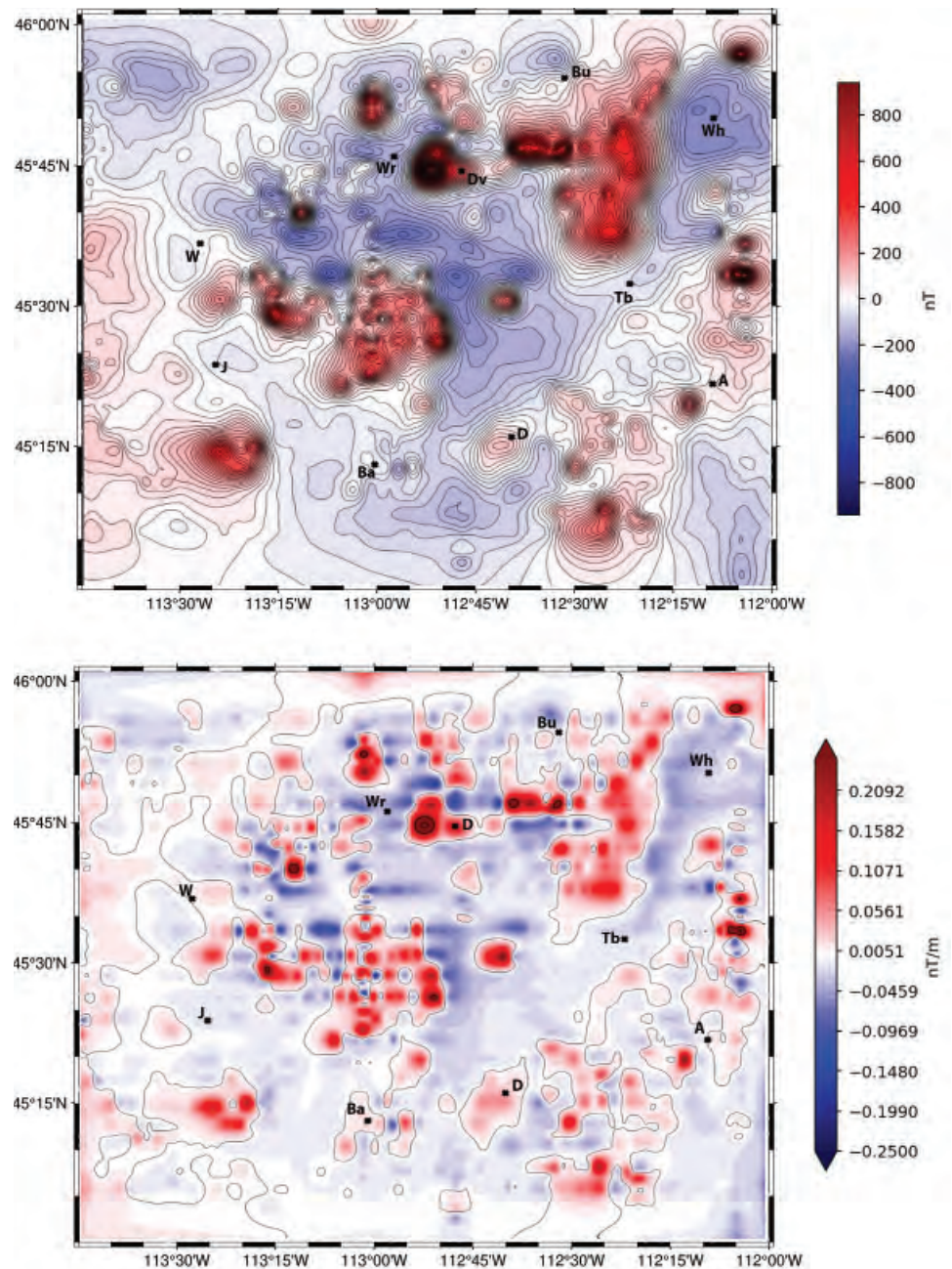


Figure 2. (Top) Map of aeromagnetic residuals. Note complex anomaly patterns characterized by paired positive and negative features. (Bottom) Vertical component of magnetic field. Compared to residual map, note simplified response shifted towards the location of the source of the anomaly. The zero contour is shown, which ideally corresponds to the edge of the source. Localities on map: A, Alder; Ba, Ban-nack; Bu, Butte; D, Dillon; Dv, Divide; J, Jackson; Tb, Twin Bridges; W, Wisdom; Wh, Whitehall; Wr, Wise River.

netic attribute responses and the surface outcrop of igneous bodies is also observed (fig. 1, fig. 3 bottom). Gradient techniques clarify the location of probable buried igneous bodies, such as the prominent anomaly located south of Wisdom, and the subsurface extensions of mapped surface intrusions. In the southeastern portion of the study area, prominent anomalies appear to correspond to the mapped extent of mafic/ultramafic metamorphic rocks and banded iron formation in the Ruby Range (James, 1990; Karasevich, 1981). The general position and orientation of the frontal fold and thrust belt are indicated by the presence of

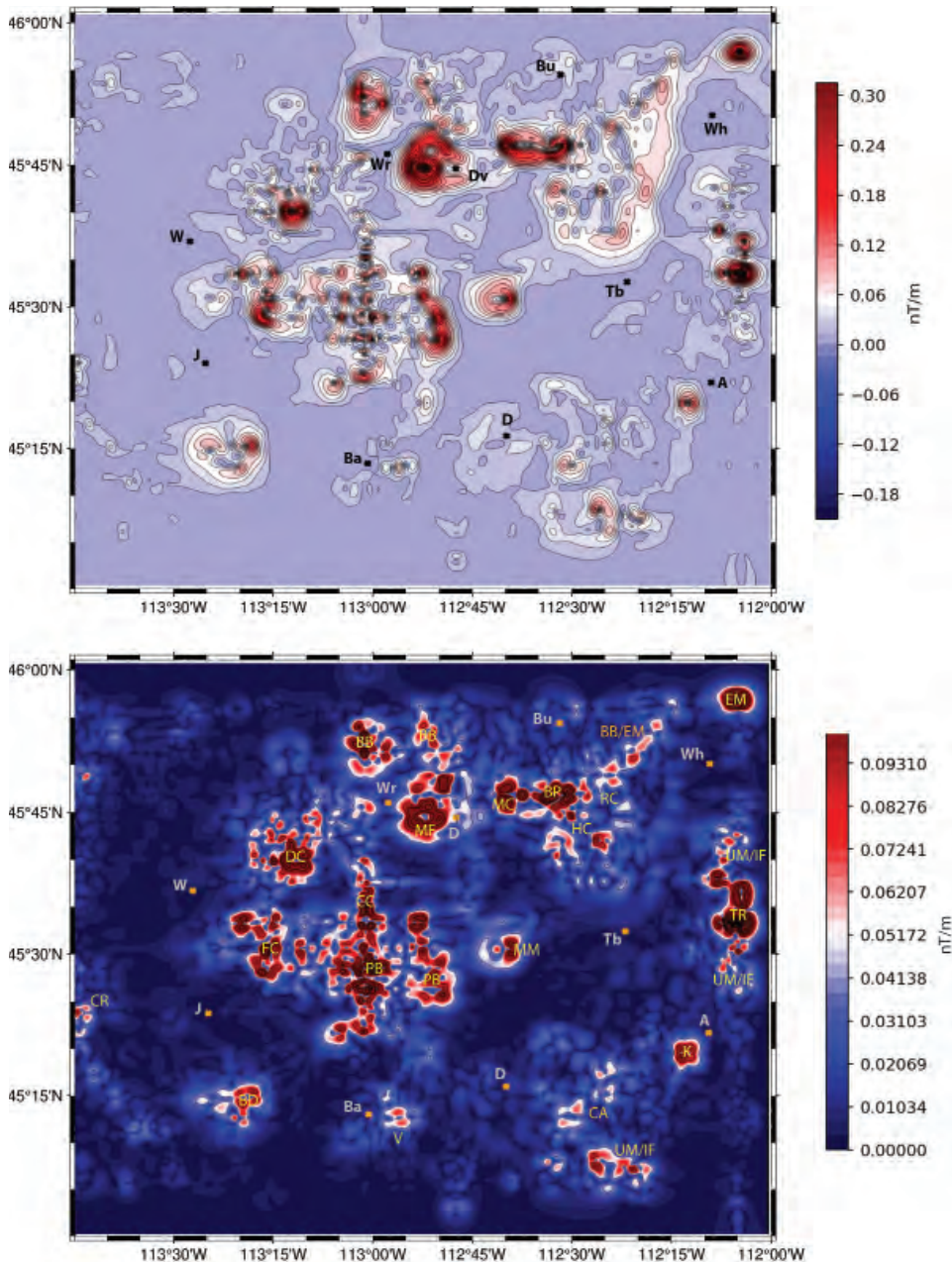


Figure 3. (Top) Total horizontal derivative (HDR) of aeromagnetic data. (Bottom) Curvature attribute of aeromagnetic data. Both curvature and HDR serve to highlight the shape and location of sources. In the study area these are principally igneous rocks and iron-rich metasedimentary rocks. Localities on map (white) are same as figure 2. Magnetic features (yellow) are: BB, Boulder batholith plutons; BD, Bloody Dick pluton; BP, Burton Park pluton; CA, Carter Creek iron deposit; CC, Clifford Creek pluton; CR, Carmen pluton; DC, Doolittle Creek pluton; EM, Elkhorn Mountain volcanics; FC, Francis Creek pluton; HC, Hell Canyon pluton; K, Kelly iron deposit; MC, Moose Creek pluton; MF, Mt. Fleecer pluton; PB, Pioneer batholith plutons; RC, Radar Creek pluton; TR, Tobacco Root batholith; UM/IF, ultramafic rocks/Iron formation; V, possible Bannack/Dillon volcanics eruptive vent.

non-responsive zones on the processed magnetic attribute data. A good example is observed in the area between the main Pioneer batholith and the McCartney Mountain stock (fig. 2). This area correlates with non-magnetic, Phanerozoic sedimentary rocks of the McCartney Mountain salient of the fold and thrust belt (fig. 1).

Faulting, folding, and other crustal structures are not as prominently imaged as in the regional gravity data, most likely due to a lesser contrast in magnetic than density

properties between fault-separated rocks. However, several structures have a distinct magnetic profile, especially where magnetic igneous rocks are juxtaposed against nonmagnetic sedimentary rocks. Examples from the Pioneer Mountains include the Comet and Fourth of July fault zones as well as the apparent northward continuation of the Ermont/Kelley thrust fault system—where igneous rocks of the Pioneer batholith were apparently emplaced along thrust fault planes. The northwest-trending McCartney Mountain, Blacktail, and Beaverhead Divide fault zones as well as the east-west-oriented Badger Pass fault zone are also apparent in the aeromagnetic data. On a larger scale, the magnetic data indicate the presence of a convex to the east indentation of the basement corresponding to the location of the McCartney Mountain salient. This feature corresponds to an arc of non-magnetic rocks that, starting from south of Bannack, runs northeast to its apex near the McCartney Mountain salient and then swings back toward the northwest along the McCartney Mountain fault zone. To the west, the strongly magnetic rocks of the Pioneer batholith occupy the core of this feature, while the eastern boundary is outlined by the moderately magnetic rocks of the Archean craton. It is proposed here that this arcuate feature marks the location of a Precambrian rift margin that, during Cretaceous orogeny, guided the emplacement of the fold and thrust belt. The presence of a preexisting step in the Archean basement would serve as a natural focus for later deformation, and in this scenario would represent an inverted rift margin. A deep-seated fault-penetrating Archean basement

could also have served as a primary conduit for the emplacement of igneous melt along the reactivated thrust fault planes (Kalakay and others, 2000, 2021; Sears, 2016). This hypothesis is consistent with geochemical data suggesting a Proterozoic crustal boundary under the Pioneer Mountains (Foster and others, 2012).



Gravity Data

In the Pioneer Mountains of southwestern Montana, the total HDR and curvature attributes, in combination with the tilt angle (TA), proved to be the most efficient tools for structural interpretation of the gravity field data. HDR represents the absolute value of the total gravity gradient in the horizontal plane (table 1) and its maxima should highlight the location of geologic contacts (Saibi and others, 2019). The TA refers to the dip angle of the gradient of the gravity field (table 1); it is ideally positive over the source of an anomaly and negative outside of it (Dentith and Mudge, 2018). The zero values of TA are generally interpreted as geological contacts and are therefore a useful edge finder for delineating geological structure (Bell, 1997). Tilt angle can also be used to obtain a rough estimate of the depth to a geologic contact. When contacts are approximately vertical, the depth to the top of the source is equal to half the distance between the $+45^\circ$ to -45° (± 0.7854 radian) contours on the map (Lee and others, 2010; Salem and others, 2007)

In the study area, a good correspondence between mapped geologic structures and observed gravity anomalies is apparent (figs. 1, 4, 5). Cenozoic normal faults juxtaposing low-density basin fill and higher density rock are the most prominent features. Major crustal structures, such as the northwest-trending McCartney Mountain and Blacktail fault zones, as well as the east–west-oriented Badger Pass fault zone, are also readily visible. East of Bannack, a prominent circular feature is particularly well-defined on the TA and HDR attribute maps (fig. 4, bottom, fig. 5). This feature corresponds to the surface outcrop of the Cretaceous Bannack and Cenozoic Dillon volcanic fields (Mosolf, 2018; Ruppel and others, 1993) and probably represents a long-lived ring fracture/volcanic dome complex. Of particular interest is the ability of the gradient-based interpretation methods to trace the location of the Cretaceous fold and thrust belt. On the HDR, TA, and curvature maps, the fold and thrust belt appears as a continuous convex eastward arcuate feature that follows the known trace of the Ermont/Kelley thrusts

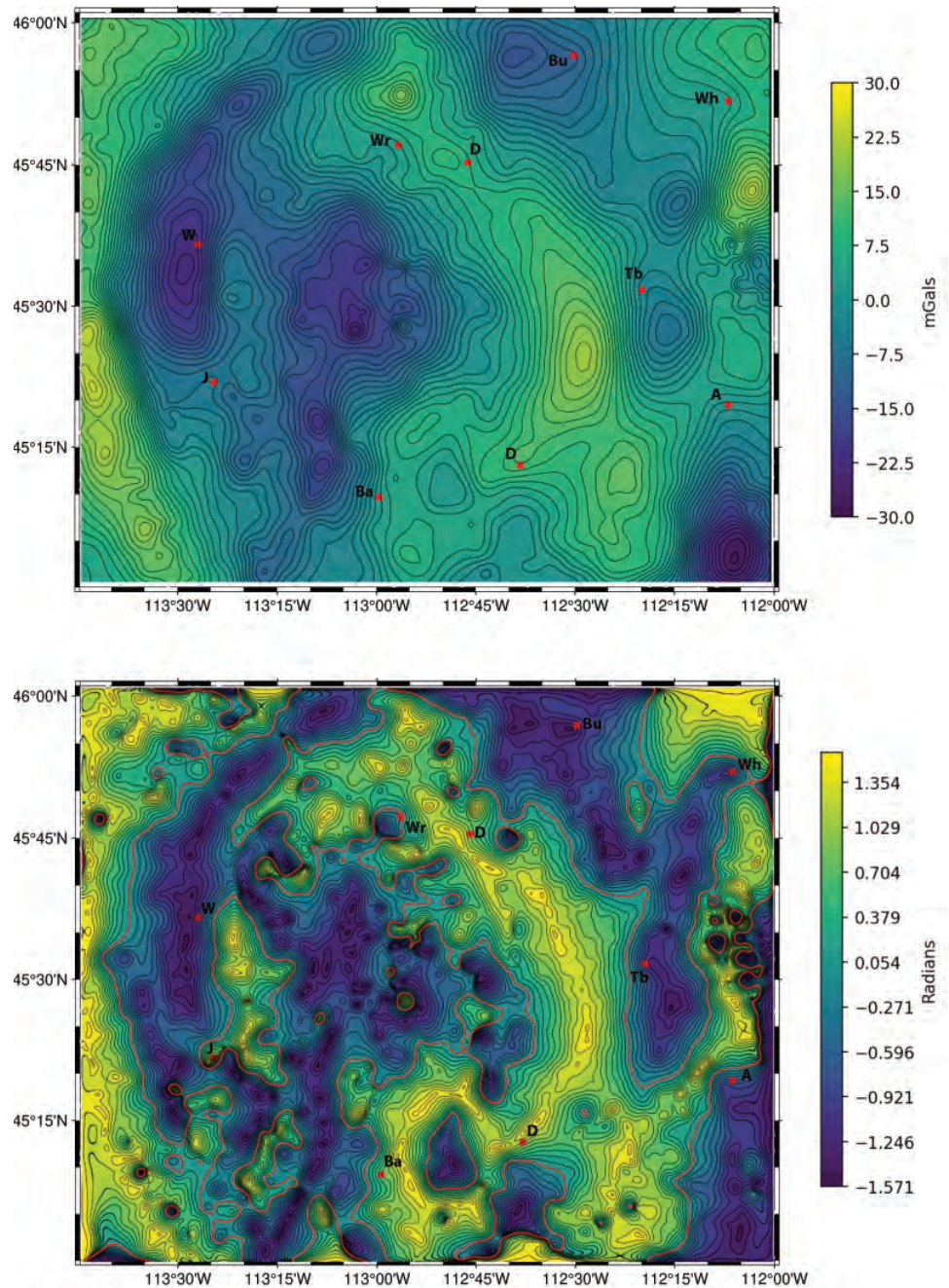


Figure 4. (Top) Map of gravity residuals. (Bottom) Tilt angle of gravity field. The red line on the map is the zero contour of the tilt angle, which is generally interpreted as a source edge finder. Localities on map: A, Alder; Ba, Bannack; Bu, Butte; D, Dillon; J, Jackson; Tb, Twin Bridges; W, Wisdom; Wh, Whitehall; Wr, Wise River.

through the Bannack area northward to where it disappears into the igneous terrain of the Pioneer batholith, eventually curving northwest into the McCartney Mountain fault zone. The deformation front in the McCartney Mountain area appears as a smaller, but otherwise similar, convex east arc that includes the area of the surface exposure of Cretaceous thrust faults within the salient (Kalakay, 2007). Intriguingly, this gravity feature corresponds in location and shape to the similar arcuate feature seen on the aeromagnetic data. This correspondence suggests that this arc represents a major crustal structure. It is proposed that this feature is a Precambrian rift margin reactivated and inverted as a thrust ramp during Cretaceous deformation. A possible analog

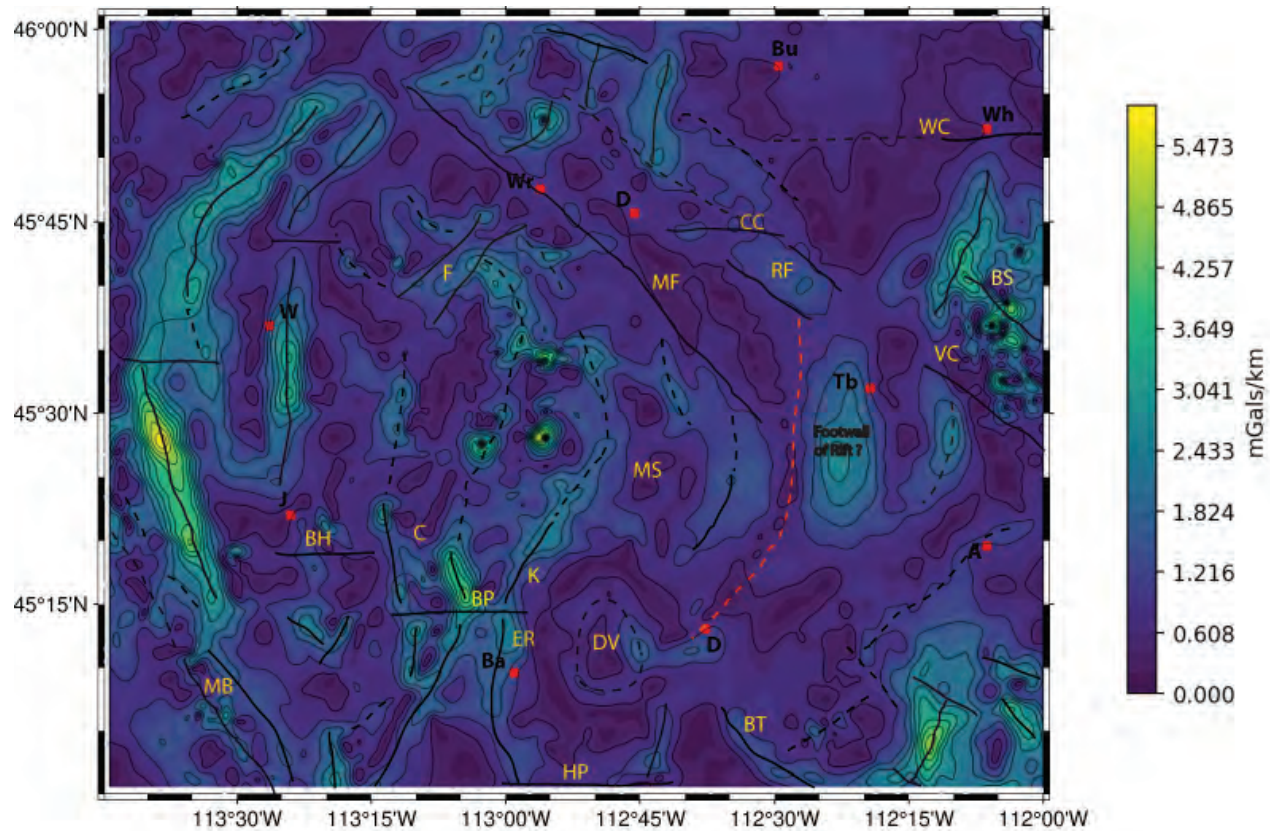


Figure 5. Total horizontal derivative of gravity map showing known (solid) and interpreted (dashed) tectonic features. Compare to surface geology in figure 1. Localities on map (black) are same as figure 4. Interpreted gravity features (yellow) are: BH, Big Hole fault zone; BP, Badger Pass fault zone; BS, Bismark/Spanish Peaks fault zone; BT, Blacktail fault zone; C, Comet fault zone; CC, Camp Creek fault zone; DV, Dillon/Bannack Volcanics ring structure; ER, Ermont thrust; F, Fourth of July fault; HP, Horse Prairie fault zone; K, Kelley thrust; MB, Miner Lake/Bitterroot fault zone; MF, McCartney Mtn fault zone; MS, McCartney Mtn salient; RF, Rochester faults; VC, Virginia City fault zone; WC, Willow Creek fault zone. Red dashed line is interpreted edge of unrifted craton.

is seen in the Galicia continental margin, where compressional stress reactivated the former rift detachment at the base of the crust; then, as compressional stresses increased, deformation propagated along the detachment, leading eventually to the inversion of preexisting normal faults within the continental basement and the deformation of the post-rift sedimentary cover (Druet and others, 2018).

Forward Modeling of the Vertical Component of Gravity Data

The interpretation of an inverted rift margin can be tested by forward modeling of the vertical component ($\partial f/\partial z$) of the gravity field. An east–west profile across the central Pioneer Mountains through the McCartney Mountain salient east into largely undeformed basement was tested (fig. 6). The observed surface geology (Pearson and Zen, 1985; Ruppel and others, 1993; Tysdal and others, 1994; Vuke and others, 2007), along with laboratory measurements of rock density (Hanna and others, 1993), serve as important constraints on the proposed model. An east–west seismic line, ground-truthed by several deep wells from the Pioneer Batholith through the McCartney Mountain salient, parallels part of this profile (Lopez and Schmidt, 1985). The central feature of this model is a stepped basement contact that deepens towards the west. These base-

ment steps represent Precambrian normal faults that bound grabens filled with rift-related volcanics and sediments. These basement steps were reactivated during Cretaceous time as thrust ramps, and Cenozoic extension later reused these same zones of crustal weakness. Cretaceous magma employed these same crustal discontinuities as emplacement conduits with intrusions preferentially located near the top of thrust ramps (Kalakay and others, 2001; Lageson and others, 2019; Sears, 2016). The thin-skinned fold and thrust belt including the McCartney Mountain salient sole into a regional decollement at or near the top of the basement (Lopez and Schmidt, 1985). East of the McCartney Mountain salient, within the footwall of the rift, thrusting has stepped down into the basement, potentially forming a “double-decker” structure (Parker and Pearson, 2021; Parker and others, 2022).

As can be seen from figure 7, a good match is obtained between the observed $\partial f/\partial z$ and the $\partial f/\partial z$ response calculated from the proposed model. This suggests that the inverted rift model is a viable, geologically realistic explanation of the regional geology. It accords well with several previously proposed ideas for the tectonic evolution of this region. It also suggests that the northwest-trending McCartney Mountain fault zone is a major crustal structure that greatly influenced Cretaceous compressional deformation



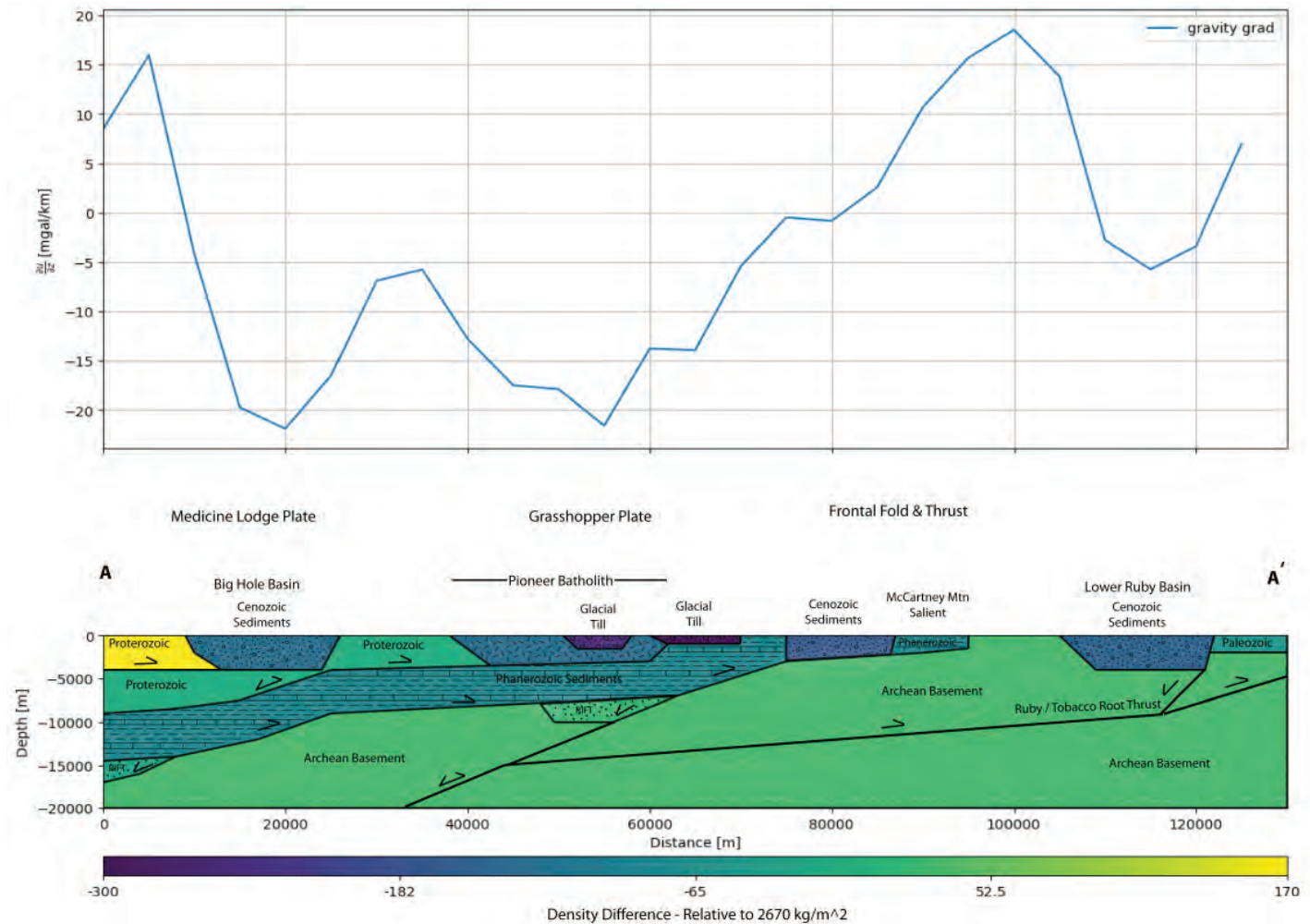


Figure 6. Geologic model for vertical component of gravity field for profile across the McCartney Mountain salient. Line of profile corresponds to line A–A' on figure 1. Vertical component of gravity calculated using the pyGIMLi program for geophysical modeling (Rucker and others, 2017) is shown above. Note the stepped geometry of the basement contact, the presence of Precambrian rift sediments, reactivation of rift margins first as thrust ramps and then as Cenozoic normal faults, the relationship of igneous intrusions to tectonic structure, and the downstepping of Cretaceous thrust faulting eastward into the cratonic basement.

and igneous intrusion as well as Precambrian rifting. Along with other northwest-trending structures in southwestern Montana, this fault zone likely had a beginning in the Archean orogenic events that sutured together the northwestern margin of the North American craton (Mogk and others, 2020).

CONCLUSIONS

The results of this study suggest the following conclusions:

To a first approximation, Cretaceous deformation in the Pioneer Mountains area can be understood as a tectonically inverted Precambrian rift that also guided the emplacement of granitic magmas.

The Ermont/Kelley thrust system can be observed to pass north into the area now occupied by the intrusions of the Pioneer batholith. In the Bannack area, the Ermont thrust is associated with the contemporary Bannack volcanic field, which suggests this area is a structurally higher

segment of the same paired deformation/igneous system. These observations support the hypothesis that magma was intruded along the contemporary fault zone. The surface expression of this thrust system appears to have been largely removed by erosion northwards except for a few scattered klippe. Originally, the fault zone likely merged into the northwest-trending McCartney Mountain high-angle oblique fault, as indicated by both gravity and magnetic data.

The McCartney Mountain fault zone is a major crustal discontinuity with a Precambrian ancestry. This zone acted as a northern buttress for the Cretaceous fold and thrust belt. East-verging folds and thrusts are observed to be deflected northwestward into this zone, and it also appears to have served as a lateral ramp for Cretaceous deformation. The McCartney Mountain thrust zone appears to be the southern boundary of a broader zone of complex deformation bounded on the north by the westward continuations of the Spanish Peaks/Bismarck and Willow fault zones. This broad zone of deformation links the fold and thrust belt of



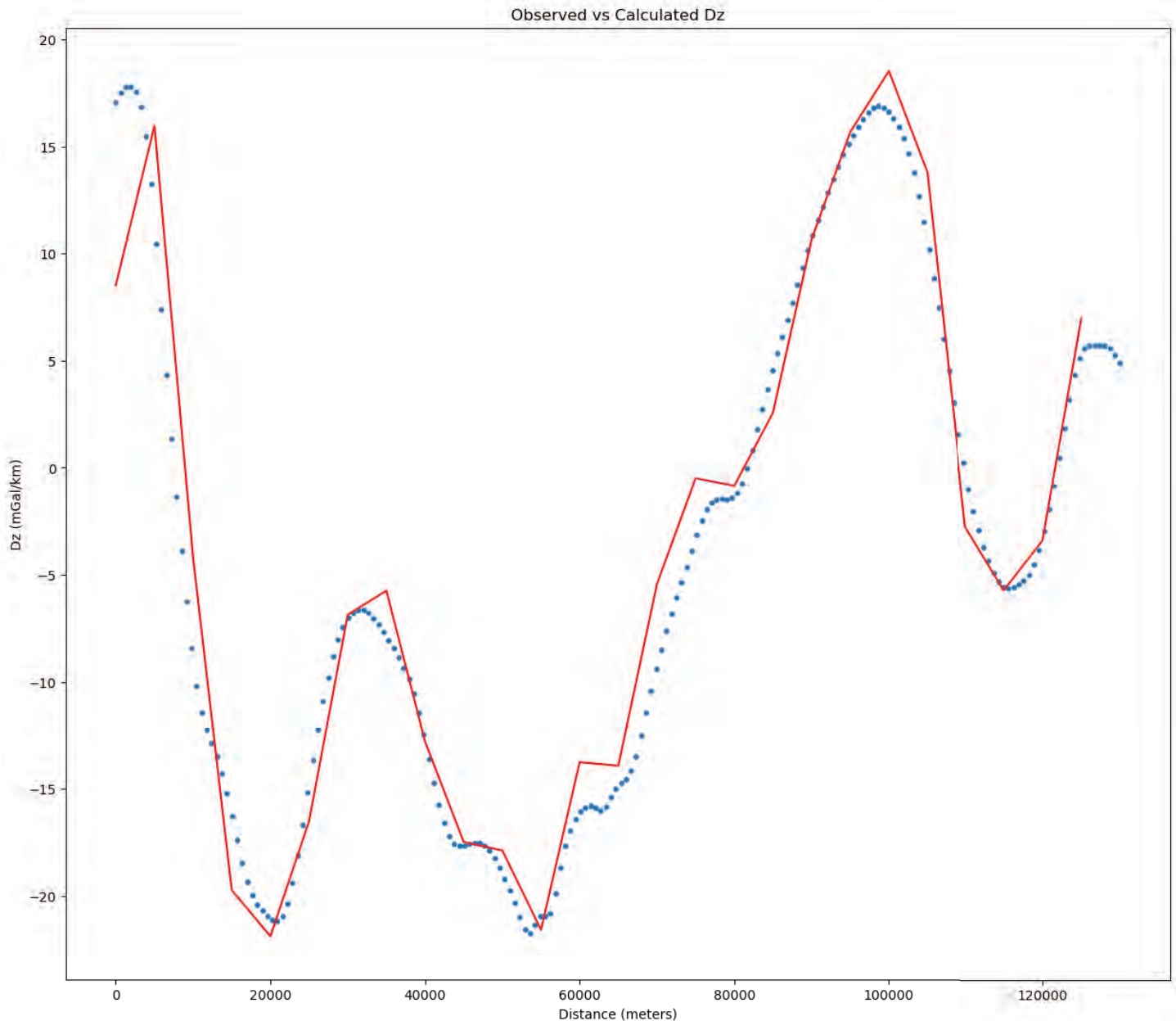


Figure 7. Comparison of measured (blue circles) and calculated (red line) values for the vertical component of the gravity data for the McCartney Mountain profile in figure 6.

southwestern Montana with the Cretaceous deformation of the Helena salient.

The eastern boundary of the Grasshopper plate was controlled by the western edge of undeformed Precambrian basement, which is interpreted here as a convex eastward arc that originated as a late Proterozoic/early Cambrian rift margin. The McCartney Mountain fault zone bounds this surface to the north, but its southward extension is uncertain. The McCartney Mountain salient of the fold and thrust belt occupies the center of this convex surface and was thrust over the edge of the rift margin onto the Precambrian basement.

Gravity and magnetic data support the hypothesis that igneous intrusions, including the Pioneer and McCartney Mountain batholiths, are localized above Cretaceous struc-

tural culminations that created accommodation space for the emplacement of igneous material (Kalakay and others, 2000, 2001).

Cenozoic normal faults predominately reactivated Cretaceous thrust ramps that were themselves controlled by the locations of Precambrian rift margins. The unusual curved shape; steep, almost vertical boundary faults; unusually thick basin-fill; and segmentation by east-west-oriented high-angle faults suggest that the Big Hole Basin is the modern expression of an original Proterozoic rift graben.

East-west-oriented high-angle faults, such as the Horse Prairie and Badger Pass fault zones, are interpreted here as having originated as faults that segmented the Proterozoic rift margin. The role they played during later compressional deformation is uncertain. Fault zones of Archean ancestry



may also have distributed tensional stresses during Proterozoic rifting.

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FIELD GUIDES

THE TILTED TERTIARY SECTION AT MILL POINT NEAR BANNACK, MONTANA

Jeff Lonn and Colleen Elliott

Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

INTRODUCTION

This trip requires about 4 mi of off-trail hiking with 1,000 ft of elevation gain. No water is available and much of the hike lacks shade. The trip will examine a tilted section of Renova-age, Tertiary sedimentary rocks, walking down-section through a coarsening succession to its lower contact with the underlying bedrock. The cobble–boulder conglomerate of this succession is displayed on Mill Point on the drive in to the start, and was mapped as Tertiary–Cretaceous Beaverhead conglomerate by Ruppel and others (1993). However, the tilted section we will view shows that the conglomerate is vertically gradational with overlying finer-grained Tertiary deposits, and therefore these strata have been reinterpreted as late Oligocene–Eocene (Janecke and others, 2005) Renova-age Medicine Lodge beds (Lonn and others, 2019). We will walk down-section from fine-grained tuffaceous strata into cobble conglomerate, then into angular breccia, and finally into the underlying quartzite bedrock interpreted to be Cambrian. We will then proceed to Red Butte, an isolated hill of Mesoproterozoic Belt quartzite surrounded by bright red Tertiary sediments (fig. 1).

TO THE TRAILHEAD

Drive north out of the Bannack State Park to Highway 278. Turn left on 278 and proceed 3.7 mi to Taylor Road, a gravel road on the right just beyond a house beneath Mill Point. Turn right, cross Grasshopper Creek, and pass under the steep slopes of Mill Point. The hill above the road (Mill Point) is a dip slope developed on rounded to angular, cobble–boulder conglomerate assigned to the Medicine Lodge beds (Tmlc). We will be examining these rocks and their stratigraphic position as part of this field trip. At 0.5 mi, turn left onto the Red Butte–Dyce dirt road just beyond some dumpsters. Proceed 0.3 mi to the top of a hill and some bulldozed prospects (45° 14.3049° , -113° 4.1613°). Park here to start the hike.

THE HIKE

Figure 1 shows the hiking route overlain on a geologic map. First, examine the prospects. They expose rounded to subangular, multi-lithological, pebble–cobble conglomerate interbedded with white, ashy fine-grained sandstone, some containing matrix-supported pebbles. The beds strike northerly and dip 15° – 20° east. What

were these bulldozed trenches seeking? The horizontal sediment layer that caps these strata is unconsolidated, angular to rounded, cobble gravel about 6 in thick. Without seeing a cross-section through these gravels, one might be tempted to call them an overlying depositional unit like Sixmile Creek Formation or Quaternary alluvium, but the view of the trench walls suggests they are erosional lag gravels rather than a mappable deposit (fig. 2). Mappers beware: erosional surfaces such as pediments abound in western Montana and are easily mistaken for younger deposits.

From the prospects, walk WNW to the white and pink outcrops visible on the hillside (**Stop 2**). Along the way note the extensive caliche development, especially on the sides of the shallow gullies. Resistant beds or lenses of poorly sorted, coarse-grained gritty sandstone protrude from the surface and strike generally north and dip 35° – 40° east. Grains are mostly angular quartz but include abundant and variable lithics. Continue down section into the white and pink outcrop area, where red, white, and pale yellow clay, silt, and sand beds compose most of the strata. Calcareous sandstone beds are also present. Rounded cobbles littering the surface are being shed from the slopes above.

Continue up the slope to a slight bench, then do a climbing traverse northwest to the lowest visible outcrops. These outcrops are clast-supported cobble–boulder conglomerate (fig. 3) with some sandstone interbeds. Clasts are dominantly well-rounded quartzite cobbles of variable lithology, and the matrix is coarse-grained sand.

Follow the outcrops upward. This is essentially a dip slope, so we are not crossing much section. Lithologies continue to be dominated by cobble–boulder–pebble conglomerate. Most clasts are composed of feldspathic quartzite, but are multicolored, and rare volcanic clasts can also be found. Most clasts are rounded, but a few are subangular or angular.

Walk to the top of the ridge (**Stop 3**), where there is a thick interval of massive, coarse- to very coarse-grained sandstone. Grains are angular and include abundant feldspar and lithic fragments. Is this bedrock underlying the Tertiary conglomerate? Probably not: this gritty sandstone does not share much similarity



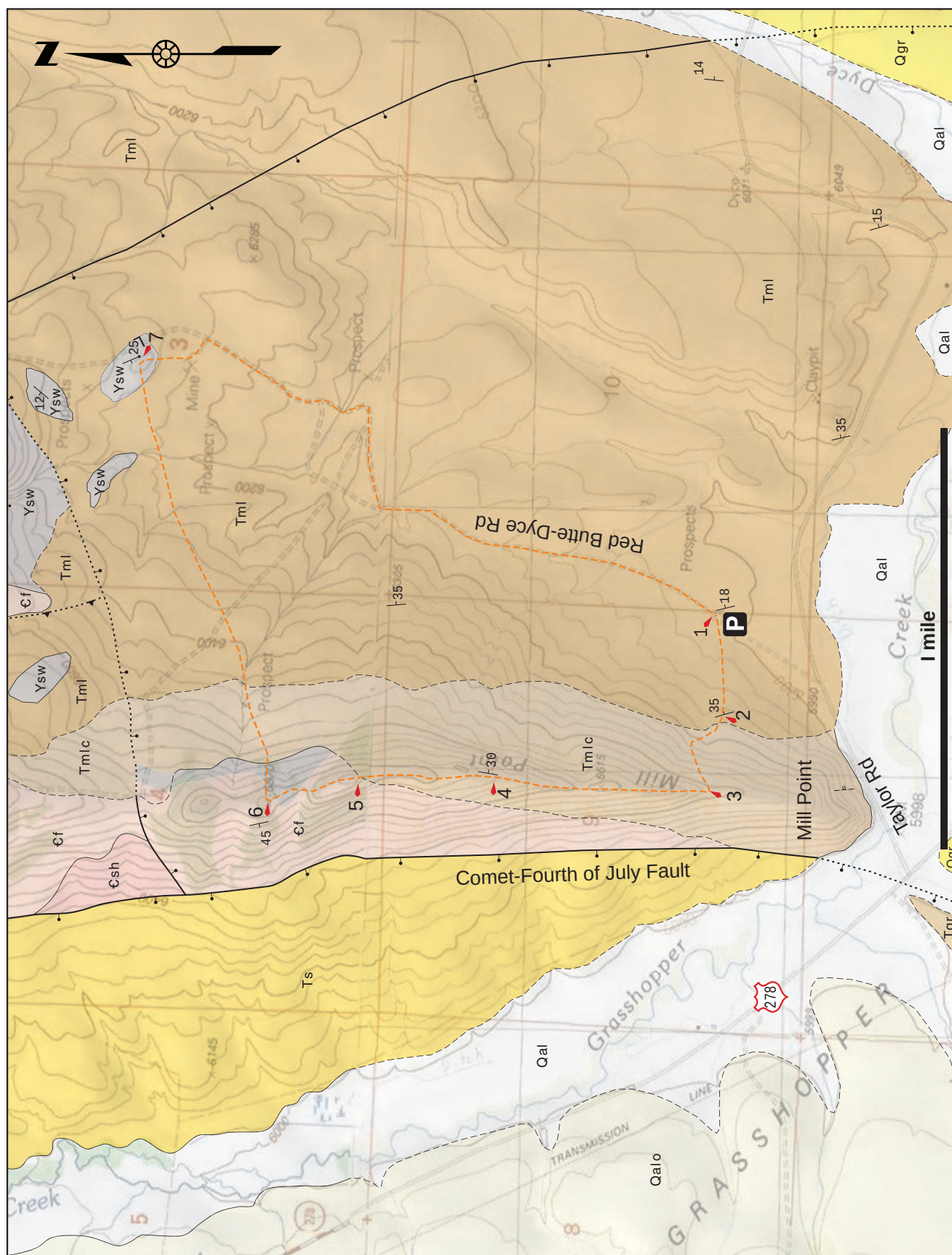


Figure 1. Geologic map of the field trip area. The hiking route is shown by the orange dashed line. Numbered points are stops listed in the text. Units from oldest to youngest are: Ysw, Swauger Fm; Cf, Cambrian quartzite; Csh, Cambrian quartzite; Ts, sedimentary rocks, undivided; Tgr, gravel; Tmlc, Medicine Lodge conglomerate; Tml, Medicine Lodge beds; Qgr, gravel; Qal, older alluvium; and Qal, alluvium. Modified from Lonn and Lewis (2012) and Lonn and others (2019).



Figure 2. View of the trench wall at parking area (Stop 1) showing light-colored Oligocene–Eocene Medicine Lodge beds, which dip away from the camera, capped by a 6-in-thick layer of lag gravels on a nearly horizontal surface.



Figure 3. Rounded cobble–boulder conglomerate common in the lower part of Oligocene–Eocene Medicine Lodge beds (Tmlc on map). Note clasts of different lithologies. Wristwatch for scale.





Figure 4. Lowest part of the Medicine Lodge conglomerate (Stop 4) is dominated by angular clasts, but note the rounded boulder beneath the pencil confirming the sedimentary, rather than tectonic, origin of this "breccia."

with the Belt or Cambrian quartzites that dominate the bedrock in this area. Those are finer-grained, with less angular grains and fewer lithics.

Walk north along the ridge crest. The north–south Mill Point ridge is a hogback developed on the resistant, moderately east-dipping Tertiary conglomerates and sandstones, but its west side is bounded by the down-to-the-west Comet–Fourth of July normal fault that uplifted the steep western face of the East Pioneer Mountains between here and Wise River (Elliott and others, 2025). Walk north; the walking is slow over the lichen-covered boulders and through the scrubby juniper–Douglas fir forest. Soon more cobble–boulder conglomerate appears. Clasts here are mostly angular, although some are rounded. Continue north on the ridge crest until you reach a series of good outcrops for discussion (**Stop 4**, $45^{\circ} 14.7545'$, $-113^{\circ} 4.7015'$).

These outcrops provide some clean exposures of the conglomerate. Clasts are of all sizes from pebbles to boulders, and most are angular (fig. 4). One might be tempted to call this a tectonic breccia, but a closer look reveals rounded clasts as well, and also a variety of quartzite lithologies. Tertiary(?) sedimentary breccia like this is extensively developed in western Montana and can easily be mistaken for fault breccia if cemented, or bedrock if unconsolidated.

Continue walking the ridge north to the base of a bedrock knob (**Stop 5**, $45^{\circ} 14.9852'$, $-113^{\circ} 4.7069'$). The rock exposed here and on the knob is the underlying quartzite bedrock. It is white, massive, quartz-rich, well-sorted, and medium- to coarse-grained. It is interpreted as Cambrian in age due to its quartz content (93–97%), its stratigraphic position above feldspathic Belt quartzite, and the presence in some areas of black argillite interbeds with trace fossils. Cambrian quartzite in this region may be as much as 600 m (2,000 ft) thick (McDonald and Lonn, in review; McDonald and Yakovlev, 2019).

Walk up and over the knob, and carefully descend its steep north side to a smooth saddle. Continue to the next knobby outcrops atop the ridge (**Stop 6**). These are composed of the same massive light-colored quartzite. Bedding is difficult to find. One good surface in this area displays planar cross-beds on a scale of 0.3 m (1 ft). It appears that the quartzite strikes roughly north and dips moderately west, in contrast to the east-dipping Tertiary strata.

Descend eastward, following a string of outcrops down a subtle ridge. These are the same white quartzite, but are highly fractured. The outcrops end, and the ground is again covered with unconsolidated boulders and cobbles of mostly quartzite that is probably derived

from underlying Tertiary conglomerate. Some bulldozer trenches that expose clayey material beneath the clasts suggests that some is colluvial material derived from the slopes above.

Continue towards the obvious bright red rocks surrounding a small hill ($45^{\circ} 15.5036'$, $-113^{\circ} 3.5282'$); this is the real Red Butte as opposed to the higher one farther north labeled as Red Butte on topographic maps. Cross a dirt road and, when you encounter a fence, look for a large boulder along the fence that allows an easy crossing. As Red Butte is approached, note that the bright red material exposed by roads and trenches on its south side appears to extend to the top of the hill. Bedded quartzite can be seen on the left (north) side. Head for the contact between the red material and the bedded quartzite and follow it up the hill. The quartzite differs from the Cambrian quartzite we observed earlier. It is reddish and contains 10–20% chalky white feldspar grains and abundant mud chips. It is interpreted as Mesoproterozoic Swauger Formation of the Lemhi subbasin, upper Belt Supergroup (Missoula Group equivalent). Looking northwest, west-dipping beds of this quartzite are visible (fig. 5). A resistant ledge at the top of this section is the lowest part of the Cambrian quartzite we viewed earlier (fig. 1). As you walk up,

note that the Swauger quartzite is covered by reddish stain, derived from overlying Tertiary sediments that must have covered it.

At the top of the hill (**Stop 7**), multi-lithology conglomerate similar to that seen earlier appears to be in depositional contact with the underlying Swauger Formation. Tertiary sedimentary rocks, many of them red, appear to surround this island of Swauger Formation. Janecke and others (2005) interpreted bedrock islands like this as huge landslide blocks originating in adjacent highlands, but the bedding attitude here is consistent with bedding just to the north, and the Swauger is not brecciated. Therefore, we interpret such islands of bedrock in a sea of Tertiary sediments as paleotopographic highs.

Walk south down the hill to the trenches and prospects in the bright red material, which is clay-rich and contains intervals with floating angular pebbles. Bedding is difficult to discern except in the largest pit, where subtle color changes appear to define beds that strike north and dip gently to moderately east. The red beds extend both north and south from Red Butte, supporting this orientation interpretation. Clearly, the trenches were seeking the red material, but for what reason?



Figure 5. View northwest from Stop 7 of moderately west-dipping Swauger Formation. The prominent ledge on the left side of the photo marks the lowest beds of the overlying Cambrian quartzite that we viewed earlier.



Local lore suggests that it was mined for pottery.

Walk a short distance south to a cattle guard and take the dirt road south back to the vehicles. At first, the ground is littered with cobbles and boulders that may represent either the underlying lithology, lag derived from now-eroded overlying beds, or Quaternary gravel washed in from the highlands to the north. Mapping unconsolidated valley deposits in western Montana is challenging! Soon the substrate becomes light brown clay and silt, representing the fine-grained Medicine Lodge beds, that continues along strike back to the vehicles.

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BANNACK STATE PARK TO BADGER PASS, MONTANA: PLACER GOLD AND CENOZOIC LANDSCAPE EVOLUTION

Jim Sears

Emeritus Professor, University of Montana

Bannack had the first big gold strike in what became Montana. But where did the gold come from? How did it get there? Can a geologic field trip from Bannack to Badger Pass give answers, or raise more questions?

The Bannack placer gold deposits were in Pleistocene gravel in the bed of Grasshopper Creek. Rich bedrock lode deposits were found nearby, but they were *downstream* of the rich placers. Some paleo-placer mines were dug in gravel above the modern creek bed, within 1 km downstream of Bannack. One paleo-placer was perched 100 m above the valley bottom. Extensive perched paleo-placers also occur *upstream* of Bannack on the early Pleistocene Bannack Bench 20 m above Grasshopper Creek.

But what was the source of the gold? There are no significant gold districts upstream of Bannack along Grasshopper Creek. In this paper, I develop a landscape model to explain how to get Bannack's placer gold *upstream* of its sources.

In today's landscape, Grasshopper Creek makes a sharp left turn from the wide Grasshopper valley to cross a major fault zone and enter a deep canyon (fig. 1). The canyon cuts east through a high ridge of Paleozoic rocks. A gold-bearing Cretaceous pluton is exposed in the canyon within 1 km downstream of Bannack. The fault zone is the Muddy–Grasshopper detachment, and the ridge is its uplifted footwall block. The detachment is thought to be of Eocene–Oligocene age (Lonn and others, 2019). It is part of a system of faults traced for 90 km from the Tendoy Range north to the Pioneer Mountains. It bounded a large graben system to the west that was active during eruption of the thick Eocene Challis volcanics and deposition of the Eocene–Oligocene Medicine Lodge beds (equivalent to the Renova Formation; Lonn and others, 2019).

In Paleocene to early Eocene time, many paleo-valleys of the Rockies originally drained eastward, down a regional slope from the Continental Divide. Grasshopper canyon may have originated as one of those old east-draining valleys. The canyon has incised meanders that suggest it originally followed an early, low-gradient valley before uplift of the detachment footwall.

The north-trending Muddy–Grasshopper detachment system crossed the older east-trending valleys, beheading many drainages and leaving them as wind gaps in uplifted footwall blocks. Some of the beheaded drainages have large

debris flows at their eastern ends, as if tectonic tilting of the footwall blocks had evacuated their sedimentary fills, which then cascaded as slurries into neighboring grabens. The debris flows date to the 18 Ma Middle Miocene, when a major tectonic event associated with the initiation of the Basin-Range province reactivated Eocene fault zones in southwest Montana (Sears and others, 2007).

Badger Pass is one such wind gap, one of the beheaded east-trending paleo-valleys of the footwall ridge, seen at STOP 4 (figs. 1, 4). A thin deposit of river gravel is preserved at the summit of Badger Pass (6,760 ft elevation). It has well-rounded Belt Supergroup pebbles, their nearest sources being 15 km to the west. Montana Highway 278 follows the Badger Pass paleo-valley across the footwall ridge to the head of a Miocene gravel fan that slopes eastward to the Beaverhead River.

It appears that a pre-detachment stream was beheaded at Badger Pass, when the detachment footwall was uplifted. But by contrast, Grasshopper Creek was able to cut a canyon completely across the uplift at Bannack. Why?

At Bannack, bouldery gravel directly overlies the detachment surface (fig. 2). The boulders were eroded from the footwall of the detachment. Lowell (1965) mapped the gravel as the sedimentary onlap of alluvial fans on the detachment. The gravel flanks the footwall block from Bannack north to Badger Pass, but at Bannack, it is cross-cut by the modern bed of Grasshopper Creek.

It appears that the Grasshopper canyon cut through the footwall block after the gravel fans had been deposited. Headward erosion of the canyon removed the gravel fan and cut across the footwall block, gradually lowering the drainage divide between the Grasshopper basin and a tributary of the Beaverhead River. The tributary was the descendant of an original east-flowing stream that had been beheaded by uplift of the detachment footwall. Headward erosion of the Bannack end of the footwall block eventually lowered the divide enough to allow throughgoing drainage for Grasshopper Creek. Until that came to pass, gold could be washed *downstream to the west* to the Bannack placers from the bedrock gold veins of the Cretaceous intrusions.

The Grasshopper–Beaverhead connection may have occurred in the Pleistocene when run-off from the high Pioneer Mountains overtopped the low drainage divide in Grasshopper canyon. The low divide may have been

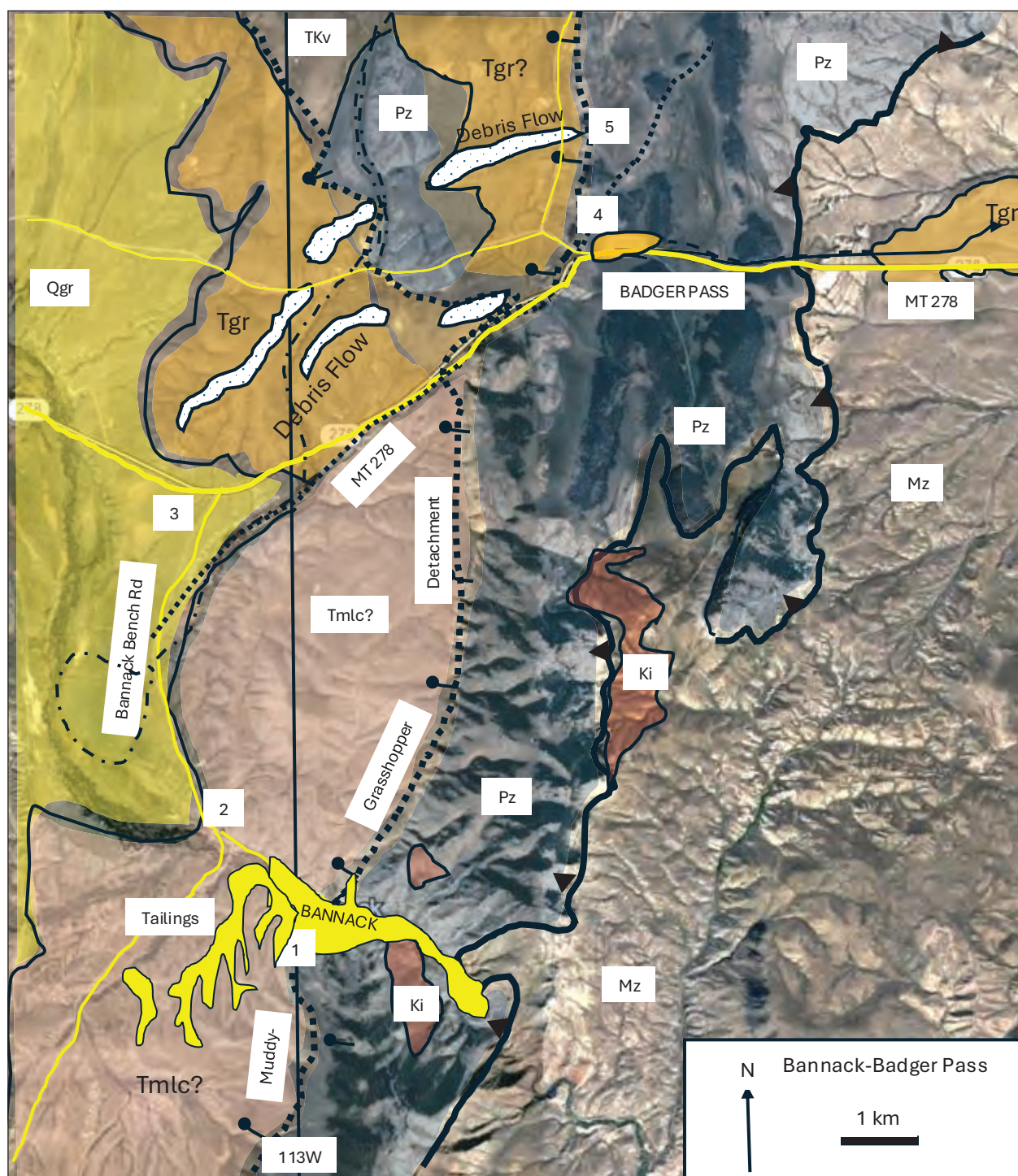


Figure 1. Geologic sketch map of Bannack–Badger Pass area of field trip. Numbered white squares locate field stops. Pz, Paleozoic; Mz, Mesozoic; Ki, Cretaceous intrusives; TKv, Tertiary–Cretaceous volcanics; Tmlc, Eocene–Oligocene Medicine Lodge conglomerate after Lonn and others (2017); Tgr, Miocene gravel after Lonn and others (2017); Qgr, Quaternary gravel after Lonn and others (2017). Stippled pattern shows debris flows.

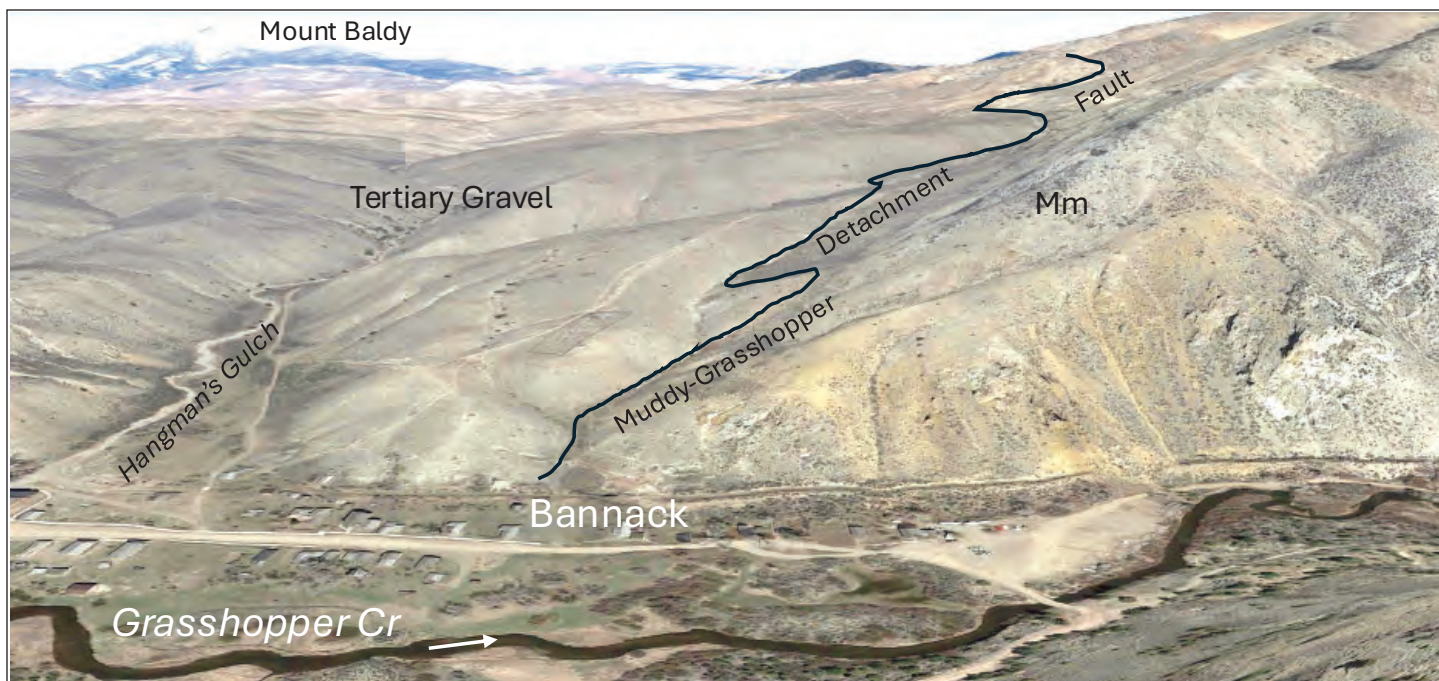


Figure 2. View north from Hendrick's Mill of Muddy-Grasshopper detachment fault. Note that gold-bearing Quaternary gravel of Grasshopper Creek valley cross-cuts the Tertiary gravel. Was Tertiary gravel cut by the detachment or was it deposited on the detachment? Is it Eocene or Miocene?

located 2 km downstream from Bannack in a zone of complex faulting where the resistant limestone rimrock of the canyon descends to the creek bed.

After the stream capture, some gold was reworked in the Bannack placers, and some washed downstream along the newly minted Grasshopper Creek. Some gold remained in perched placers in old alluvial fans on the sides of Grasshopper canyon and on the Bannack Bench at the mouth of the canyon.

ROAD LOG

0.00 mi Begin at Road Agent Campground, Bannack State Park. Turn left on Bannack Road.

STOP 1—0.51 mi Hendrick's Mill. View north of Muddy-Grasshopper Detachment fault (fig. 2). Hangman's Gulch is on the hanging wall of the detachment, in gravel interpreted as Eocene-Oligocene Medicine Lodge conglomerate faulted against Mississippian Madison Group in the footwall. It may, however, be that the gravel is in depositional contact with the detachment fault and represents alluvial fans from the footwall block. The gravel is crossed by the Pleistocene gold-bearing gravel of Bannack. Note that placer mines were upstream to the west of the bedrock canyon.

The detachment fault dips gently west (to the left), and cuts across beds in the footwall that were folded in the Sevier/Laramide orogeny. The Mississippian Madison Group limestone holds up the rimrock of the canyon. Mount Baldy in the background is a glacial horn of Belt Supergroup in the high Pioneer Mountains.

Return to Road Agent Campground and proceed to intersection of main road to Bannack. Turn left on main road.

STOP 2—0.1 mi Park along the side of the road to examine the gravel deposit. This gravel deposit is mapped as the Eocene-Oligocene Medicine Lodge conglomerate (Lonn and others, 2019). Take a look at the gravel's composition and texture for comparison to gravel mapped as Miocene to be seen at STOPS 3 and 5. Is the gravel cut by the detachment or does it onlap the detachment?

Continue west on Bannack Road.

0.4 mi Turn right on Bannack Bench Road, along a Pleistocene terrace at the foot of the gravel deposit.

0.43 mi Bannack Cemetery on right.

STOP 3—2.5 mi Intersection of Bannack Bench Road and Montana Highway 278 (fig. 3). A northeast-trending fault is mapped between the Eocene-Oligocene Medicine Lodge conglomerate to the southeast and Miocene gravel to the northwest (Lonn and others, 2019). The Miocene gravel is described as mature gravel of well-rounded quartzite cobbles interlayered with varying amounts of fine-grained and ashy clastic sediments. Its base is dated regionally at 21–18 Ma (Lonn and others, 2019). Compare the mapped Miocene gravel to the gravel we saw at STOP 2. Notice that the Miocene gravel at STOP 2 [?] contains large blocks of Quadrant Quartzite derived from the footwall of the detachment.



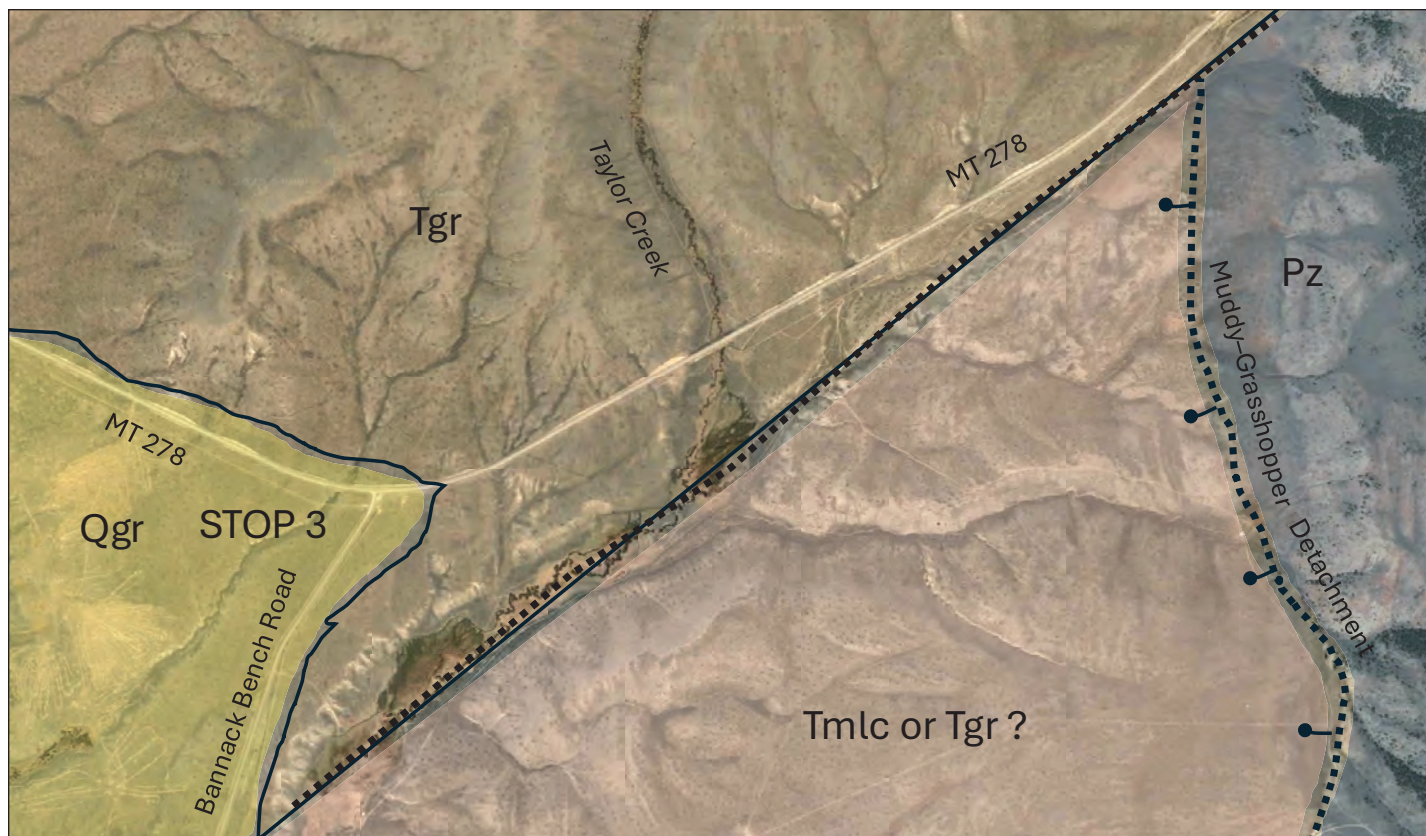


Figure 3. Area of STOP 3. View north. Intersection of Bannack Bench Road and Montana Highway 278. Same symbols as figure 1. Is Tgr actually the same unit as Tmlc? Note that Taylor Creek cut a Pleistocene channel through Tgr. From Google Earth image.

0.75 mi Cross Taylor Creek, which eroded a Pleistocene valley through the Miocene gravel. We will examine the valley at STOP 5.

STOP 4—2.66 mi Turn left and park at Taylor Creek Road pullout (fig. 4). Badger Pass wind gap. Here we see a small remnant of stream gravel. Examine the gravel. Notice the rounding of the pebbles and their compositions. Do you recognize any Belt Supergroup pebbles? The paleo-river that transported these pebbles was beheaded by the detachment fault as the footwall was raised into this ridge and Grasshopper valley collapsed into a broad graben.

This paleo-valley was left as a dry streambed, while at Bannack, the Grasshopper valley cut the canyon across the footwall.

This beheaded paleo-valley winds down across the Quadrant Quartzite ridges and meets a Miocene alluvial fan that slopes east, down to the Beaverhead River. The fan is part of a system that unconformably overlies the Eocene–Oligocene Renova Formation (Medicine Lodge equivalent), and is thought to correlate with nearby fans interlayered with Miocene tephra beds.

Proceed west on Taylor Creek Road.

0.37 mi Turn right on Badger Pass Road, proceed north.

STOP 5—0.74 Top of a low rise (fig. 5). Miocene-Pliocene debris flow. Large blocks of white Quadrant Quartzite are scattered on the rise southwest of the road.

The low rise is the topographically inverted valley of a debris flow that trends southwest for 5 kms from here down to Stop 3. Notice the large angular blocks of white Quadrant Quartzite scattered through the deposit. Traverse for 1.5 km along the debris flow and notice the decrease in size of the Quadrant blocks. The next ridge to the southeast is a similar debris flow, also armored by Quadrant blocks. Early people fashioned tools and weapons from well-cemented quartzite blocks on that ridge.

About 1.5 km southwest of the road the debris flow is cut across by the 100 m deep Pleistocene canyon of Taylor Creek. The canyon meanders south from the flank of the high Pioneer Mountains. The drainage direction shifted by 90 degrees from the west-flowing debris flow to the south-flowing Taylor Creek. Here the canyon cuts across Paleozoic rocks. It is floored by recent alluvium, but downstream at the confluence with Grasshopper Creek it cuts across an older alluvial terrace to an alluvial fan.

END OF ROAD LOG
Return to Montana Highway 278.

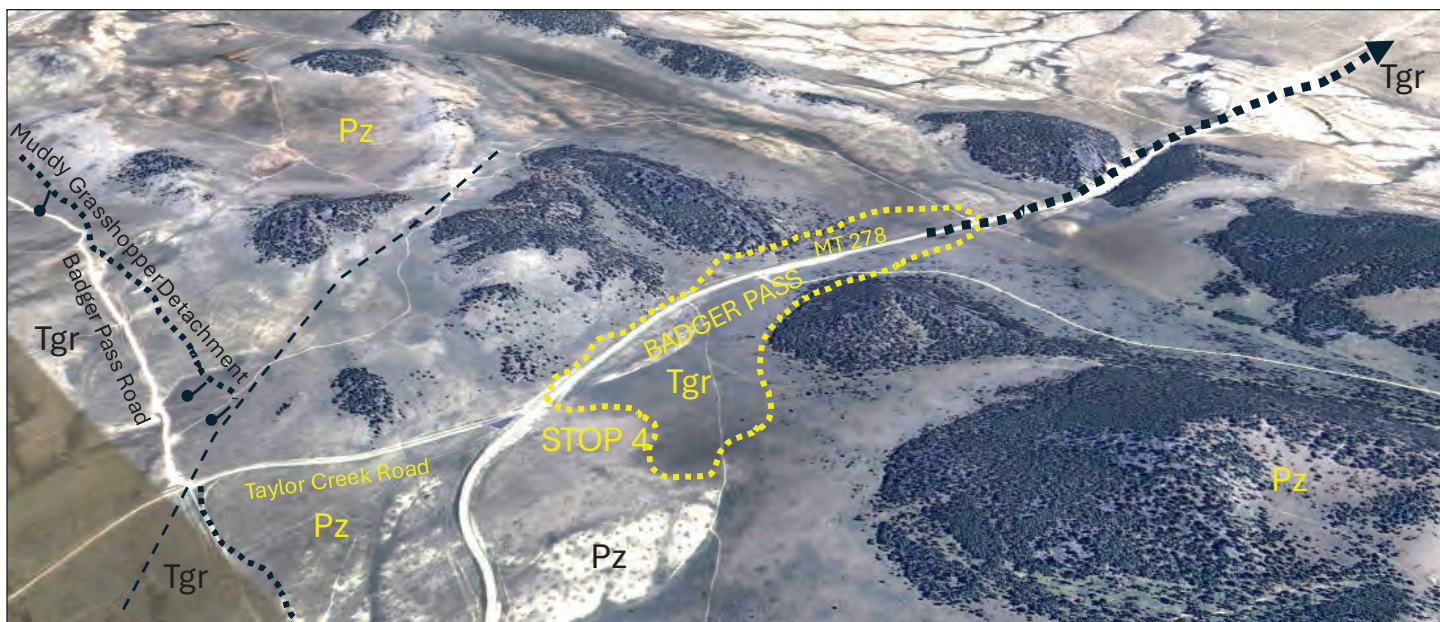


Figure 4. Area of STOP 4. View northeast. Intersection of Montana Highway 278 and Taylor Creek Road at Badger Pass. Badger Pass Tgr is a remnant of stream gravel that was cut off by the Muddy–Grasshopper detachment. Dotted black arrow shows flow path of paleo-stream from river gravel to alluvial fan. From Google Earth image.

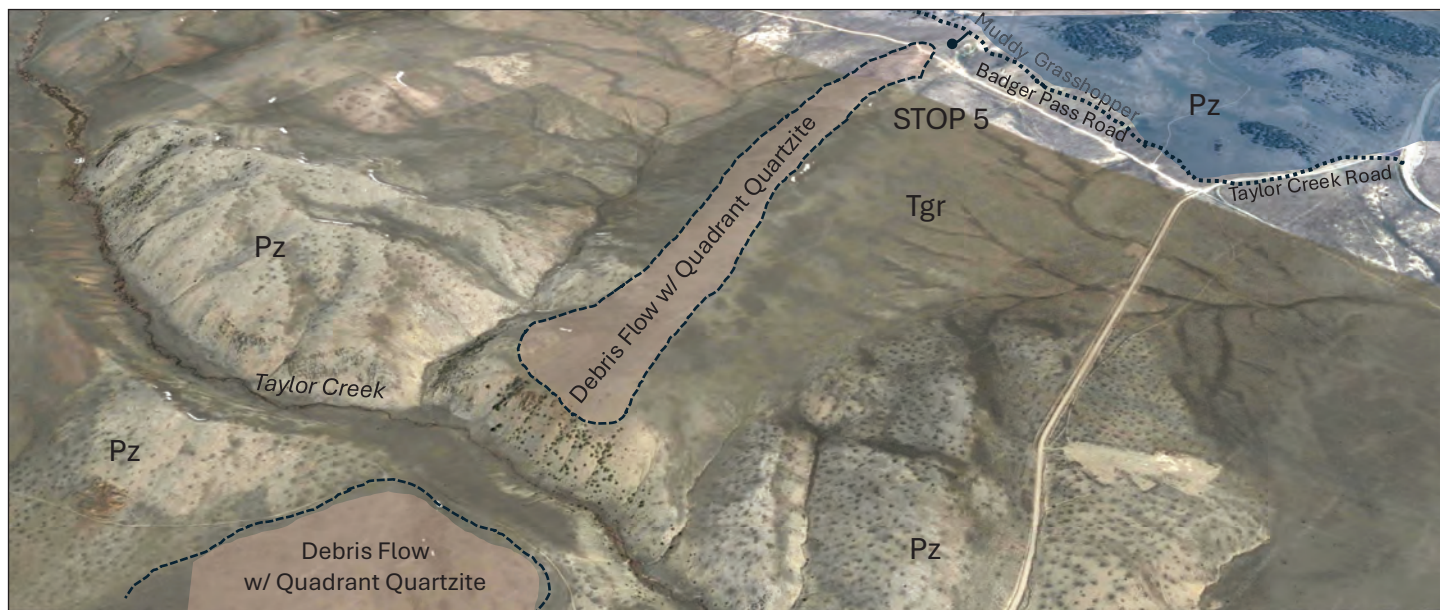


Figure 5. Area of STOP 5. View northeast. Debris flow cut by Pleistocene Taylor Creek Valley. The flow has large blocks of white Quadrant Quartzite eroded from the footwall of the Muddy–Grasshopper detachment. The blocks decrease in size southwest along the debris flow. From Google Earth image.

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FIELD GUIDE TO THE EOCENE VOLCANIC ROCKS IN BEAVERHEAD CANYON, SOUTH-WEST MONTANA

Jesse G. Mosolf and Raymond Salazar

Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

INTRODUCTION

This field guide explores the Eocene magmatic rocks of the Dillon Volcanic Group (DVG) exposed in Beaverhead Canyon, 16 km (10 mi) southwest of Dillon, Montana. The DVG erupted during a period of transitory tectonism as Cretaceous–Paleogene crustal shortening ended and the extensional collapse of the Cordilleran orogenic belt began. Eocene magmatism was widespread in Montana, erupting from several major vent complexes in the west-central part of the State (fig. 1). The DVG was a relatively small volcanic complex bridging two major volcanic fields: the Challis field to the west and the Absaroka field to the east. Most of the DVG's eruptive units are easily accessed along the I-15 corridor in Beaverhead Canyon, where spectacular 2D cross-sections of volcanogenic deposits can be viewed.

Beaverhead Canyon is northeast-trending and approximately 17.5 km (10.9 mi) long, extending northeast from Clark Canyon Reservoir to the Blacktail Deer Mountains escarpment (fig. 2), with the canyon walls exceeding 300 m (1,000 ft) of vertical relief. Several strath terraces along the modern Beaverhead River are covered by remnants of the Sixmile Creek Formation, constraining the onset of canyon incision to middle-Miocene time. The river canyon breaches the Blacktail fault escarpment at Barretts Rock, shaping the “Beaverhead Canyon Gateway” first described by the Lewis and Clark expedition in August 1805. The expedition traveled upriver through Beaverhead Canyon to “Camp Fortunate” at the confluence of Red Rock and Horse Prairie Creeks, declaring it the upper Missouri River's navigable limit (Moulton, 1983–2001). Today, Beaverhead Canyon is an important transportation corridor in southwest Montana, accommodating the I-15 freeway and Pacific Union railway, which are frequently impacted by mass wasting hazards, including several large landslide complexes impinging the roads and railway.

This day-long trip begins at Clark Canyon Reservoir, travels northeast through Beaverhead Canyon, and ends at Barretts Rock, with an optional stop at the Dillon West Rhyolite a couple miles northwest of Dillon. The field trip stops will generally traverse the DVG units from oldest to youngest. Participants are urged to be cognizant of traffic—most stops are along narrow paved and gravel roads commonly congested by recreational tourism. Three stops entail moderate off-trail hikes up to 1.6 km (1 mi) long. Participants should bring food, water, and off-trail hiking apparel.

GEOLOGIC SETTING

Regional Setting

Eocene volcanogenic rocks in Beaverhead Canyon are part of the regional Challis–Kamloops volcanic belt (CKVB) that stretches over 10 degrees of latitude in the North American Cordillera and exhibits an equally wide array of magmatic assemblages. The CKVB is defined by several major volcanic fields comprising the Buck Creek, Kamloops, Penticton, Princeton, Colville, Clarno, Challis, Absaroka, and Montana alkalic fields, most of which were erupted ca. 55–40 Ma and have high-potassium calc-alkaline geochemical signatures (e.g., Breitsprecher and others, 2003). Volcanic centers in Montana exemplify Eocene magmatic diversity, ranging from silicic granitoids in the western part of the State to highly alkaline intrusions in Late Cretaceous sedimentary rocks in the Cordilleran foreland to the east (fig. 1). Tectonomagmatic models attempt to reconcile spatially and compositionally diffuse Eocene magmatism in the southern CKVB, with most linking magmatism to North America's active plate margin. More traditional models portray magmatism at a subduction margin, entailing flat slab subduction followed by slab rollback (e.g., Constenius, 1996), slab foundering (Humphreys, 1995; Feeley, 2003), and subduction of a slab window (Breitsprecher and others, 2003). Other models don't invoke melts generated by subduction processes but instead suggest that widespread magmatism in the southern CKVB is the product of heating and decompression melting of the crust in response to the extensional collapse of the Cordilleran orogenic belt (e.g., Meen and Eggler, 1987; Dudas, 1991; Norman and Mertzman, 1991; Morris and Hopper, 1997). Many of the Eocene magmatic centers in the northern Cordillera overlap regions of pronounced crustal attenuation and the formation of metamorphic core complexes, perhaps signifying a relationship between magmatism and crustal extension (Foster and others, 2007).

Local Setting

The oldest rocks mapped in Beaverhead Canyon are Paleozoic sedimentary sequences deposited during marine transgressive–regressive cycles (figs. 2, 3). The Paleozoic rocks are unconformably overlain by Triassic through Late Cretaceous continental deposits that accumulated in the foreland basin of the eastward advancing Cordilleran fold-thrust belt. Conglomeratic deposits of the Beaverhead Group rest on an unconformity truncating the older Paleo-

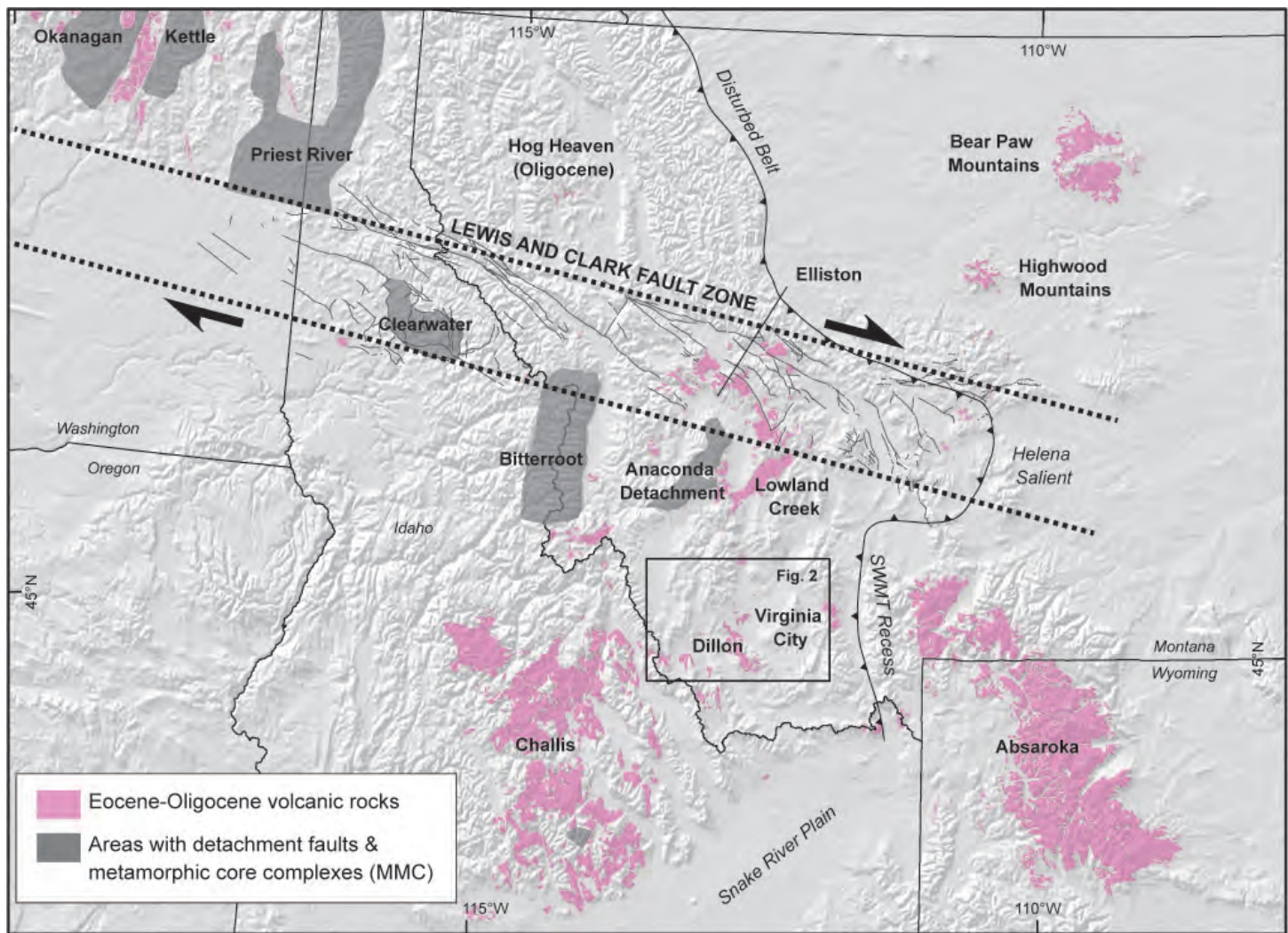


Figure 1. Regional tectonostratigraphic map showing the regional extent of Eocene volcanic rocks of the Challis–Kamloops volcanic belt and areas of high-magnitude crustal extension in northern Cordillera. Major volcanic centers are labeled. The location of figure 2 is shown by the black box. The simplified geologic overlay was compiled from Case and others (2013), Lewis and others (2012), Massey and others (2005), Schuster (2005), Vuke and others (2007), and Walker and MacLeod (1991).

zoic–Mesozoic units, marking the arrival of the Cordilleran thrust belt and synorogenic sedimentation within the local Armstead and Ermont thrust systems. Cenozoic extensional faults overprint the shortening structures in the Beaverhead Canyon area, including the Muddy–Grasshopper fault, the area’s most significant extensional fault system, stretching over 90 km (56 mi) and connecting the Grasshopper and Muddy Creek basins. The DVG units erupted in the foot-wall of the Muddy–Grasshopper fault and may have been coeval with the early onset of crustal extension. Volcanism transitioned to dominantly continental sedimentation of the Bozeman Group during the late Eocene–Oligocene, followed by Basin and Range block faulting and continued sedimentation during the middle Miocene. Several extensional structures are active in the area, including the northwest-striking Quaternary Blacktail fault. Mass wasting deposits are prevalent in the area, and several large landslide complexes formed in the Eocene volcanic units have shed into Beaverhead Canyon.

Dillon Group Stratigraphy

Fritz and others (2007) grouped Eocene through Miocene sequences in southwest Montana as the Dillon Volcanic Member of the Renova Formation. Their grouping includes three submembers: (1) the Eocene “lower Dillon volcanics” (49–40 Ma); (2) the late Eocene to Oligocene “middle Dillon volcanics” (39–27 Ma); and (3) the early Miocene “upper Dillon volcanics” (22–17 Ma). Mosolf and McDonald (2022) mapped the Eocene volcanogenic sequences near Beaverhead Canyon as distinct formations of the DVG (fig. 3). Here, the DVG comprises thin and discontinuous high-K calc-alkaline volcanic intervals and associated subvolcanic intrusions emplaced circa 50–46 Ma (figs. 3, 4). The volcanic units rest on an erosional unconformity truncating Paleozoic–Mesozoic rocks deformed by shortening structures, signifying that local thrust-belt deformation had ceased by Eocene time. Eocene volcanic rocks of lower DVG erupted before significant Tertiary sedimentation occurred and are postulated to have



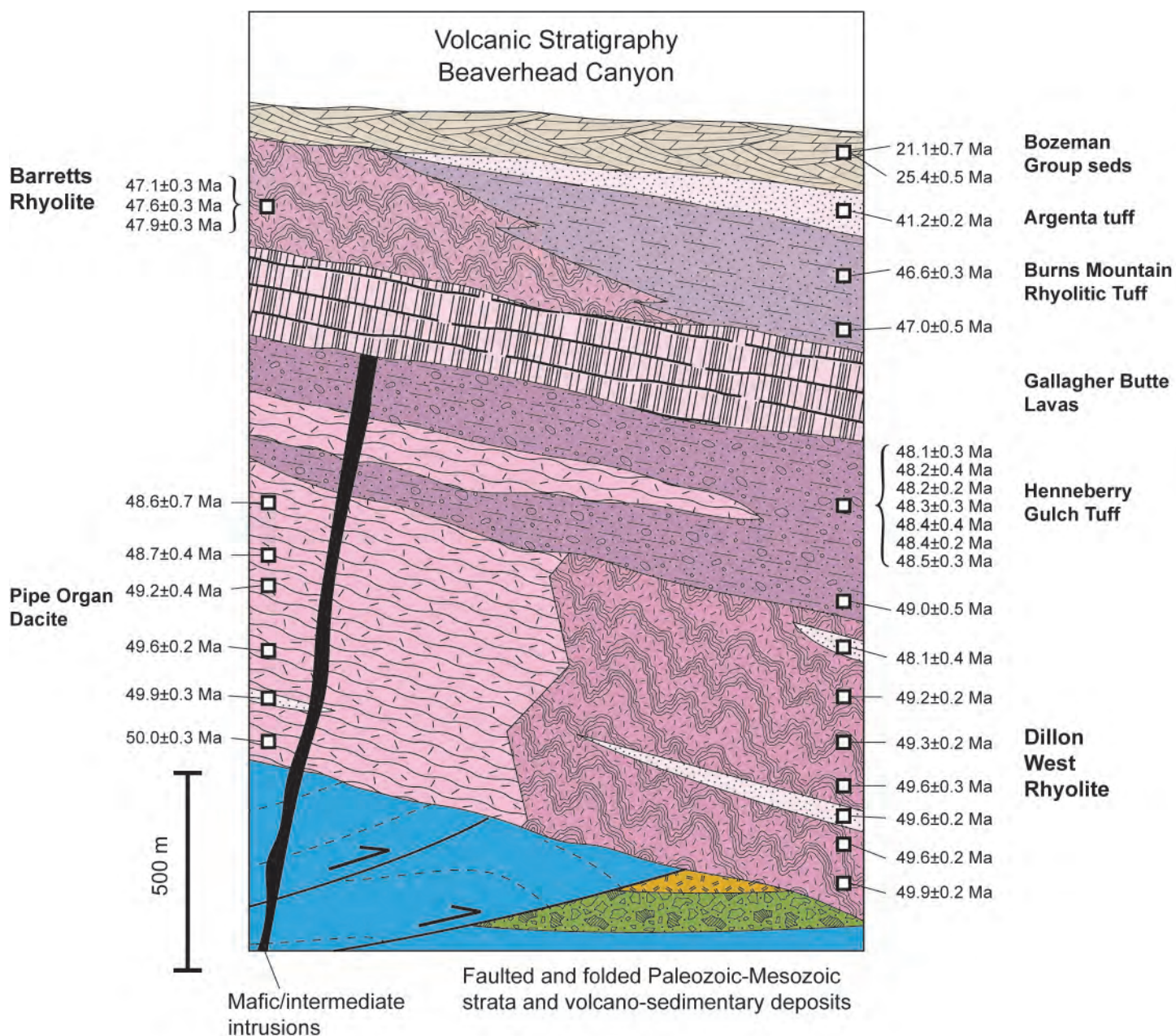


Figure 3. Schematic stratigraphy for the DVG in the Beaverhead Canyon area. Note that the stratigraphic thickness of the volcano-sedimentary units varies significantly. Numerical ages are U-Pb zircon ages reported by Mosolf and McDonald (2022). See table 1 for a complete list of ages and sample attribute data.

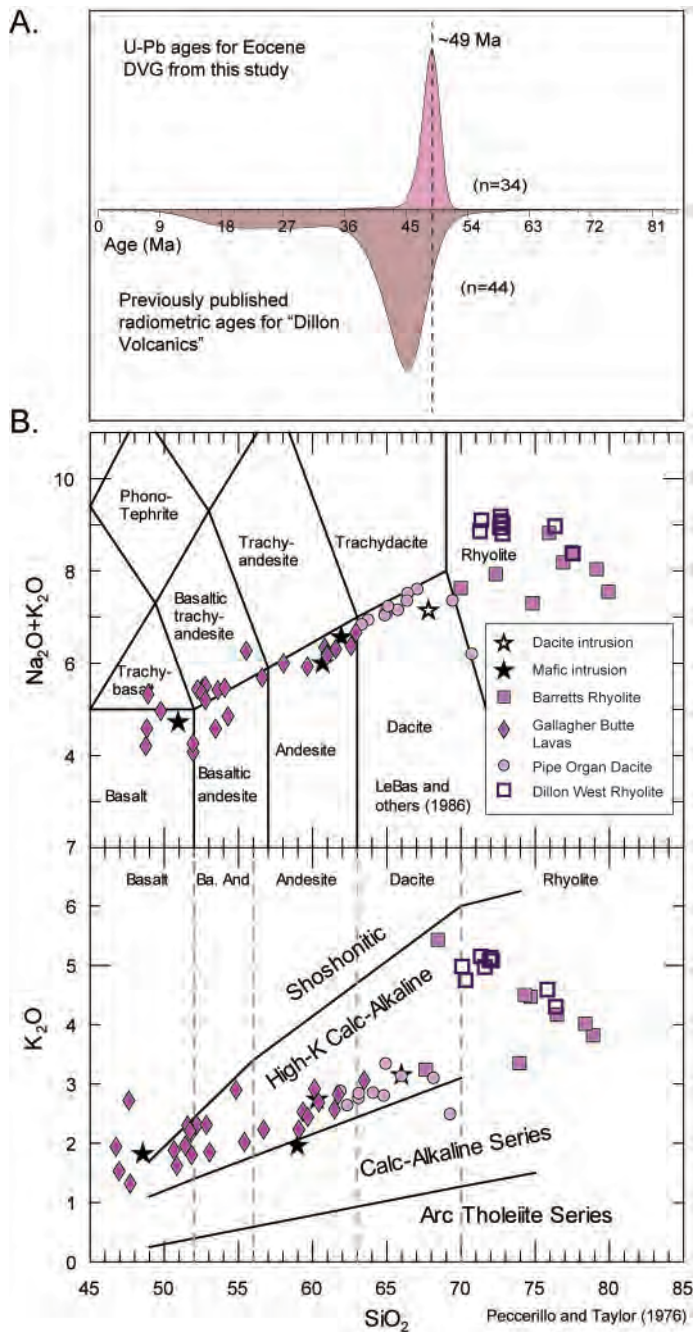


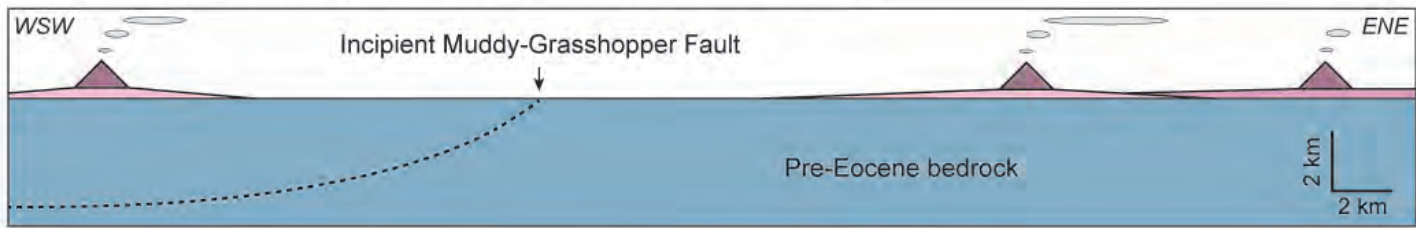
Figure 4. Absolute age and geochemistry data from the DVG in the Dillon 30' x 60' quadrangle (Mosolf and McDonald, 2022; McDonald and Mosolf, 2023; Mosolf, 2024). (A) Age distribution plots. The upper curve shows U-Pb age data from this study. The lower curve is K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ age data for the "Dillon Volcanics" published by Fritz and others (2007). (B) Whole-rock geochemical data from the DVG plotted on rock-type diagrams after LeBas and others (1986) and Peccerillo and Taylor (1976).

formed in a broad basin (Dillon–Renova basin) ringed by Eocene volcanic highlands (Fritz and others, 2007) or on the shoulder of a rift margin formed by supradetachment faults (Janecke, 1994; Janecke and others, 1999). The lower DVG units show some evidence of extensional deformation and are locally tilted and cut by brittle faults with various orientations, most of which were likely formed during Miocene Basin and Range-style faulting (Reynolds, 1979; Chadwick, 1980) and crustal doming related to the passage of the Yellowstone hotspot track in Late Miocene–Pleistocene time (Fritz and Sears, 1993). A subtle yet progressive angular unconformity within the DVG sequence suggests Eocene magmatism may have been concurrent with extensional faulting (fig. 5; Janecke and others, 1999).

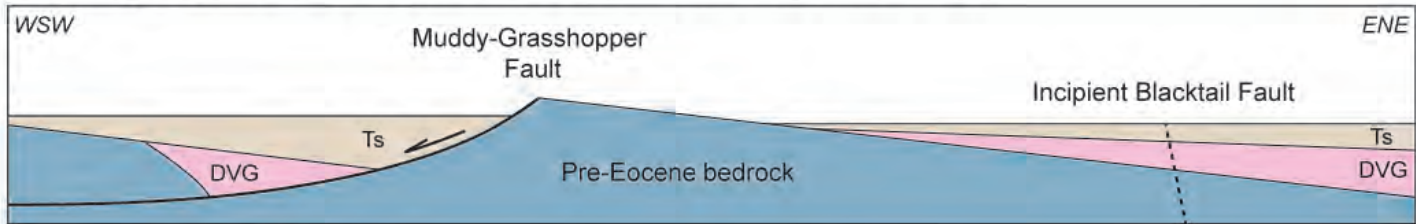
The age equivalent Pipe Organ Dacite and Dillon West Rhyolite are the oldest DVG units mapped in the Dillon area, which were erupted from separate vents at approximately the same time (fig. 3; table 1). The Dillon West Rhyolite comprises a chemically uniform sequence of rhyolite lavas and domes with minor tuff intercalations (49.9–49.2 Ma) underlying the low hills approximately 2 km (1.2 mi) northwest of Dillon. In Beaverhead Canyon, the Pipe Organ Dacite unit is a complex sequence of dacite–andesite flows (50.0–48.6 Ma) with rhyolite–dacite tuff intercalations (49.9–49.3 Ma). The Pipe Organ Dacite is the most recognizable DVG unit in Beaverhead Canyon, creating dramatic columnar jointed cliffs along the interstate corridor. The Henneberry Gulch Tuff (49.0–48.1 Ma) and overlying Gallagher Butte Lavas rest unconformably on the Pipe Organ Dacite unit but are thin and laterally discontinuous, with stratigraphic truncations and pinch-outs occurring over short distances. The Gallagher Butte Lavas are mostly mafic basaltic flows with diagnostic columnar jointing. The DVG sequence is capped by the Barretts Rhyolite (47.9–47.1 Ma) and comagmatic Burns Mountain Tuff (47.0–46.6 Ma). The silicic units are nearly concordant and contain distinctive quartz-eye phenocrysts indicative of a shared parental melt and/or vent. Mafic–intermediate intrusions and breccia pipes cut the Eocene volcanic rocks and country rock throughout the Beaverhead Canyon area. The DVG units generally exhibit high-K calc-alkaline geochemical signatures (fig. 4B) and contain abundant Precambrian zircon xenocrysts that reflect substantial fractionation, mixing, and assimilation of melts generated by an active plate margin. Local Eocene magmatism was likely contemporaneous with early extension on the Muddy–Grasshopper fault that daylights approximately 18 km (11.2 mi) west of Beaverhead Canyon (figs. 2, 5).



Eocene ca. 50 Ma: onset of volcanism and coeval(?) crustal extension



Eocene-Oligocene ca. 49-25 Ma: continued volcanism/extension; sedimentation



Miocene-present ca. 25-0 Ma: Basin-and-range faulting and sedimentation

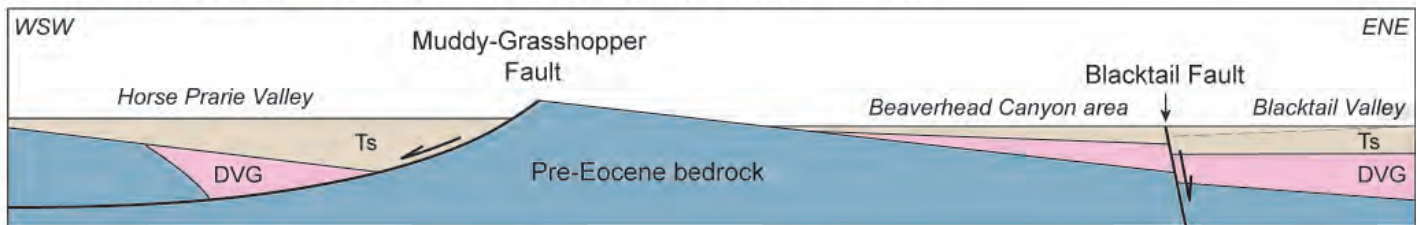


Figure 5. Conceptual model of volcanism and extensional faulting in the Beaverhead Canyon Area from the Eocene to the present. DVG, Dillon Group; Ts, Tertiary sediment.

PREVIOUS GEOLOGIC MAPPING AND RESEARCH

The field trip materials are mainly derived from a Montana Bureau of Mines and Geology mapping project supported by the STATEMAP component of the U.S. Geological Survey's National Cooperative Geologic Mapping Program. Recent MBMG mapping and analytical work is included in several draft and published quadrangle maps, including Burns Mountain (Mosolf, 2024, 1:24,000 scale), Dalys (Mosolf, 2016, 1:24,000 scale), Dillon West (McDonald and Mosolf, 2023, 1:24,000 scale), and the west half of the Dillon 30' x 60' (Mosolf and McDonald, 2022, 1:100,000 scale). Beaverhead Canyon and adjacent areas were included in a small-scale map by Ruppel and others (1993, 1:250,000 scale) and larger-scale maps by Lowell (1965, 1:31,680 scale), Coryell (1983, 1:12,000 scale), Johnson (1986; 1:24,000 scale), Tysdall (1988, 1:24,000 scale), and Leeth (1998, 1:24,000 scale).

Mapping of the Tertiary sedimentary units in the area was based on lithological descriptions by Ripley (1995), Thomas and others (1995), Fritz and others (2007), Sears and others (2009), Thomas and Sears (2020), and Vuke (2020). The Eocene volcanic map units were defined by lithology, stratigraphy, age, geochemical composition, and distribution reported by the MBMG and Fritz and others (2007). The informal subdivision of the Beaverhead Group follows the terminology defined by Lowell (1965)

and Johnson (1986). See Sears (2007) for a field guide to middle Miocene erosional surfaces, debris flows, and neotectonic faults in the Clark Canyon reservoir area.

ROAD LOG

The road log begins in the large parking lot on the southeast end of Clark Canyon dam, near the entrance to Beaverhead Campground and the Highway 324 and Interstate 15 interchange (44.9989°N, 112.8548°W). Figures 6 and 7 provide a simplified geologic map and legend for the field trip route. Mileage is cumulative driving distance; coordinates are in the WGS84 datum.

Drive southeast from the parking lot for 150 m (490 ft), following Highway 324 across the I-15 overpass. Turn left onto High Bridge Road and drive approximately 2.6 km (1.6 mi) to STOP 1, passing the Armstead Campground and several large boulders (2–3 m) of quartzite, metamorphic basement, and volcanic rock along the river. Park under the freeway overpass. This is a popular boat launch for fishermen and is commonly congested during the summer months.

STOP 1—Pipe Organ Dacite and Andesite 1.7 mi (45.0177°N, 112.8384°W)

The informal Pipe Organ Dacite and Tuff (Tpod) is the oldest DVG unit mapped in Beaverhead Canyon and comprises intermediate lava flows, intercalated tuffs,



TABLE 1. LA-ICPMS U-Pb zircon geochronology data from the Dillon 30' x 60' quadrangle

Sample	Lithology	Unit or Formation	7.5' quadrangle	Latitude (°N)	Longitude (°W)	Method	Age (Ma)	2 σ	MSWD*
JM19BM17	sandstone	Renova	Burns Mountain	45.1534	112.7843	WM 206/238	21.0	0.7	2.4
JM19BM16	sandstone	Renova	Burns Mountain	45.1503	112.7994	WM 206/238	25.4	0.5	3.1
DN-AR1	tuff	Renova	Argenta	45.2895	112.8186	WM 206/238	41.2	0.2	0.2
19B03	tuffaceous sand	Renova	Block Mountain	45.4331	112.5100	MDA 206/238	44.9	0.3	0.6
JM19BM08	tuff	Burns Mountain Tuff	Burns Mountain	45.1385	112.7921	WM 206/238	46.6	0.3	0.6
JM19BM27	tuff	Burns Mountain Tuff	Burns Mountain	45.1723	112.8129	WM 206/238	47.0	0.5	0.5
19B01	tuff	Burns Mountain Tuff	Beaverhead Rock	45.4903	112.4841	WM 206/238	47.0	0.4	0.6
DN-31	rhyolite	Barretts Rhyolite	Dalys	45.1222	112.7601	WM 206/238	47.1	0.3	1.0
JM19BM01	rhyolite	Barretts Rhyolite	Burns Mountain	45.1263	112.7519	WM 206/238	47.6	0.3	0.9
DN-02	rhyolite	Barretts Rhyolite	Dalys	45.0913	112.7565	WM 206/238	47.9	0.3	1.3
DN-27	tuff	Henneberry Gulch Tuff	Dalys	45.0979	112.7936	WM 206/238	48.1	0.3	0.7
JM19DW10	rhyolite tuff	Henneberry Gulch Tuff	Dillon West	45.1306	112.7428	WM 206/238	48.1	0.4	1.4
DN-13	tuff	Henneberry Gulch Tuff	Dalys	45.1177	112.8122	WM 206/238	48.2	0.4	1.4
DN-17	tuff	Henneberry Gulch Tuff	Dalys	45.1141	112.7830	WM 206/238	48.2	0.2	1.0
DN-41	tuff	Henneberry Gulch Tuff	Dalys	45.0391	112.7634	WM 206/238	48.3	0.3	1.3
DN-30	tuff	Henneberry Gulch Tuff	Dalys	45.1162	112.7931	WM 206/238	48.3	0.3	1.1
JM19BM24	tuff	Henneberry Gulch Tuff	Burns Mountain	45.1697	112.8284	WM 206/238	48.4	0.4	0.8
DN-05	tuff	Henneberry Gulch Tuff	Dalys	45.0802	112.7726	WM 206/238	48.4	0.2	0.4
DN-33	tuff	Henneberry Gulch Tuff	Dalys	45.0100	112.8009	WM 206/238	48.5	0.3	0.5
DN-29	tuff	Henneberry Gulch Tuff	Dalys	45.1087	112.7887	WM 206/238	49.0	0.5	1.4
DN-45	dacite	Pipe Organ Dacite	Dalys	45.0551	112.7774	WM 206/238	48.6	0.7	1.3
DN-23	dacite	Pipe Organ Dacite	Dalys	45.0938	112.8393	WM 206/238	48.7	0.4	0.9
JM19BM23	dacite	Pipe Organ Dacite	Burns Mountain	45.1695	112.8155	WM 206/238	49.2	0.4	0.9
DN-38	tuff	Pipe Organ Dacite	Dalys	45.1139	112.7609	WM 206/238	49.3	0.4	1.8
JM19BM35	tuff	Pipe Organ Dacite	Burns Mountain	45.1567	112.8642	WM 206/238	49.4	0.4	0.8
DN-46	dacite	Pipe Organ Dacite	Dalys	45.0685	112.8014	WM 206/238	49.6	0.4	0.5
DN-47	dacite	Pipe Organ Dacite	Dalys	45.0997	112.8117	WM 206/238	49.6	0.3	1.3
DN-40	tuff	Pipe Organ Dacite	Dalys	45.0178	112.8355	WM 206/238	49.9	0.3	1.0
JM19BM40	dacite	Pipe Organ Dacite	Burns Mountain	45.1760	112.8544	WM 206/238	50.0	0.3	1.0
JM19DW04	rhyolite	Dillon West Rhyolite	Dillon West	45.2385	112.6697	WM 206/238	49.2	0.2	1.2
KM19DW28	rhyolite tuff	Dillon West Rhyolite	Dillon West	45.2179	112.7048	WM 206/238	49.3	0.2	0.8
KM19DW50	rhyolite tuff	Dillon West Rhyolite	Dillon West	45.1859	112.7020	WM 206/238	49.6	0.2	1.0
JM19DW03	rhyolite	Dillon West Rhyolite	Dillon West	45.2350	112.6950	WM 206/238	49.6	0.3	0.8
JM19DW09	rhyolite	Dillon West Rhyolite	Dillon West	45.2152	112.6667	WM 206/238	49.6	0.3	0.7
JM19DW06	rhyolite	Dillon West Rhyolite	Dillon West	45.2494	112.6566	WM 206/238	49.9	0.2	0.9

*MSWD is the Mean Square Weighted Deviation.

Method: WM 206/238 weighted mean of select $^{206}\text{Pb}/^{238}\text{U}$ dates
MDA 206/238 max depositional age, weighted mean of youngest $^{206}\text{Pb}/^{238}\text{U}$ dates

Data compiled from Mosolf and McDonald (2022)

Zircon separates were prepared at MBMG and analyzed by LA-ICPMS at the University of California, Santa Barbara.

Latitudes and longitudes are in the 1984 World Geodetic Survey (WGS84) datum.



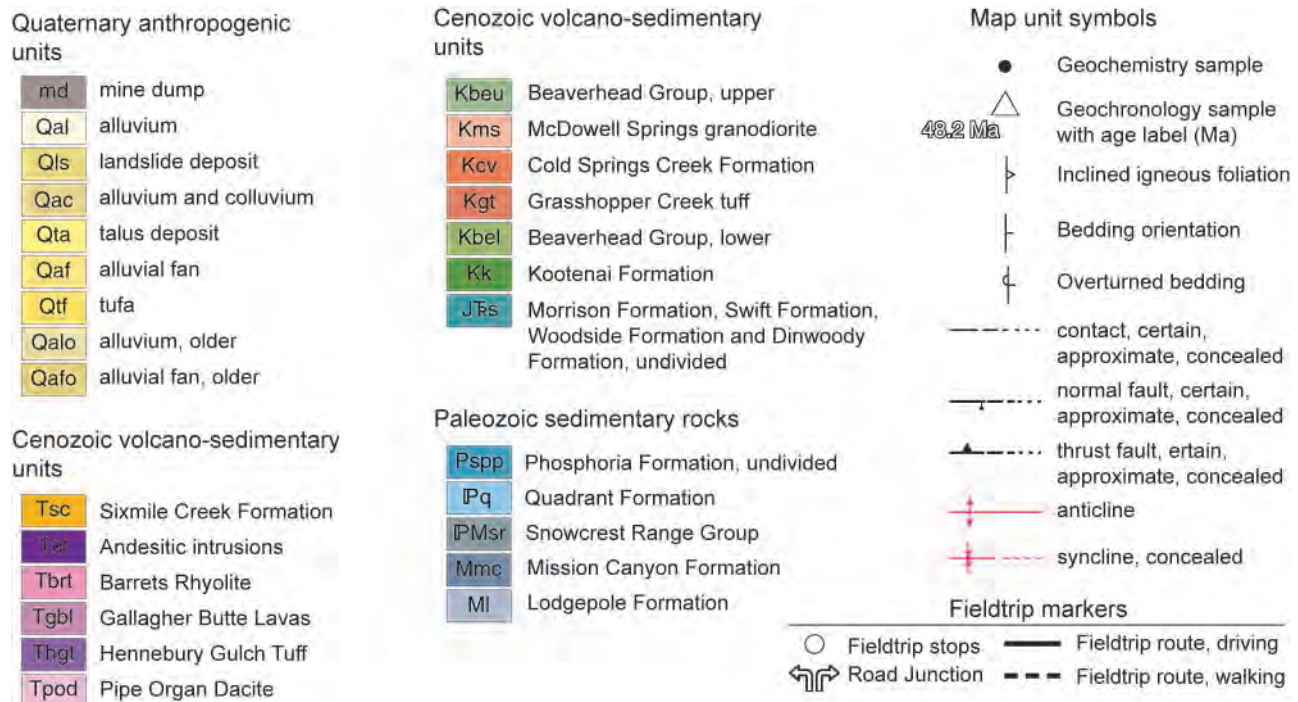


Figure 6. Legend for geologic maps in figures 7, 9, and 17.

and subvolcanic intrusions emplaced ca. 50–48 Ma. Here, the unit is a dark to light gray aphanitic rock containing less than 1 percent xenocrystic quartz. A strong igneous foliation and systematic vertical fractures give the outcrop a flaggy appearance, weathering the rocks into plates and blocks (fig. 8A). Geochemical and geochronological data were not obtained at this locale; however, textural similarities and superposition of the tuff dated at STOP 1a suggest the rock may be a dacite or andesite correlative to the Pipe Organ Dacite unit discussed in further detail at STOP 2. It's unclear if this outcrop was formed from a lava or intrusion.

Looking west across the river, the toe of a large landslide complex impinges the Beaverhead River. The landslide deposit formed in unconsolidated debris flow deposits of the Sixmile Creek Formation (Sears, 2007). The unconsolidated deposits can be traced upslope to a gap in the Madison Group bedrock and westward across the core of the Armstead anticline to another gap in the Quadrant Formation. The debris flows likely flowed westward through a paleovalley crossing the Armstead anticline to the proto-Beaverhead Canyon, where they were mixed with fluvial gravel and transported downstream.

Walk around the southern end of the dacite outcrop and climb the road switchback to the railroad grade (fig. 9). Turn north and climb the small hill on the westerly side of the railroad grade. **Stay off the railroad tracks.**

STOP 1a—Vent Complex, Basal Sixmile Creek Formation Unconformity, and the Monument fault 0.25 mi hike (45.0176°N, 112.8366°W)

The railroad grade cut exposes a cross-sectional profile of a near-vent volcanic sequence (fig. 8B). Dacite–andesite intervals with compositional textures similar to the previous stop occur at the base of the sequence and are overlain by pink, gray, and white weathering volcanic breccia consisting of chaotic block and ash deposits several meters thick. Breccia clasts are monolithologic and resemble the intact dacite–andesite outcrop visited at STOP 1. The breccias are likely the result of near-vent phreatic explosions. The lavas and breccia are intruded by a dark gray stock several meters wide with compositional textures resembling the outcrop at high bridge. The stock is interpreted to be a volcanic plug formed in a vent that sourced the enveloping lava–breccia sequence. The volcanic breccias are capped by a thin, conspicuously white rhyolitic tuff containing 10–15 percent phenocrysts of quartz, biotite, hornblende, and sanidine (fig. 8B) that yielded a U–Pb zircon age of 49.9 ± 0.4 Ma. The volcanic units are tilted eastward, possibly by block rotation in the footwall of the Muddy–Grasshopper fault. To the northeast, the Pipe Organ Dacite is overlain by the Hennebury Gulch Tuff and Gallagher Butte Lavas visited at STOP 3.

Near the top of the railroad cut, polymictic cobble to boulder conglomerate deposits of the Sixmile Creek Formation overlie the Eocene volcanic sequence in angular unconformity (fig. 8C). The channel-fill conglomeratic deposits are massive, consisting of poorly



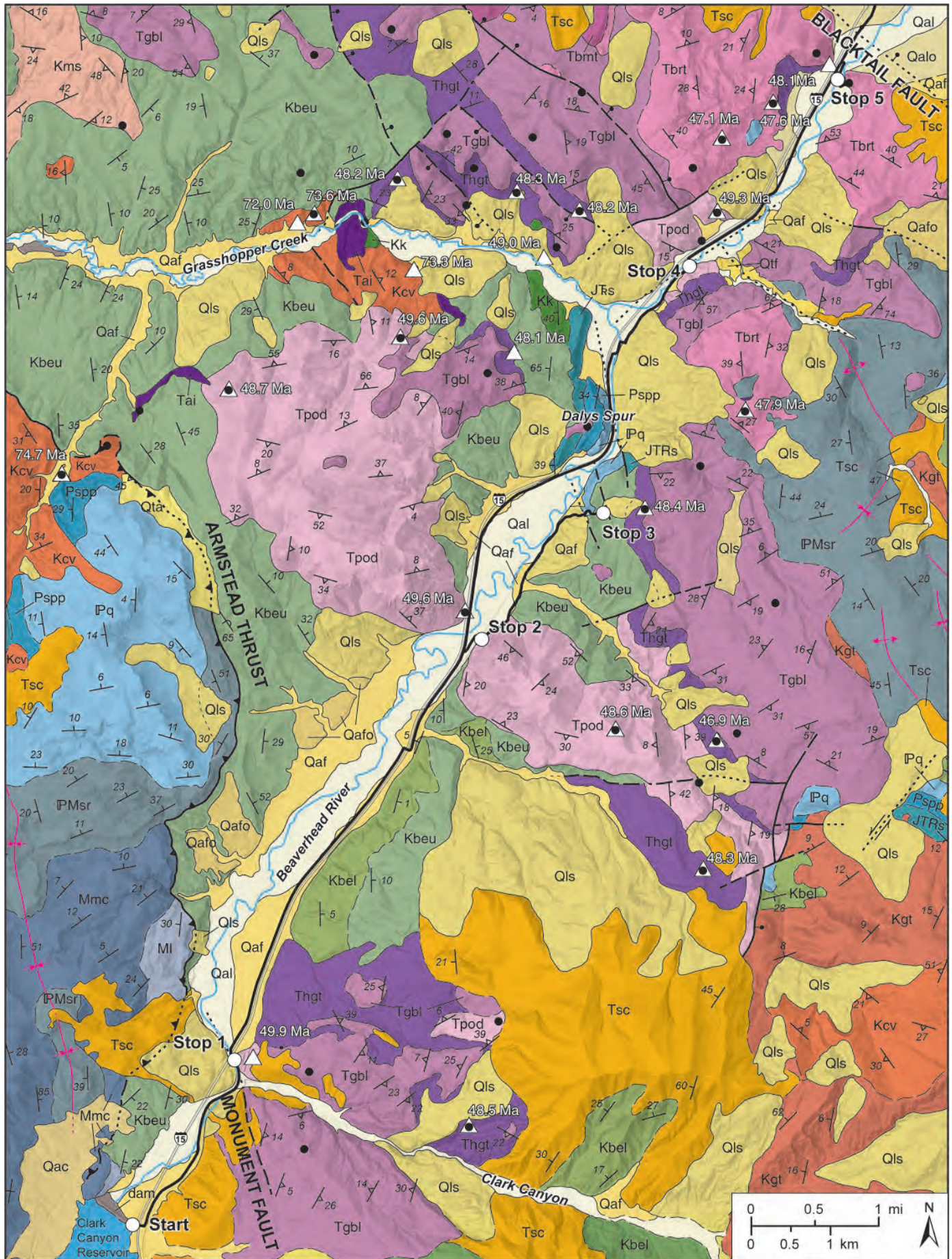


Figure 7. Simplified geologic map of Beaverhead Canyon showing the field trip route. Modified from Mosolf and McDonald (2022).



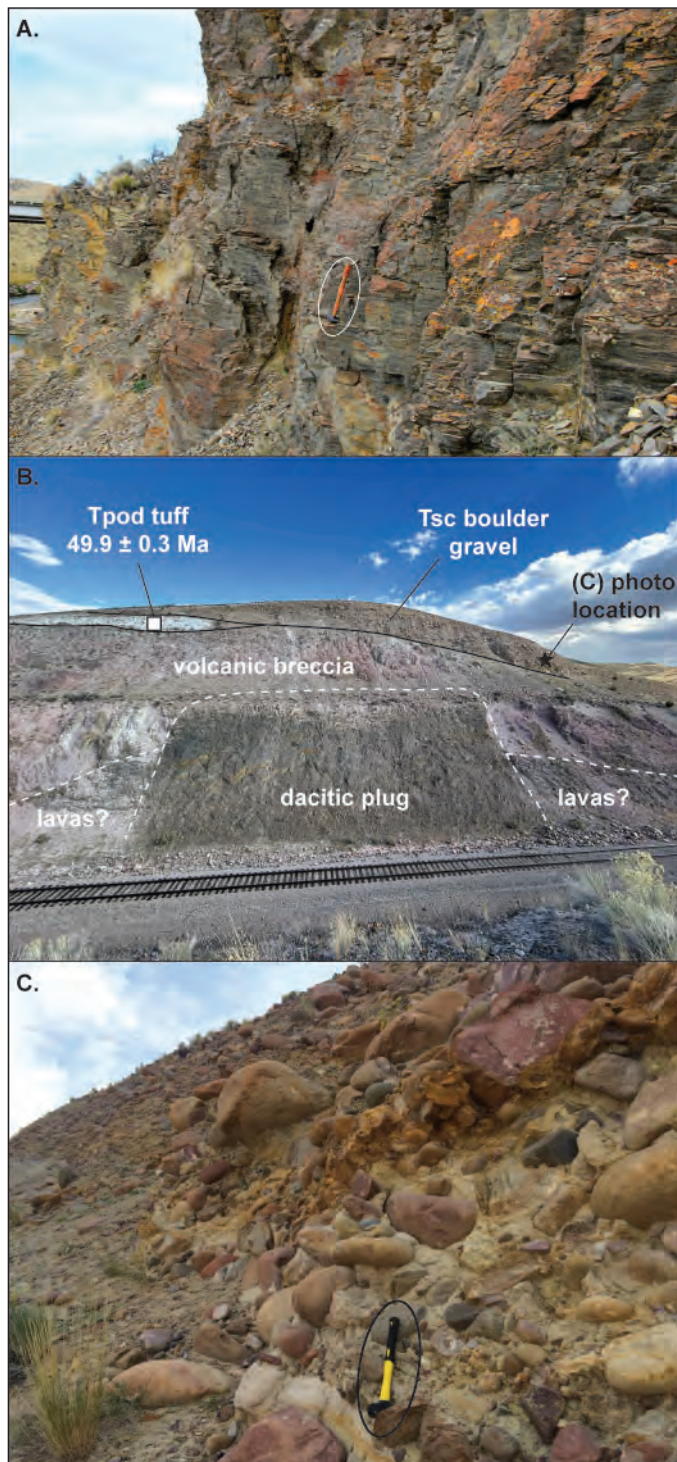


Figure 8. (A) Flaggy outcrops of volcanic or subvolcanic dacite along High Bridge Road at STOP 1; correlated to the Pipe Organ Dacite unit (Tpod). (B) Railroad cut through a dacitic plug overlain by volcanic breccia. A thin layer of white dacitic tuff (Tpod; 49.9 Ma) caps the volcanic sequence. Poorly consolidated gravel deposits of the Sixmile Creek Formation (Tsc) rest on the volcanic sequence in angular unconformity. (C) Sixmile Creek gravel deposit with abundant boulder clasts of subrounded quartzite. The location of the photo is shown in (B).

sorted, subrounded to rounded boulders (up to 1 m) of quartzite and Eocene volcanic rock. The deposits were likely deposited by a high-gradient river system and associated debris flows. The conglomeratic sequence was gently tilted southward by the neotectonic Monument fault (Sears, 2007). Southward, the trace of the fault approximately follows the trend of a small unnamed gulch, juxtaposing the Sixmile Creek Formation (west) against the Gallagher Butte Lavas (east) with a minimum down-to-the-southwest vertical displacement of 70 m (230 ft). The northern trace of fault projects under the freeway overpass, where it is obscured by modern river alluvium.

Walk northward, contouring around the hill back to the vehicles, inspecting dacite and breccia outcrops along the way. Continue driving northward on the frontage road for 2.6 km (1.6 mi). Looking northwest across the river, the Armstead thrust places Madison Group carbonates (gray) over clastic sediments of the Beaverhead Group (red). The red beds on the southeast side of the road are also Beaverhead Group. Turn right at the Heneberry Fishing access site, drive under the freeway overpass, and continue on the frontage road for another 1.5 km (1 mi). Turn right at the Pipestone Fishing access site and drive another 350 m (0.2 mi) towards the railroad crossing. Park in the small pullouts on either side of the railroad tracks. Exit the vehicles and walk to the outcrop on the southeast side of the tracks. Be cognizant of railroad traffic.

STOP 2—Pipe Organ Dacite **5.7 mi (45.0654°N, 112.7988°W)**

Dramatic red/maroon cliffs of dacite lavas, subvolcanic intrusions, and subordinate tuff deposits composing the Pipe Organ Dacite unit can be seen on both sides of Beaverhead Canyon (figs. 10A, 11A). The volcanogenic deposits unconformably rest on and possibly intrude the Beaverhead Group, indicating local Cordilleran shortening had ceased by Eocene time (fig. 10A). Aphanitic to porphyritic dacite–andesite lavas and intrusions are commonly gray to red weathering and strongly foliated. Aphanitic lavas generally have a trachytic texture consisting of plagioclase with minor ortho- and clinopyroxene microlites, and Fe-Ti oxides. Porphyries contain phenocrysts of plagioclase (10–20 percent; up to 4 mm); stubby to tabular hornblende (15–20 percent; up to 7 mm); and quartz (<5 percent; up to 3 mm) in a gray to pink groundmass of weakly aligned plagioclase and subordinate Fe-oxide microlites. Quartz and plagioclase phenocrysts are commonly embayed or skeletal, and possibly xenocrystic. Porphyritic rock is very dense and weathers into resistant blocks. Lava flows are locally flow-banded and separated by zones of autobreccia up to 4 m (13 ft) thick. Intercalations of

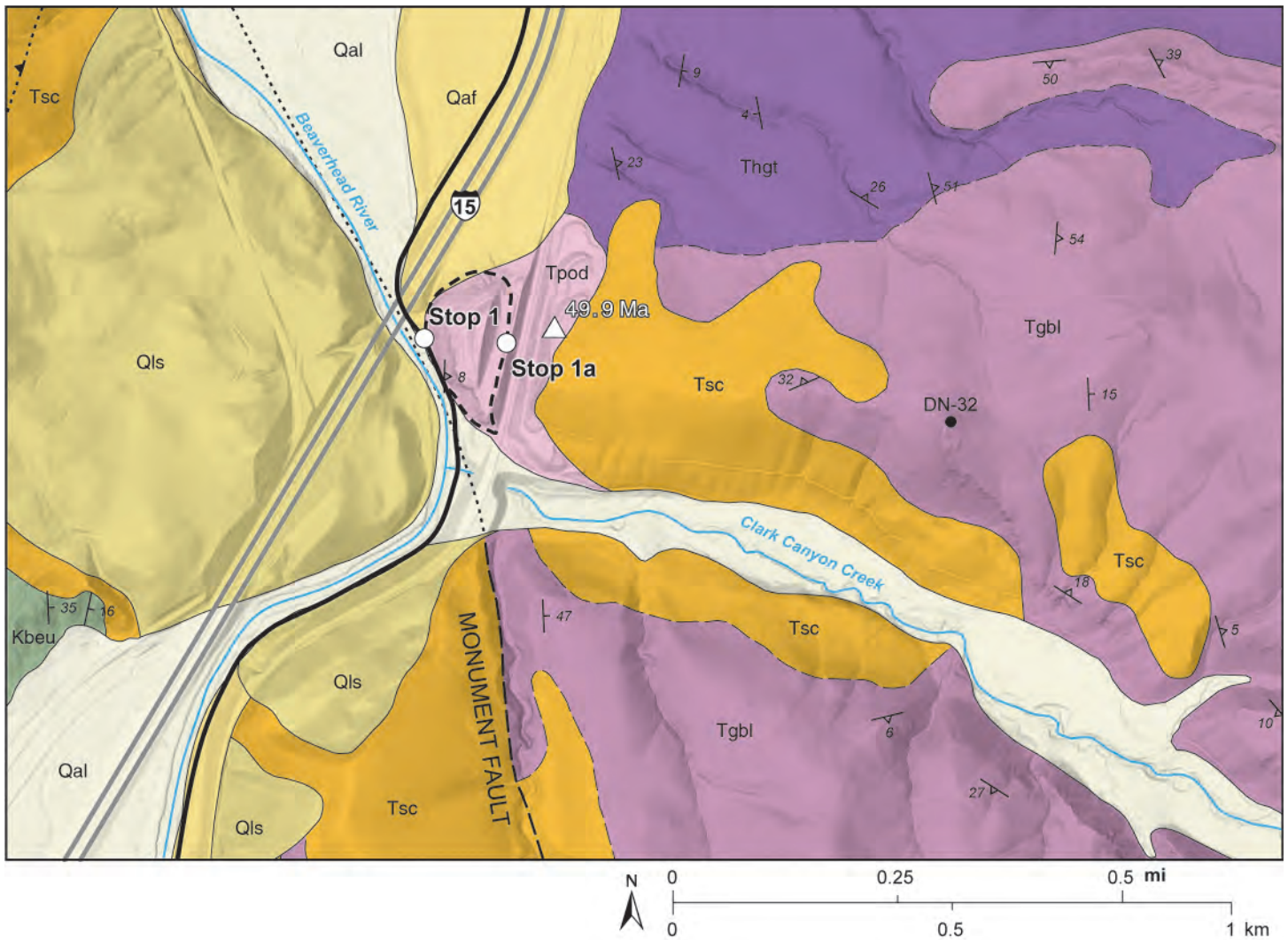


Figure 9. Simplified geologic map showing the walking route for STOP 1. See figure 6 for the map legend. Modified from Mosolf and McDonald (2022).

subordinate tuff exhibit similar mineral assemblages. Pronounced columnar jointing is observed throughout the map unit and is well developed at Pipe Organ Rock, visible on the northwest side of the freeway (fig. 11A). U-Pb zircon ages for this unit span 50.0 ± 0.3 Ma to 48.6 ± 0.7 Ma. Leeth (1998) reported a stratigraphically unreconcilable K-Ar age of 46.8 ± 1.0 Ma for this unit. The dacite sequence is generally susceptible to mass wasting, and a large landslide can be seen on the south-west flank of Pipe Organ rock.

The Pipe Organ Dacite formed near a volcanic vent, likely as a series of lava domes erupted from a central volcanic neck. At river level, a near-vertical igneous foliation is observed on both sides of the freeway, suggesting the melts flowed vertically during neck emplacement and eruption. Several tuffcite dikes and breccia pipes cut the dacite sequences in the Pipe Organ area, which are typically formed by explosive gas-release and hydrothermal eruptions in an epithermal system associated with a volcanic neck (fig. 11B). The apparent eastward dip of the contact separating the

Beaverhead Group and the overlying Pipe Organ Dacite is the surface trace of a steeply dipping volcanic neck (fig. 10A). Alternatively, the contact is an eastward-dipping angular unconformity rotated in the footwall block of the Muddy–Grasshopper fault to the west (fig. 5). Breccia pipes hosted by the Pipe Organ Dacite will be viewed at STOP 4.

Return to the vehicles and continue driving north 1.9 km (1.2 mi) to the next road junction, crossing the railroad tracks and traversing beneath red weathering slopes underlain with landslide deposits formed in the Beaverhead Group on the southeast side of the road. Turn right and drive up the rough dirt road 536 m (0.35 mi) to STOP 3, parking in the large pullout beside a historic cabin (fig. 12).





Figure 10. (A) View north of the Pipe Organ Dacite unit (Tpod) and the northern trace of the Armstead thrust. The volcanic units unconformably rest on the upper Beaverhead Group (Kbeu), with an apparent eastward tilt. The Armstead thrust places Paleozoic strata over synkinematic conglomerate deposits of the Beaverhead Group. Recessive deposits of the Hennebury Gulch Tuff (Thgt) can be seen in the foreground. (B) Henneberry Gulch Tuff (Thgt) and Gallagher Butte Lavas (Tgbl) in the lower Grasshopper Creek drainage, viewed towards the north.

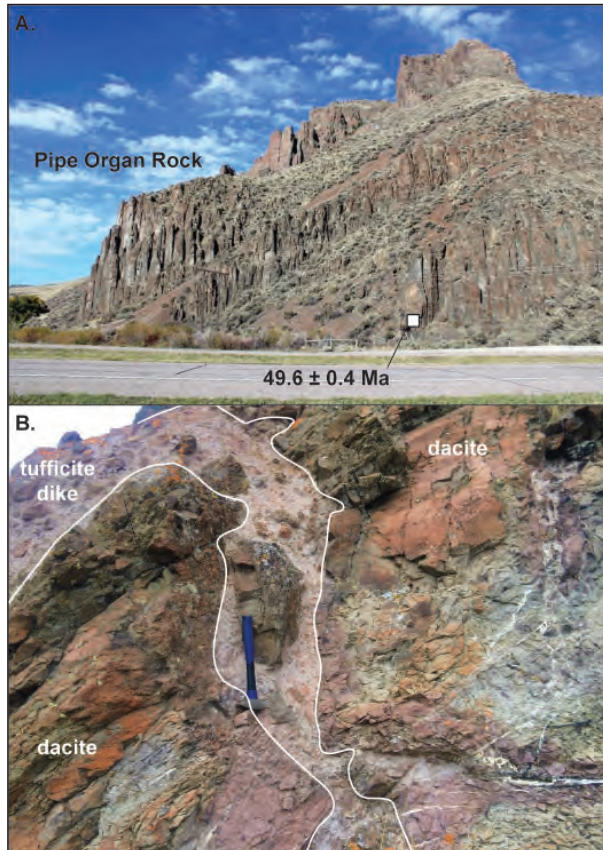


Figure 11. (A) Pipe Organ Dacite unit cropping out in the I-15 corridor south of Dillon. Note the well-developed columnar jointing, which gives the outcrops the appearance of an organ. (B) Tuffcite dike hosted by the Pipe Organ Dacite. Tuffcite forms through the violet gas release, commonly near a volcanic neck or vent.

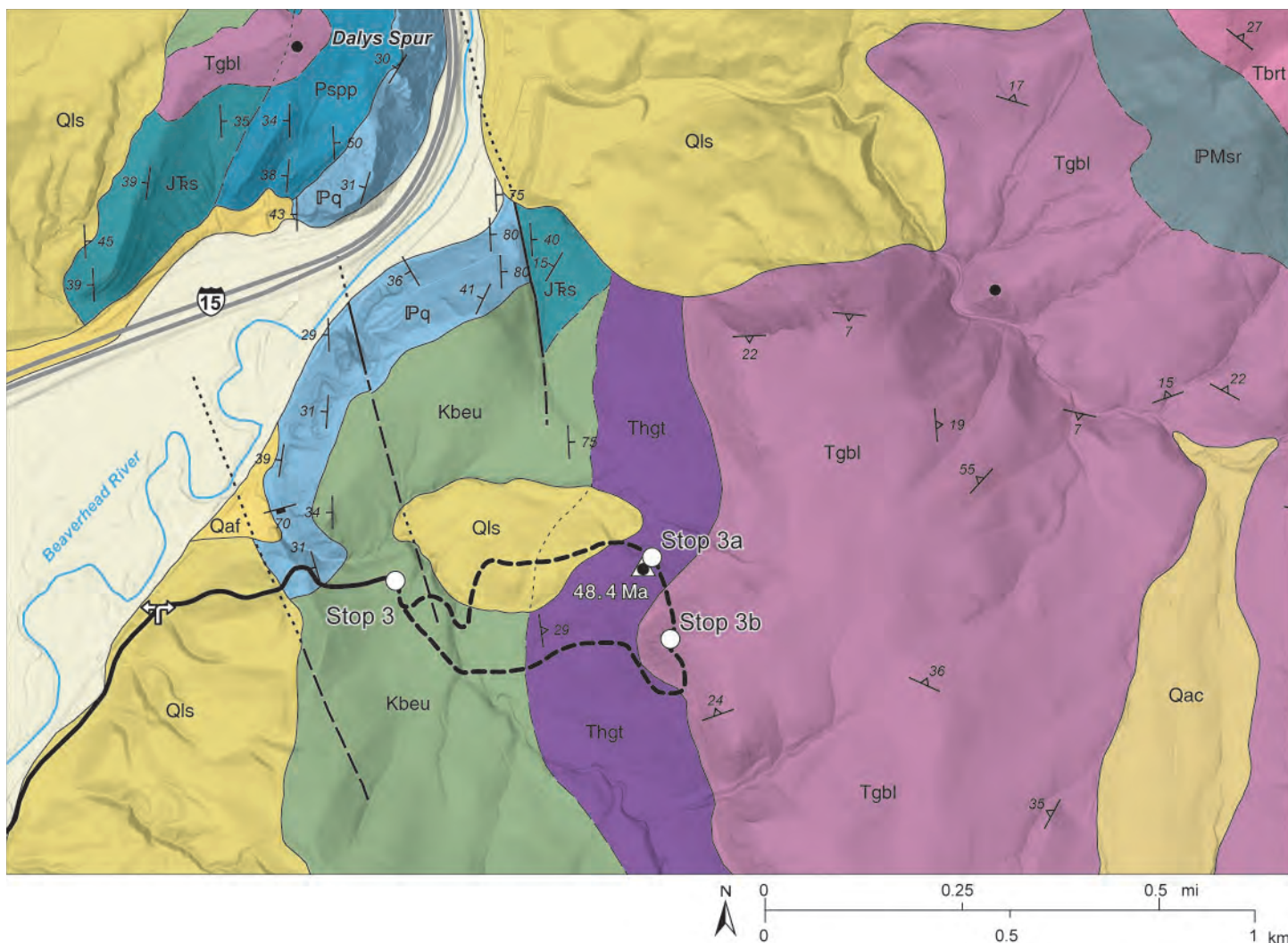


Figure 12. Simplified geologic map showing the hiking route at STOP 3. Modified from Mosolf and McDonald (2022).

STOP 3— Upper Beaverhead Group and Walking Tour of Eocene Volcanic Rocks

7.2 mi (45.0798°N, 112.7793°W)

Poorly lithified clastic deposits composed of rounded to well-rounded quartzite cobbles underlie the parking area at STOP 3. Rounded quartzite clasts are diagnostic of the Cretaceous Upper Beaverhead Group, or the Tertiary Sixmile Creek Formation. Here, the clastic deposits are likely part of the Beaverhead Group, but recycling of the Sixmile Creek Formation cannot be ruled out. Northwest of Beaverhead Canyon, towards Bannack, exposures of the Upper Beaverhead Group consist of poorly lithified, clast-supported conglomerate beds composed of well-rounded and imbricated pebbles and cobbles. Approximately 50–80 percent of conglomerate clasts are red to maroon weathering quartzite derived from the Belt Supergroup and Paleozoic units; the other clasts are predominantly Paleozoic limestone, chert, and rare dacite–andesite clasts that are deeply weathered. Conglomerate beds are 10–100 cm (4–39 in) thick and have tabular and lenticular geom-

etries. The stratal thickness of the unit exceeds 400 m (1,300 ft). Detrital zircon max depositional ages of 71.6 ± 3.4 Ma and 65.9 ± 1.3 Ma were reported for the Upper Beaverhead Group in the Dalys and Bannack 7.5' quadrangles, respectively (Laskowski and others, 2013; Garber and others, 2020). The Upper Beaverhead Group was synorogenic with local shortening, accumulating as a wedge in the foreland basin of the Armstead thrust system.

Looking eastward, cliffs of columnar jointed basalt rest on recessive slopes of white weathering tuff (fig. 13). Begin hiking on the road eastward and then climb off trail towards the white slopes, crossing landslide hummocks *en route* to STOP 3a. Look for pieces of petrified wood in the float.

STOP 3a—Henneberry Gulch Tuff Hike 0.5 mi (45.0798°N, 112.7792°W)

The Henneberry Gulch Tuff generally forms recessive slopes beneath the Gallagher Butte Lavas throughout the Beaverhead Canyon area. Where exposed, the unit is composed of a multi-storied sequence of buff, light





Figure 13. (A) Henneberry Gulch Tuff unit (Thgt) at STOP 3a that forms a recessive slope below the Gallagher Butte Lavas unit (Tgbl). (B) Petrified wood weathered from Thgt, abundant in the float at STOP 3a. (C) Columnar jointed basalt in the Gallagher Butte Lavas unit (Tgbl).

brown, and orange lithic-vitric tuff occurring in massive bedforms up to 2 m (6.6 ft) thick (fig. 13A). The tuff is often crystal-rich, containing broken phenocrysts (up to 30 percent) of quartz, feldspar, biotite, and occasional amphibole in a vitric groundmass composed of cusped glass shards that are mostly fresh but are devitrified locally. Small fragments (up to 5 mm; 0.2 in) of pumice, accidental volcanic clasts (up to 1 m; 3.3 ft), and petrified wood are commonly observed in the unit (fig. 13B). Thin intervals of planar, low-angle, cross-bedded tuffaceous sand and agglomerate beds occur locally. The Henneberry Gulch Tuff has a minimum thickness of 150 m (500 ft) with U-Pb zircon ages spanning 49.0 ± 0.5 Ma to 48.1 ± 0.3 Ma. A sample collected from this locale yielded an age of 48.4 ± 0.2 Ma. The Henneberry Gulch Tuff's indistinguishable radiometric ages and stacked bedforms suggest the sequence accumulated through rapid near-vent eruptions.

Continue climbing the hill, walking 180 m (0.1 mi) south toward the prominent basalt outcrop at Stop 3b.

STOP 3b—Gallagher Butte Lavas Hike 0.1 mi (45.0787°N, 112.7722°W)

The Gallagher Butte Lavas unit forms prominent outcrops and cliffs throughout the Beaverhead Canyon area and is named after the prominent butte in the Blacktail Mountains visible to the southeast. Basalt and basaltic andesite lavas comprise dark gray to black, strongly indurated rock; autobreccia and columnar joints are common. Lava flows can be porphyritic, containing <10 percent phenocrysts of olivine (1–3 mm), clino- and ortho- pyroxene (1–2 mm), with occasional plagioclase (1–2 mm) and xenocrystic quartz (1–2 mm). The aphanitic groundmass typically has a trachytic texture consisting of aligned microcrystals of plagioclase and sometimes includes microcrystalline olivine and pyroxene. Andesitic flows with embayed quartz occur locally. A basalt sample from Dalys Spur (fig. 12) yielded a K-Ar age of 42.8 ± 1.0 Ma (Leeth, 1998), which is not reconcilable with the volcanic stratigraphy mapped by Mosolf and McDonald (2022). Basalt and basaltic andesite occurrences are as thick as 120 m (400 ft).

The vantage point at STOP 3b offers spectacular views of the surrounding geology (fig. 14A). To the northwest, the prominent quartzite cliffs underlying Dalys Spur above I-15 are the Pennsylvanian Quadrant Formation. The Permian Phosphoria, Triassic Dinwoody, and Cretaceous Kootenai Formations overlie the Quadrant Formation, forming the eastern limb of a syncline in the foreland of the Armstead thrust (fig. 14B). The tilted panel is unconformably overlain by the Beaverhead Group and Eocene volcanic sequences. A thin interval of the Gallagher Butte Lavas caps Dalys



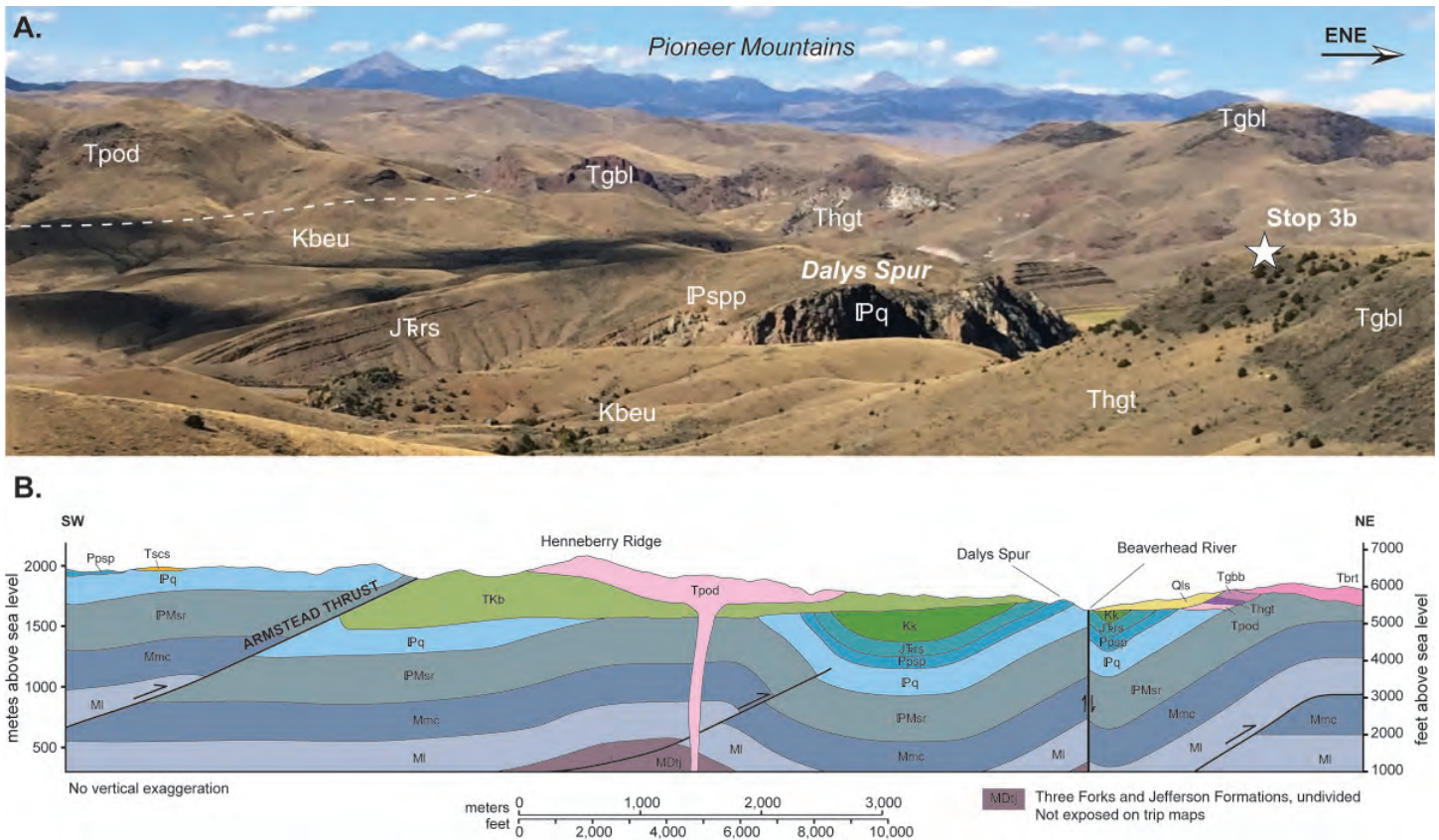


Figure 14. (A) View towards the northwest of the geology surrounding Dalys Spur and lower Grasshopper Creek. Stop 3b can be seen in the foreground. (B) Schematic geological cross-section through Dalys Spur. Cross-section line shown in figure 2. See figure 6 for symbol definitions.

Spur. The Henneberry Gulch Tuff and Gallagher Butte Lavas are visible north of Dalys Spur and Grasshopper Creek. The volcanic intervals are intruded by a volcanic neck that forms a prominent hoodoo. The volcanic units are deformed by northwest-striking normal faults with down-to-southwest displacements, similar to the orientation of the Monument fault viewed at STOP 1. To the north, the pointed hills forming the skyline are underlain by the Barretts Rhyolite visited at STOP 6. Several large landslide complexes formed in the Eocene volcanic units and shed into Beaverhead Canyon are visible to the north.

Walk back to the vehicles and dive back to High Bridge Road. Turn right and drive 2.4 km (1.5 mi) to the I-15 onramp. Merge onto the freeway and drive north 1.9 km (1.2 mi) to the Grasshopper Creek exit. Turn east onto the frontage road, cross the river and railroad tracks, and follow the dirt road northeast for 1.3 km (0.8 mi) to STOP 4. Park in the pullout next to the andesite outcrop.

STOP 4—Breccia Pipes in the Pipe Organ Dacite 13.5 mi (45.0793°N, 112.7854°W)

The outcrop near the parking area is dark gray andesite with sparse quartz xenocrysts correlated to the Pipe Organ Dacite unit observed at STOPS 1 and 2. The rock contains a steeply dipping igneous foliation and

is deformed by systematic, near-vertical fractures. To the northwest across the freeway, several breccia pipes can be seen cross-cutting the Pipe Organ Dacite unit (fig. 15). The breccia pipes have sharp, vertical contacts and are composed of poorly sorted, angular to subangular clasts exceeding 80 cm (32 in) in diameter that are compositionally identical to the country rock. Silicification and sulfide mineralization of the breccia pipes through hydraulic fracturing and phreatic brecciation. The age of brecciation is uncertain but is assumed to be hydrothermally linked to the emplacement of the volcanic neck at Pipe Organ rock. The breccia pipes cross-cut the easterly tilted biotite-lapilli tuff tilted visibly across the freeway, constraining the breccia's maximum emplacement age to 49.3 ± 0.4 Ma (fig. 15A).

Return to the vehicles and continue driving northeast on the frontage road 3.4 km (21 mi) to Barretts Park Campground. Turn right and park in the visitor area. Do not block access to the boat ramp.



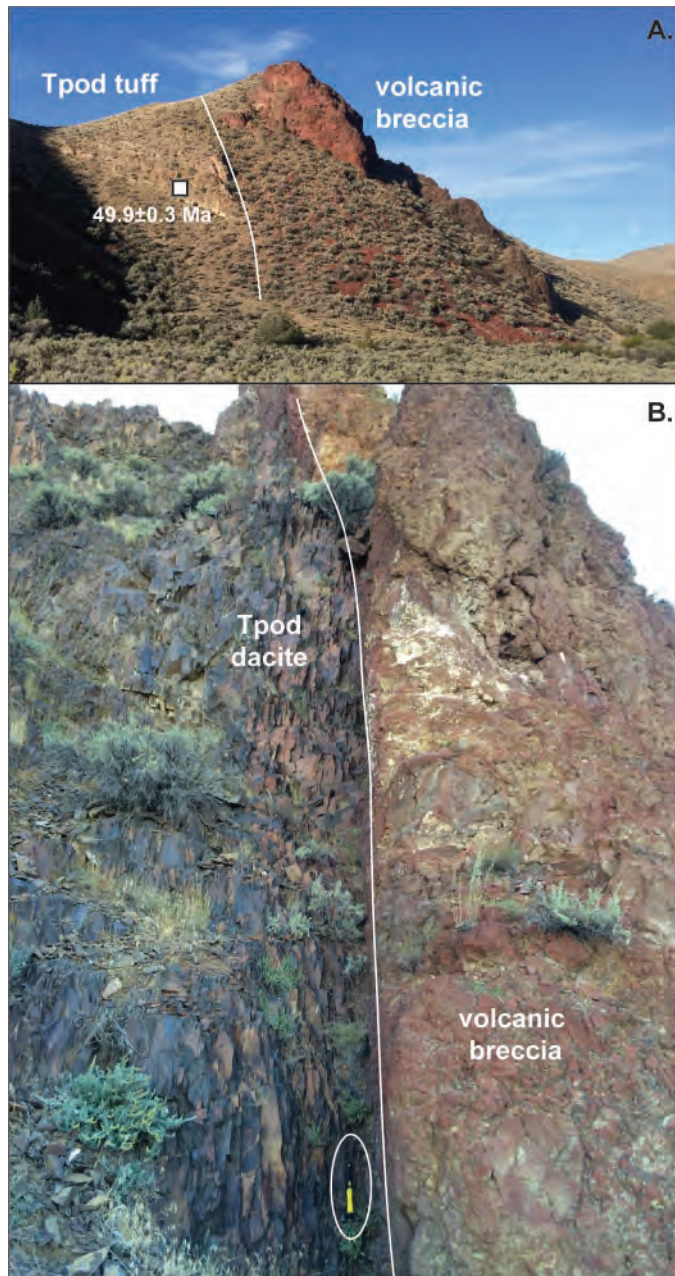


Figure 15. (A) Volcanic breccia pipe cross-cutting tilted tuff beds of the Pipe Organ Dacite unit (Tpod). (A) U-Pb zircon age of approximately 49.9 Ma was obtained for the tuff. (B) Volcanic breccia pipe hosted by Tpod. The dacite has a pronounced vertical foliation.

STOP 5—Barretts Rhyolite 15.7 mi (45.1290°N, 112.7415°W)

The Barretts Rhyolite unit forms the towering cliffs seen on both sides of the river. It is cut by the north-west-striking, Quaternary-active Blacktail fault, with an undetermined amount of down-to-the-northeast displacement, forming the northeast-facing Blacktail Mountains escarpment. The river canyon breaches the fault escarpment at this locale, creating a gap described as the “Beaverhead Canyon Gateway” by the Lewis and Clark Expedition. The Barretts Rhyolite unconformably rests on the Henneberry Gulch tuff (48.1 ± 0.4 Ma) on the opposite (northwest) side of the freeway (fig. 16A). Looking upriver to the southwest, rhyolite is entrained in several large landslide complexes shed into Beaverhead Canyon.

Rhyolite flows are pink, purple, and gray and typically form resistant ridges, blocky outcrops, and extensive talus slopes. Flow-banding (fig. 16B), autobrecciation, and vitrophere are common; a prominent interval of vitrophere up to 10 m (33 ft) thick occurs near the unit’s base locally. Rhyolite occurrences commonly contain phenocrysts of smoky quartz eyes (<5 percent; 1–3 mm), euhedral plagioclase (<20 percent; 1–2 mm), euhedral biotite (<5 percent; <1 mm), and hornblende (<5 percent; 1 mm) in a waxy groundmass of cryptocrystalline quartz. Flows are intercalated with subordinate tuff intervals exhibiting a similar mineral assemblage. Hydrothermal alteration is common, likely formed by exsolved magmatic gases. Stacked rhyolite flows have a stratal thickness exceeding 150 m (500 ft) locally. U-Pb zircon ages span 47.9 ± 0.3 Ma to 47.1 ± 0.3 Ma.

The rhyolite flows and tuffs are interpreted to have erupted as a lava dome complex during the waning stages of local Eocene volcanism in the Beaverhead Canyon area. To the northwest, rhyolite flows transition to a stack of rhyolitic pyroclastic deposits mapped as the Burns Mountain Tuff (fig. 16C; Mosolf, 2024). The pyroclastic sequence contains smoky quartz eyes resembling those in the Barretts Rhyolite and yielded U-Pb ages spanning 47.0 ± 0.4 Ma to 46.6 ± 0.3 Ma. Thus, geochronology and mineralogy data suggest the Barretts Rhyolite and Burns Mountain tuff were genetically related and possibly erupted from a shared vent or parental melt.

For a closer look at the rhyolite, cross the road bridge on foot and immediately turn left, following the climber’s trail through the willows to the outcrop.

If continuing to optional STOP 6, return to the vehicles and exit Barretts Park Campground, turning right on the frontage road. Merge onto I-15 north and drive 10 km (6.2 mi) towards Dillon. Take exit 62 for the I-15



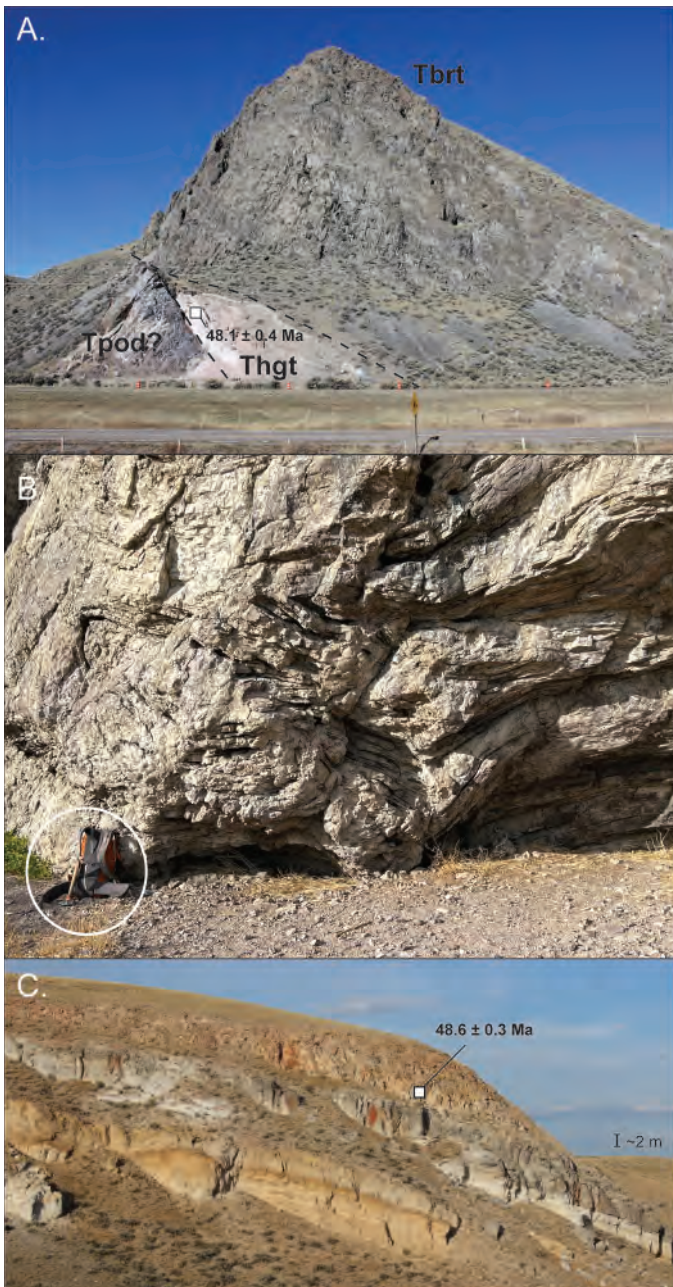


Figure 16. (A) View northwest at Stop 5 of the Barretts Rhyolite unit (Tbrt) resting on white tuff beds of the Henneberry Gulch Tuff unit (Thgt). The volcanic units are tilted northeast, towards the trace of the Blacktail fault that crosses Beaverhead Canyon Gateway. (B) Rhyolite flow at the base of Barretts Rock Park with intensely developed flow folding. (C) Gently tilted sequence of rhyolitic quartz eye tuff and ignimbrite composing the Burns Mountain Tuff unit 4 km (2.5 mi) northwest of Barretts Rock (Mosolf, 2024).

Business Loop and follow Atlantic Street into Dillon 1.8 km (1.1 mi). Turn left on Reeder Street and head northwest 0.8 km (0.5 mi) to Park Street. Turn left, cross the railroad tracks, and continue on 10 Mile Road for 1.8 km (1.1 mi) to STOP 6.

OPTIONAL STOP 6—Dillon West Rhyolite 15.7 mi (45.2151°N, -112.6668°W)

The low hills northwest of Dillon are underlain by a rhyolite field exceeding 21.6 km² (8.4 mi²), mapped as the Dillon West Rhyolite (McDonald and Mosolf, 2023). The rhyolite rests unconformably on gently folded Paleozoic–Mesozoic strata, including the Madison and Snowcrest Range Group carbonates and the Quadrant, Phosphoria, and Dinwoody Formations (fig. 17). Bozeman Group sediments, including the Renova Formation, unconformably rest on the dome complex and are intercalated with lava flows to the north. While compositionally similar to the Barretts Rhyolite visited at STOP 5, the Dillon West Rhyolite is notably older and age-correlative to the Pipe Organ Dacite unit. An exceptional two-dimensional section of the Dillon West rhyolite can be seen in the cliff face above STOP 6 (fig. 18).

Rhyolite flows are typically vuggy, massive, and variegated, occurring in shades of red, pink, gray, and brown. Sparse (<5 percent) phenocrysts include biotite, hornblende, feldspar, and quartz. Rheomorphic deformation is common, including intense flow banding and folding locally (fig. 18). Discontinuous intervals of vitrophere, autobreccia, and tuff with varying degrees of welding occur locally. U-Pb zircon ages obtained for this unit span 49.9 ± 0.2 Ma to 49.2 ± 0.2 Ma. A sample collected at STOP 6 yielded an age of 49.6 ± 0.3 Ma. The Dillon West Rhyolite is interpreted to be a lava dome complex, contemporaneous with the Pipe Organ Dacite that erupted from a separate vent approximately 17 km (10.5 mi) to the southwest. It's unclear if these eruptive centers were comagmatic, sharing a magma chamber or parental melt.

END OF TRIP



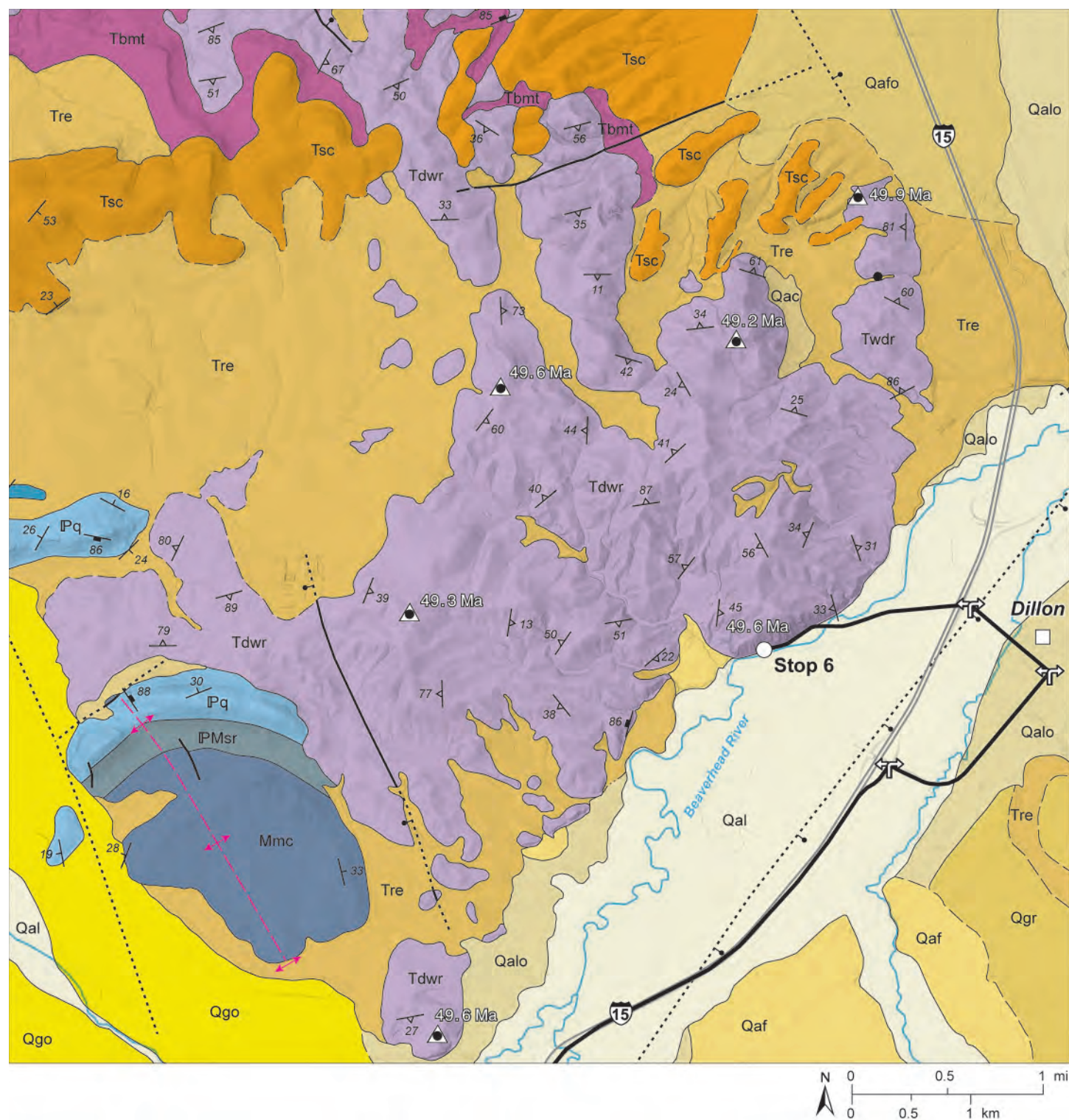


Figure 17. Simplified geology map for STOP 6. Modified from Mosolf and McDonald (2022).



Figure 18. (A) Cliffy outcrops of the Dillon West Rhyolite unit at STOP 6. (B) Flow banding and folding in rhyolite lava. Annotation outlines folded flow banding.

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THE INDIAN QUEEN MINE, UTOPIA MINING DISTRICT, PIONEER MOUNTAINS, SOUTH-WEST MONTANA

Kyle Eastman and Katie McDonald

Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

INTRODUCTION

The Indian Queen Mine is within the Utopia Mining District, located north of Birch Creek in the eastern range of the Pioneer Mountains (fig. 1). The district includes several small base metal (Cu, Au, Ag) mines developed along the contact between the Pioneer Batholith and Paleozoic carbonate rocks. This field trip will focus on the history and economic geology of the Indian Queen Mine—the largest mine in the district. The trip will include an approximately 2-mi round trip hike exploring the historical mine workings that involves some short, steep climbs over uneven terrain.

ROAD LOG—BANNACK TO BIRCH CREEK (approximately 42 mi one way)

Exit Bannack State Park, head west 1 mi, and turn right on Bannack Bench Rd. Continue north 2.9 mi to MT 278; turn right (east) and proceed 17.3 mi to I-15. Turn left onto I-15 and travel north 15 mi to the Apex/Birch Creek exit (exit 74). Turn left at bottom of the exit onto Birch Creek Road and head west approximately 2.4 mi to Stop 1. Parking is along right (north) side of Birch Creek Road.

STOP 1

GEOLOGIC OVERVIEW OF THE UTOPIA MINING DISTRICT

(45.3784°N, -112.7570°W)

The Utopia Mining District is located in the southeastern Pioneer Mountains within the Birch Creek drainage (fig. 1). The western part of the district is underlain by folded Mesoproterozoic (?) to Cretaceous sedimentary rock intruded by the Late Cretaceous Pioneer Batholith (fig. 2). The eastern third is underlain by Tertiary and Quaternary surficial deposits. The district is located in the footwall of the Grasshopper/Kelley thrust (fig. 1), which forms the leading edge of the Grasshopper thrust plate, one of the major Late Cretaceous thrust sheets of the southwest Montana fold-thrust belt (Ruppel and others, 1981; Ruppel and Lopez, 1984). The most detailed mapping of the area is by Myers (1952), Sharp (1970), Tysdal and others (1994), and McDonald and Yakovlev (2019).

The oldest rocks in the district are Mesoproterozoic (?) to Cambrian quartzites and argillites (unit €Yqa) exposed in the core of the Humbolt Mountain anticline in the southwestern part of the district (figs. 1, 2). These rocks were previously mapped as part of the Missoula Group

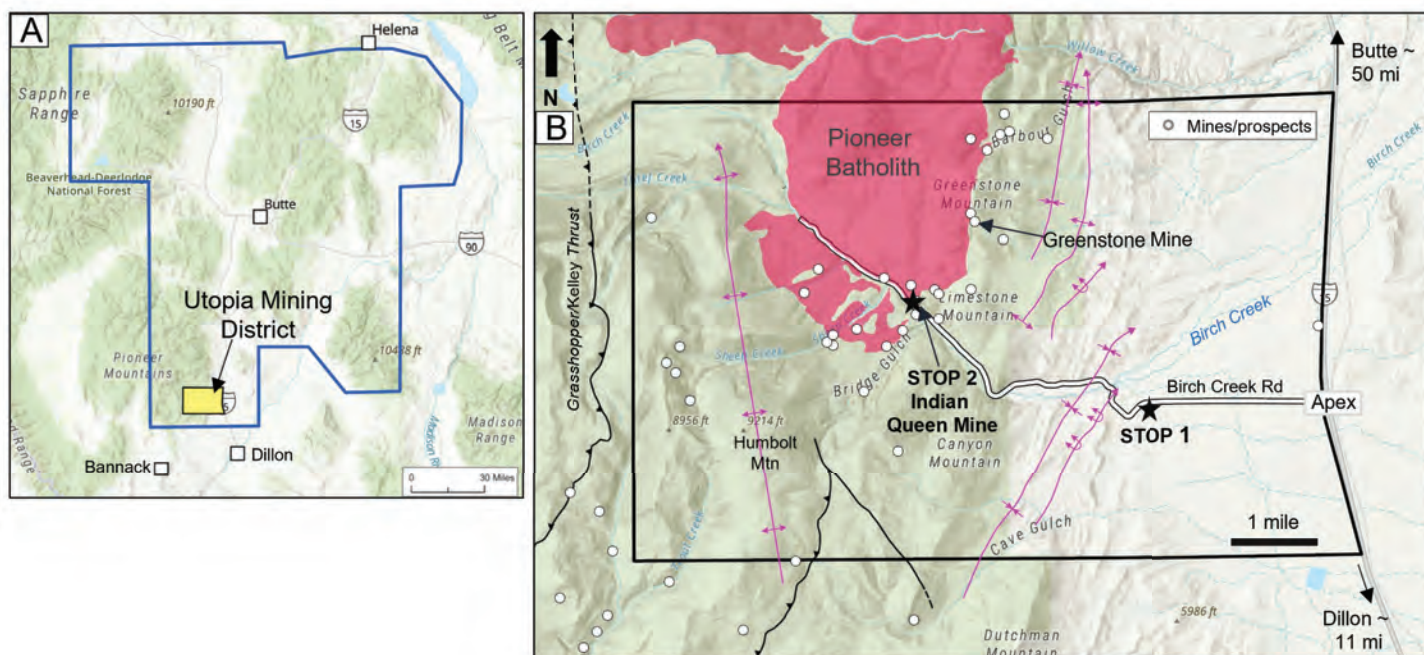
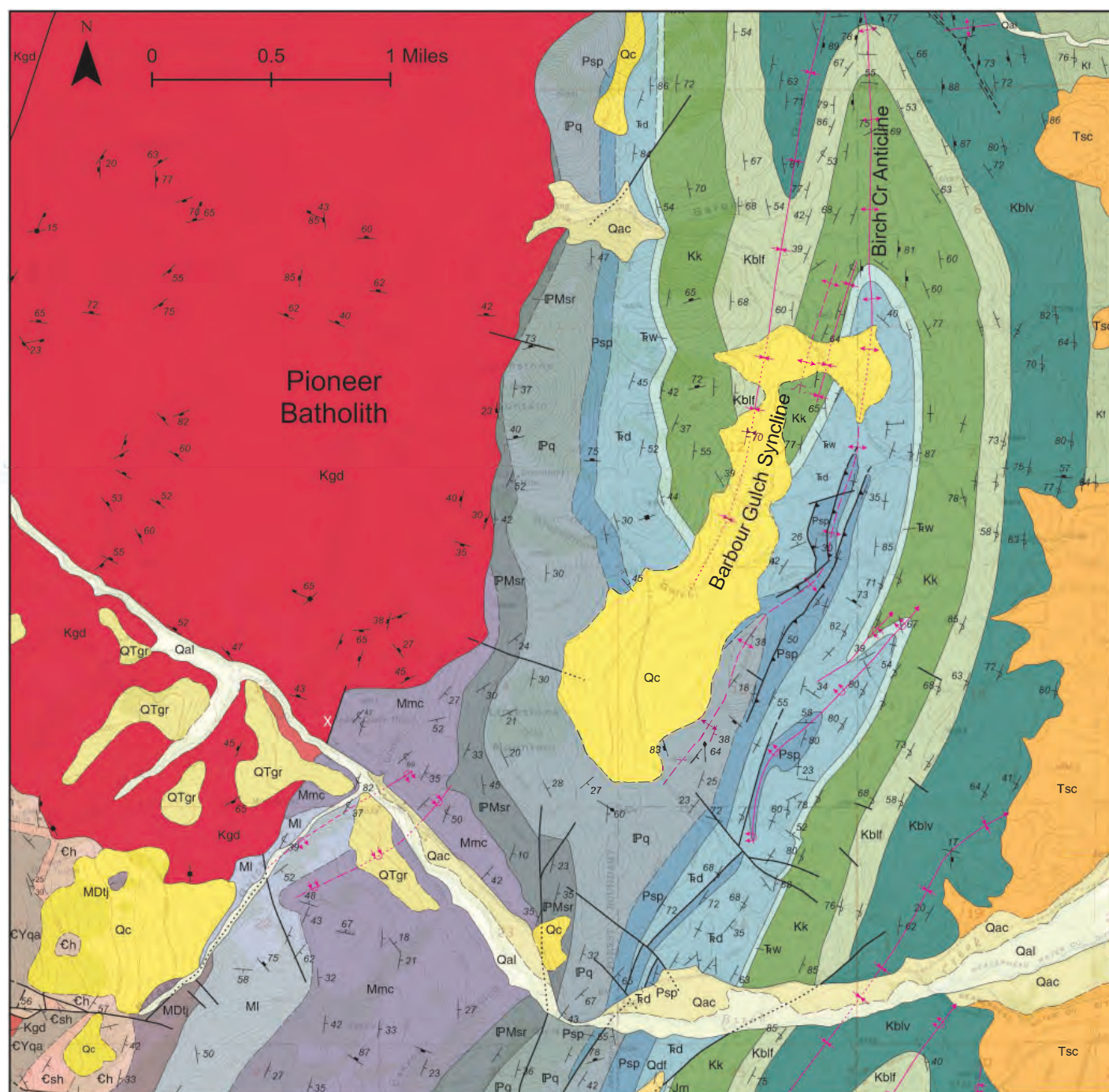


Figure 1. Location maps showing (A) the boundary of airborne radiometric data acquired by the USGS Earth MRI program (blue line); and (B) boundary of Utopia Mining District, location of mines within the district, major faults and folds, and the field trip stops.





Quaternary–Tertiary		Cretaceous–Triassic		Permian–Mississippian		Cambrian–Mesoproterozoic	
Qal	Alluvium	Kf	Frontier Fm	Psp	Shedhorn–Phosphoria Fms	Ch	Hasmark Fm
Qc	Colluvium	Kblv	Vaughn Mbr, Blackleaf Fm	IPq	Quadrant Fm	Csh	Silver Hill Fm
Qac	Alluvium-colluvium	Kblf	Flood Mbr, Blackleaf Fm	PMsr	Snowcrest Range Gp	CYqa	Quartzite–argillite
Qdf	Debris flow deposit	Kk	Kootenai Fm	Mmc	Mission Canyon Fm	Cretaceous Intrusions	
QTgr	Gravel deposit	Tw	Woodside Fm	MI	Lodgepole Fm	Kgd	
Tsc	Sixmile Cr Fm	Jm	Morrison Fm	MDtj	Three Forks–Jefferson Fms		
		Trd	Dinwoody Fm				

Figure 2. Geologic map of the area near the Indian Queen mine (modified from McDonald and Yakovlev, 2019).



of the Mesoproterozoic Belt Supergroup (Myers, 1952; Ruppel and others, 1993) but the upper part of the section, characterized by thin beds of dark green to black argillite interbedded with thicker beds of medium- to coarse-grained quartzite, is interpreted as Cambrian based on the discovery of classic Cambrian trace fossils in float at the top of Humbolt Mountain (McDonald and Yakovlev, 2019). The quartzites and argillites are overlain by shale and dolomite of the Cambrian Silver Hill and Hasmark Formations. The age of the enigmatic quartzites and argillites in the Humbolt Mountain anticline has been the subject of considerable debate during previous TRGS field trips (see Sears and Link, 2007; Sears and others, 2010, 2023).

Devonian through Cretaceous rocks are well exposed along the north side of Birch Creek Road in a series of northeast-trending, upright to overturned folds. The Devonian Jefferson through Mississippian Madison Group (Lodgepole and Mission Canyon Formations) are predominately marine carbonates that unconformably overlie Cambrian rocks. They are overlain by Late Mississippian to Early Triassic strata (Snowcrest Range Group, Quadrant and Phosphoria formations) that record the transition from dominantly carbonate deposition to marginal marine and continental siliciclastic foreland basin deposition. The siliciclastic rocks include the Triassic Woodside Formation, the Late Jurassic Morrison Formation, and the Early Cretaceous Kootenai Formation. The Early to Late Cretaceous Blackleaf and Frontier Formations record the return of marine deposition to the area.

The Late Cretaceous Pioneer Batholith intrudes the folded sedimentary rocks. The intrusive rocks are granodiorite to granite with reported ages of ~72 to ~70 Ma (McDonald and Yakovlev, 2019; Zen, 1988). Contact metamorphism along the steep, generally concordant eastern and southern contact created numerous garnet–epidote skarns that produced copper, lead, silver, and gold (Geach, 1972). The contact aureole extends up to 2 km (1.6 mi) from the contact and is characterized by marble and hornfels.

The youngest deposits in the district are Tertiary and Quaternary surficial deposits. The Tertiary deposits, correlated with the Miocene Sixmile Creek Formation, consist primarily of conglomerate, gravel, and sand, and are interpreted as debris flow and fluvial deposits (Fritz and Sears, 1993). Quaternary glacial deposits are widespread along Birch Creek and include till and outwash (Thomas and others, 2007).

Return to vehicles, and continue west approximately 3.7 mi. On the south side of the road is a pullout with a road sign that marks the ghost town of Farlin. Pull in and park here for Stop 2.

STOP 2

(45.3971°N, -112.8405°W)

Stop 2 involves an overview of the Utopia Mining district, followed by a hike (Stops 2a–2d) around the historical workings at the Indian Queen Mine. The workings include collapsed adits and shafts, so be aware of hazards. The Indian Queen is located on a patented mining claim (private property), so permission is required for access.

The Utopia Mining District Overview

The Utopia Mining District (fig. 1) includes numerous small base metal mines and prospects, mostly developed in skarns along the contact between the Pioneer Batholith and Paleozoic limestones. Geach (1972) provides the most comprehensive summary of the district, including mine maps and production figures. From 1902 to 1965, total reported production was 1,771,824 lb copper; 5,464 lb lead; 43,744 oz silver, and 308 oz gold. The Indian Queen (fig. 3) is the primary mine in the district and accounts for nearly all production. Tungsten (as scheelite) is also present, but to date, only low-grade ore has been found, with 1 ton of tungsten ore (1.2 percent WO_3 , Pattee, 1960) produced from the Greenstone Mine north of the Indian Queen (fig. 1).

The Utopia district includes mostly lode mines, but minor placer workings were located along Birch Creek. Iron ore mined for flux and dimension stone from limestone are other reported commodities. The ghost town of Farlin, which included a store and school, was home to approximately 500 people from the late 1800s to early 1900s. It was located near the smelter built to process ore from the Indian Queen.

The Indian Queen Mine

The Indian Queen Mine opened in 1867 and was worked intermittently until 1923 (Geach, 1972). The workings, now inaccessible, consisted of three adits, two small shafts, and numerous prospect pits (fig. 4). Underground workings were reported to consist of the two shafts (150 and 200 ft deep), an 820-ft tunnel, and a total of 1,600 ft of underground openings. Total recorded production was 22,907 tons of ore, which contained 1,729,404 lbs copper, 42,219 oz silver, and 299 oz gold.

The mine is located along the contact between the Pioneer Batholith and limestone of the Mission Canyon Formation, the uppermost limestone of the Mississippian Madison Group. Geach (1972) describes the ore as primarily in a garnet–epidote skarn (fig. 5) localized in irregular shoots and pods along a fault between the batholith and limestone or in the limestone, suggesting that mineralization may have been later than skarn development. Most of the ore mined was oxidized and contained native copper, malachite, azurite, chrysocolla, and cuprite. Sulfide ore below the oxidized zone consisted of massive chalcocite,



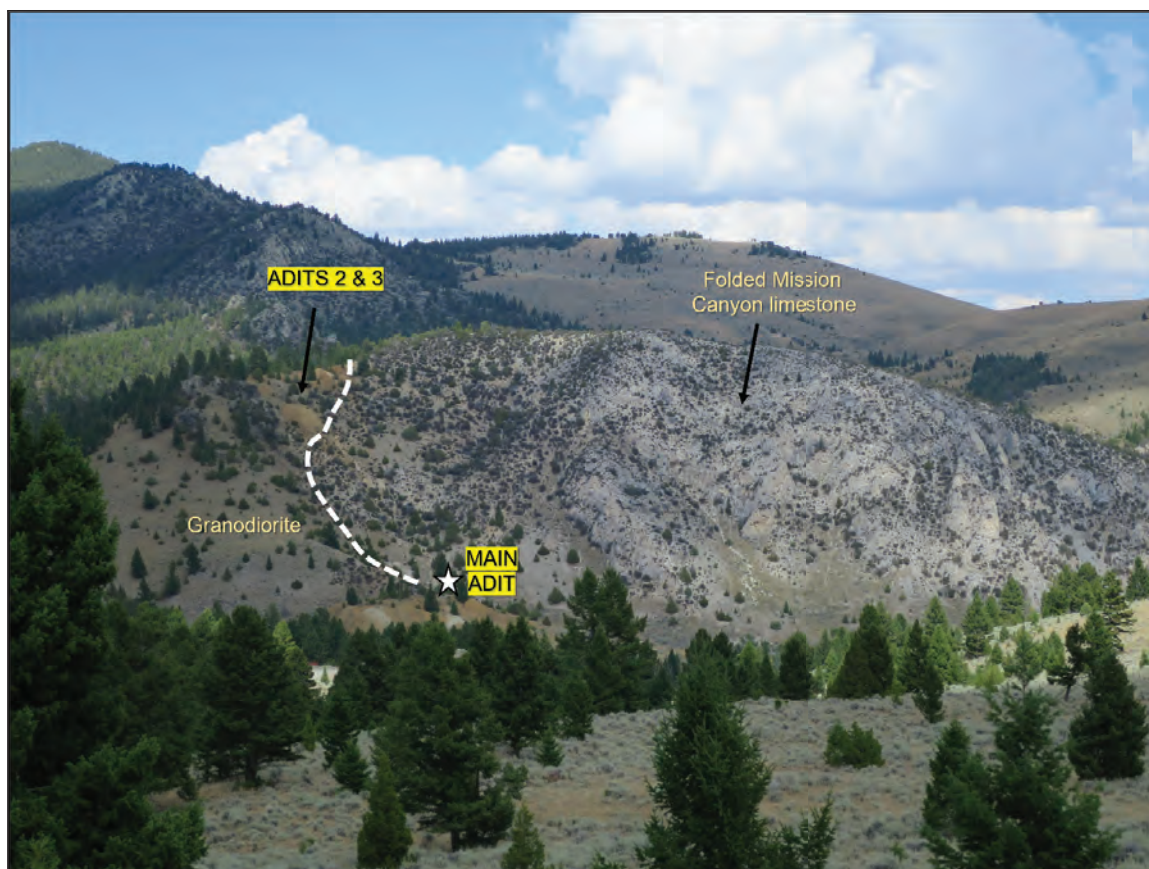


Figure 3. Photo looking north toward the Indian Queen Mine. The tightly folded Mission Canyon limestone forms cliffs to the east of the mine. The approximate contact between the Pioneer Batholith (granodiorite) and the limestone is shown by the dashed white line.

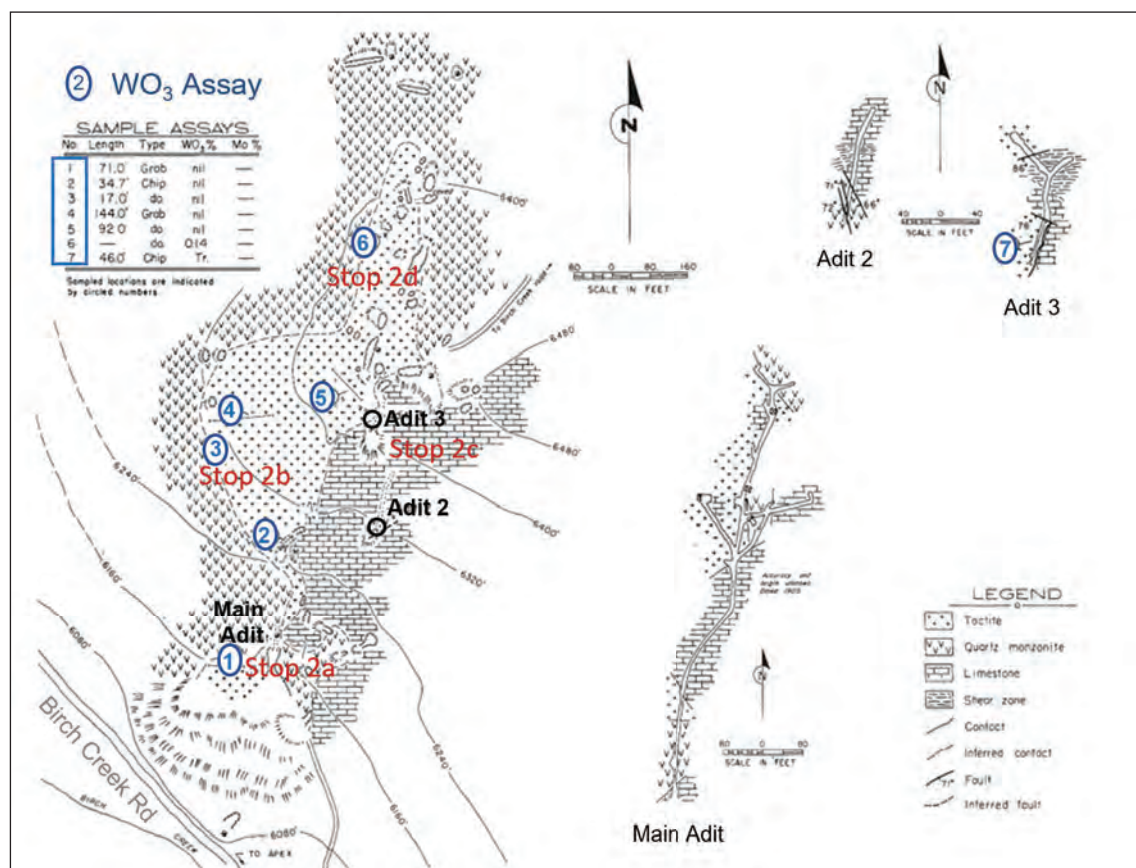


Figure 4. Map of the Indian Queen Mine showing geology and mine workings (modified from Geach, 1972). Blue circles show location of samples assayed for tungsten. Field trips stops are in red.





chalcopyrite, and bornite. The U.S. Bureau of Mines collected samples for tungsten (fig. 4), but only one sample from a small pit at the north end of the mining claim contained reportable amounts. That sample—a gray banded rock containing scheelite (CaWO_4), magnetite, and malachite—assayed 0.14 percent WO_3 (Geach, 1972; Pattee, 1960). Some scheelite contains molybdenum up to ~1,000 ppm, indicating partial substitution towards powellite

(CaMoO_4); this imparts a yellowish green short-wave UV fluorescence as compared to the usual bright blue-white fluorescence of scheelite.

The USGS Earth Mapping Resources Initiative (Earth MRI) acquired airborne magnetic and radiometric data in 2023 with coverage of the western half of the Boulder Batholith region; this survey also included the eastern Pioneer Mountains and Utopia Mining District (figs. 6, 7). This survey was flown by helicopter at 100 m altitude, with 200 m east–west line spacing, and 2,000 m north–south tie lines. The data can be accessed at <https://www.usgs.gov/special-topics/earth-mri/data>. Both radiometric and magnetic data types highlight the Pioneer Batholith, with a strong magnetic signature at the intrusive contact likely representing skarn and hornfels alteration with magnetite and/or pyrrhotite.

From the Stop 2 parking area, proceed west on foot about 0.4 mi along Birch Creek Road to the Indian Queen Mine. Due to limited parking space along the road, we will walk to the mine site, and to Stops 2a–2d described below and shown in figure 4. However, there may be an option for a few folks to drive the 0.4 mi to the mine site and park along the right (north) side of Birch Creek Road.

Stop 2a

At this stop we will discuss the mine and examine waste rock on the dump from the Main adit. The mine waste rock contains skarn minerals, Fe-oxides, and Cu minerals.

Stop 2b

Proceed uphill to Stop 2b. At this stop we will examine outcrops of the skarn exposed in several pits and prospects.

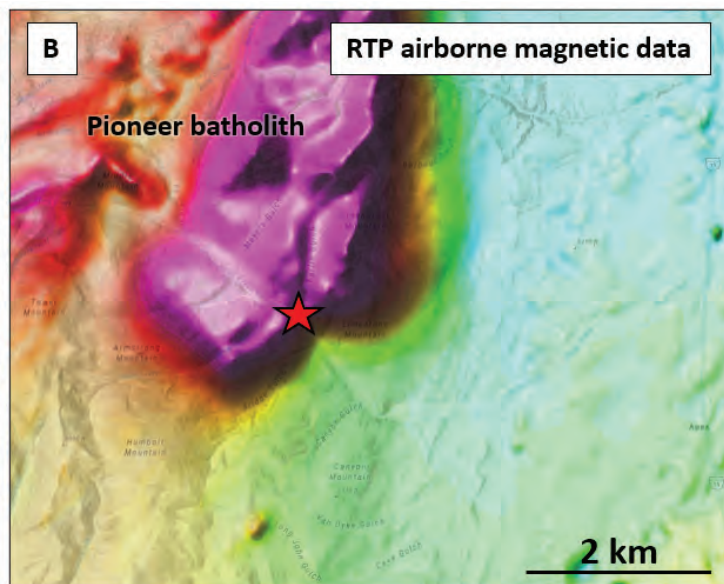
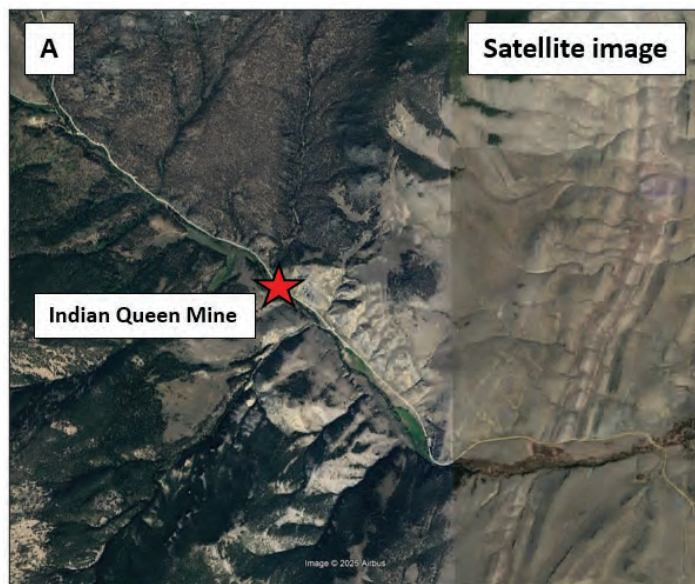


Figure 5. Hand sample of garnet–magnetite–scheelite skarn.



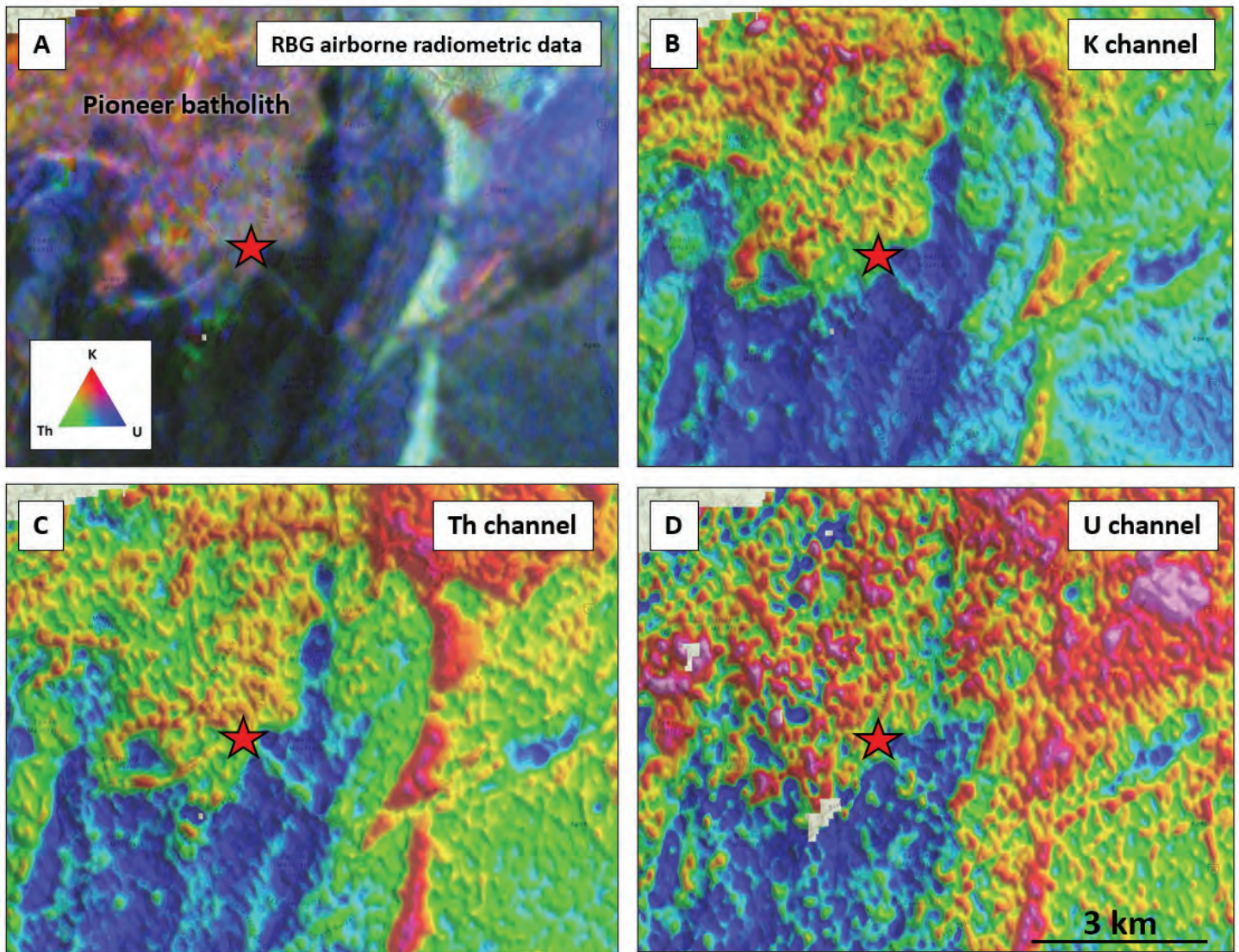


Figure 7. Airborne radiometric data acquired by the USGS Earth MRI program. (A) Three-channel red–blue–green (RGB) radiometric image. (B) K (potassium) channel radiometric image. (C) Th (thorium) channel radiometric image. (D) U (uranium) channel radiometric image. Red star is the Indian Queen Mine.

Stop 2c

Head east along the ridge toward workings near Adit 3. Here we will look at a massive garnet skarn outcrop located just west of the collapsed workings. We will also cross the contact between the marble (metamorphosed Mission Canyon Formation) and the skarn.

Stop 2d

This stop is located in the trees approximately 100 yards north of Stop 2c. Here we will examine garnet–magnetite–scheelite skarn (fig. 5) exposed in prospect pits.

Return to vehicles. There is a two-track road just north of Stop 2d that can be followed as an easier route on the return hike. The road follows Farlin Gulch and connects back to the Birch Creek Road.

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FAULT, FOLD, AND CLEAVAGE DUPLEXES OF SANDY HOLLOW

Stuart D. Parker,¹ and Steven E. Boyer²

¹Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

²Olympia, Washington

INTRODUCTION

Characterizing the geometry of a structure within deformed rocks is the first step of structural analysis. This crucial first step often occurs in the field, and requires not only good exposures, but also a robust conceptual framework to pull from. Using the incredible exposures of Sandy Hollow, southwestern Montana, we hope to help participants develop and refine their conceptual understanding of fault, fold, and cleavage duplexes, which are ubiquitous in fold-thrust belts.

A duplex is a zone of deformation that occurs between a lower (floor) and upper (roof) thrust (Dahlstrom, 1970). These structures have long been described in the field (e.g., Willis, 1902), decades before the term duplex was applied. While the duplex concept was originally applied only to faults, the concept has recently been expanded to also include fold and cleavage duplexes (Boyer and Mitra, 2019). This expanded conceptual framework not only provides more precise and efficient terminology for field descriptions within fold-thrust belts, it also accentuates the commonalities between duplexes, whether they be composed of faults, folds, cleavage, or any combination of the three. Highlighting the geometric and kinematic similarities between different styles of duplexes allows structural geologists to more easily interpret the wide variety of geometries that may be produced by shortening a sequence of rocks with variable material properties, without having to evoke changes in the stress state.

Fundamentally, duplexes act as transfer zones, transferring displacement up-section and forward through multiple detachment horizons (e.g., Mitra and Boyer, 1986). Exposures of duplexes provide a snapshot of the evolving strain path, as slip was iteratively transferred between detachment horizons. In this field trip, we not only highlight the geometric complexity of duplexes but also hope to demonstrate how they evolve through time. Viewing fault, fold, and cleavage duplexes as evolving transfer zones leads to more simple and precise interpretations of the most complicated parts of the Sandy Hollow uplift. We hope that the concepts developed during this field trip may be applied to other localities, anywhere that duplexes may be found.

GEOLOGIC BACKGROUND

Exposures of folded and faulted Mississippian to Late Cretaceous rocks between Block and McCartney mountains will herein be referred to as the Sandy Hollow uplift (fig. 1). The Sandy Hollow uplift is a small structural culmination within the thin-skinned McCartney Mountain salient. The McCartney Mountain salient is a small segment of the Cretaceous to Eocene Cordilleran (aka Sevier) fold-thrust belt, situated within the regional-scale Montana Recess (Brumbaugh, 1973; Brumbaugh and Hendrix, 1981; Brandon, 1984). Local shortening directions ($\sim 110^\circ$) are orthogonal to the trace of folds and faults (Brumbaugh and Hendrix, 1981), with vertical-axis rotation related to progressive development of the curved geometry that defines the salient (Eldredge and Van der Voo, 1988). These observations are similar to findings in the Wyoming salient, where about 75% of the present-day curvature is related to vertical-axis rotation, while the remaining 25% reflects the primary curvature of the basal detachment, which followed weak horizons in the host sedimentary rocks (Weil and others, 2010).

The McCartney Mountain salient can be viewed as the leading tip of an upper thin-skinned fold-thrust belt that overlies a kinematically linked but deeper thick-skinned system (Kulik and Schmidt, 1988; Parker and Pearson, 2023). Near the front of the salient, deformation dies out as a series of low-amplitude, fault-propagation folds (Parker and others, 2025). The Sandy Hollow uplift can therefore be viewed as a small, internally deformed culmination within a larger, fold-dominated salient.

The Sandy Hollow uplift deforms and exposes Mississippian to Cretaceous rocks, ranging from weak, argillaceous carbonates to strong sandstones and chert (fig. 2). As we'll see throughout this field trip, the deformed section has a pronounced mechanical stratigraphy, with the thicker upper package corresponding with the foreland basin deposits of the Kootenai and Blackleaf Formations (Kkl-Kblf in fig. 2).

The local timing of deformation is constrained by folded and faulted ca. 75 Ma sills that emanate from the McCartney Mountain pluton (Kalakay and others, 2001;



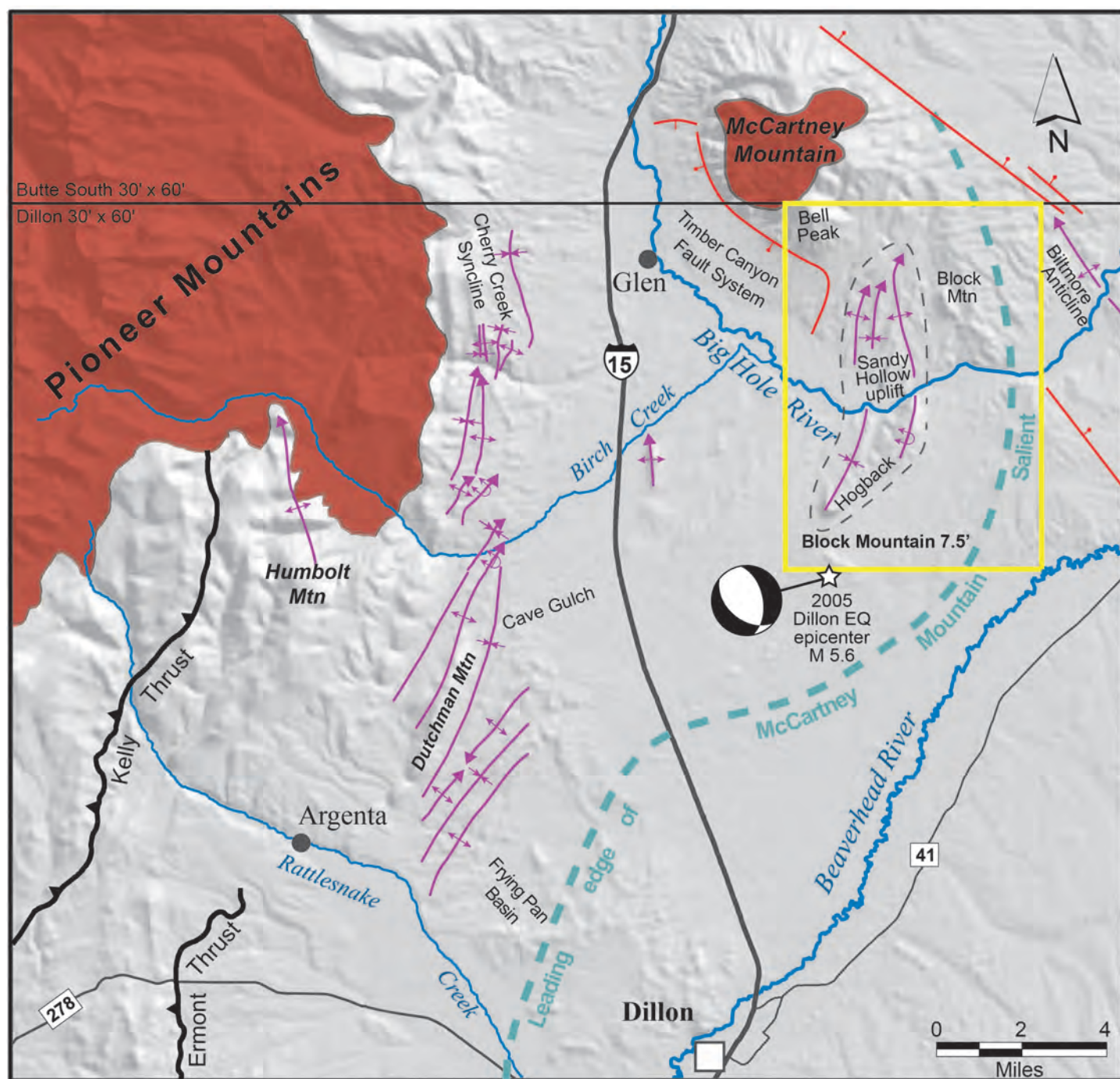


Figure 1. Simplified tectonic map of the field trip area, showing the extent of the Sandy Hollow uplift. Prominent folds shown in magenta. Quaternary normal faults shown as red lines. Plutonic rocks shown in red. Extent of the Block Mountain 7.5' (Parker and others, 2025) shown in yellow.

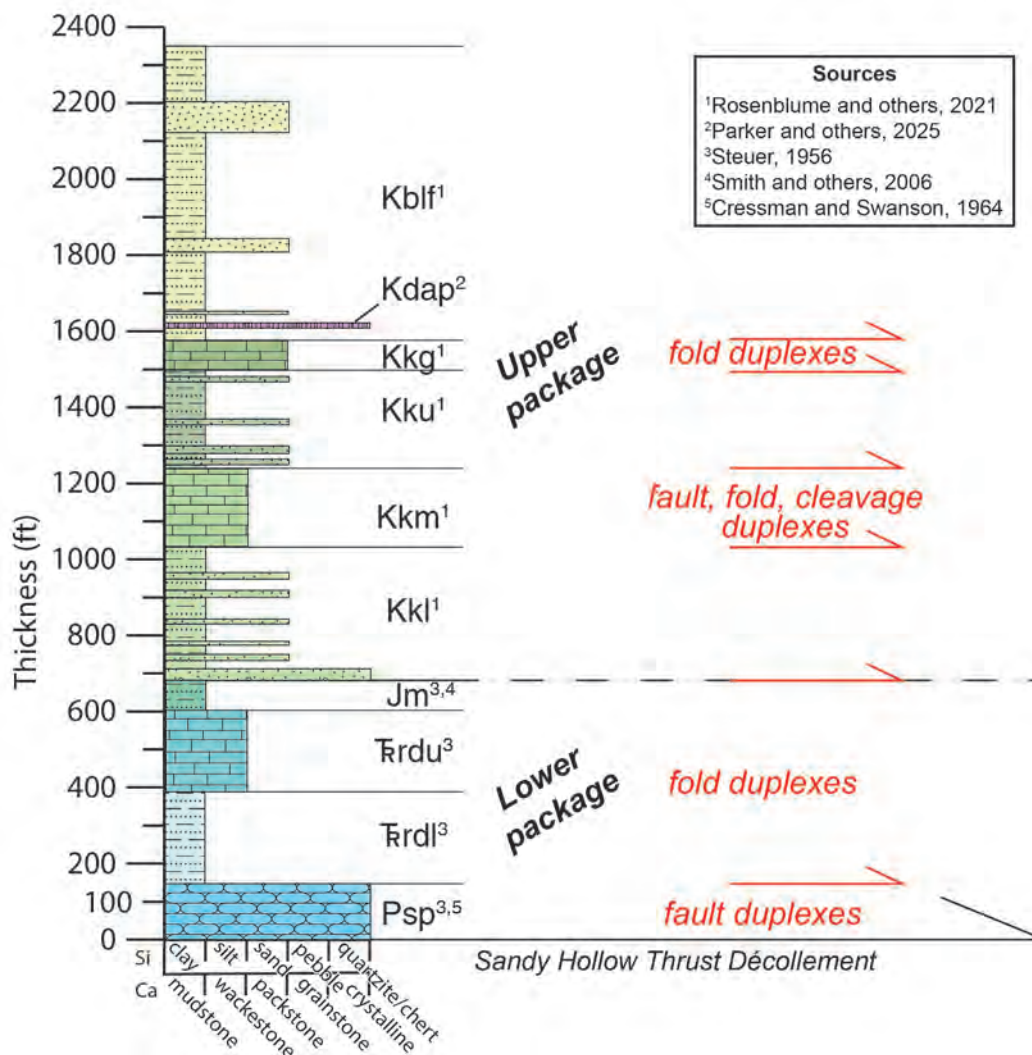


Figure 2. Simplified stratigraphic column of hanging wall rocks of the Sandy Hollow Thrust. See figure 3 for full map unit names. Map-scale structural packages shown in black. Minor detachment horizons and outcrop-scale mechanical stratigraphy is shown and described in red. All measured sections are from or near Ziegler Gulch, with sources listed.

Foster and others, 2012) and undeformed ca. 48 Ma and younger volcanic rocks (Parker and others, 2025). The McCartney Mountain pluton was emplaced about 1–3 km deep (Friberg and Vitaliano, 1981), suggesting that the deformation observed in the Sandy Hollow uplift corresponds to very shallow crustal levels.

This field trip focuses on exposures along the Sandy Hollow thrust, the main structure in the Sandy Hollow uplift north of Burma Road (fig. 3). Here, the uplift consists of two overturned, ESE-verging fold pairs, with wavelengths of about 1.5–2 km (~1 mi). The Sandy Hollow thrust cuts across and is folded by the larger amplitude fold pair. The folded footwall is termed the Sandy Hollow anticline. The folded hanging wall is termed the Ziegler Gulch anticline. Slip along the Sandy Hollow thrust has stacked these two anticlines on top of each other. These stacked fold hinges coincide with a folded hanging-wall ramp, resulting in a pronounced jog in the trace of the thrust (Brumbaugh and Dresser, 1976; Parker and others, 2025). As we will devel-

op below, the Sandy Hollow anticline is the surface expression of a fault-propagation fold, above the northern tip of the Hogback thrust. The Ziegler Gulch anticline represents the early fault-propagation fold above the growing Sandy Hollow thrust, which was translated over 1 km (3/4 mi) after the thrust propagated completely through the fold. The cross-section (fig. 4) shows the structural position of the three main field trip stops, which focus on the intermediate and small-scale structures that define the upper and lower structural packages along the folded Sandy Hollow thrust.

STRUCTURAL INTERPRETATION OF THE SANDY HOLLOW UPLIFT

While it is difficult to see the entire Sandy Hollow uplift from the ground, aerial photos capture the overall geometry of the structure nicely (fig. 5). The Sandy Hollow uplift may be modeled as a series of fault-propagation folds (Suppe and Medwedeff, 1990). In such a fold, as a thrust fault propagates, it loses displacement into a fold



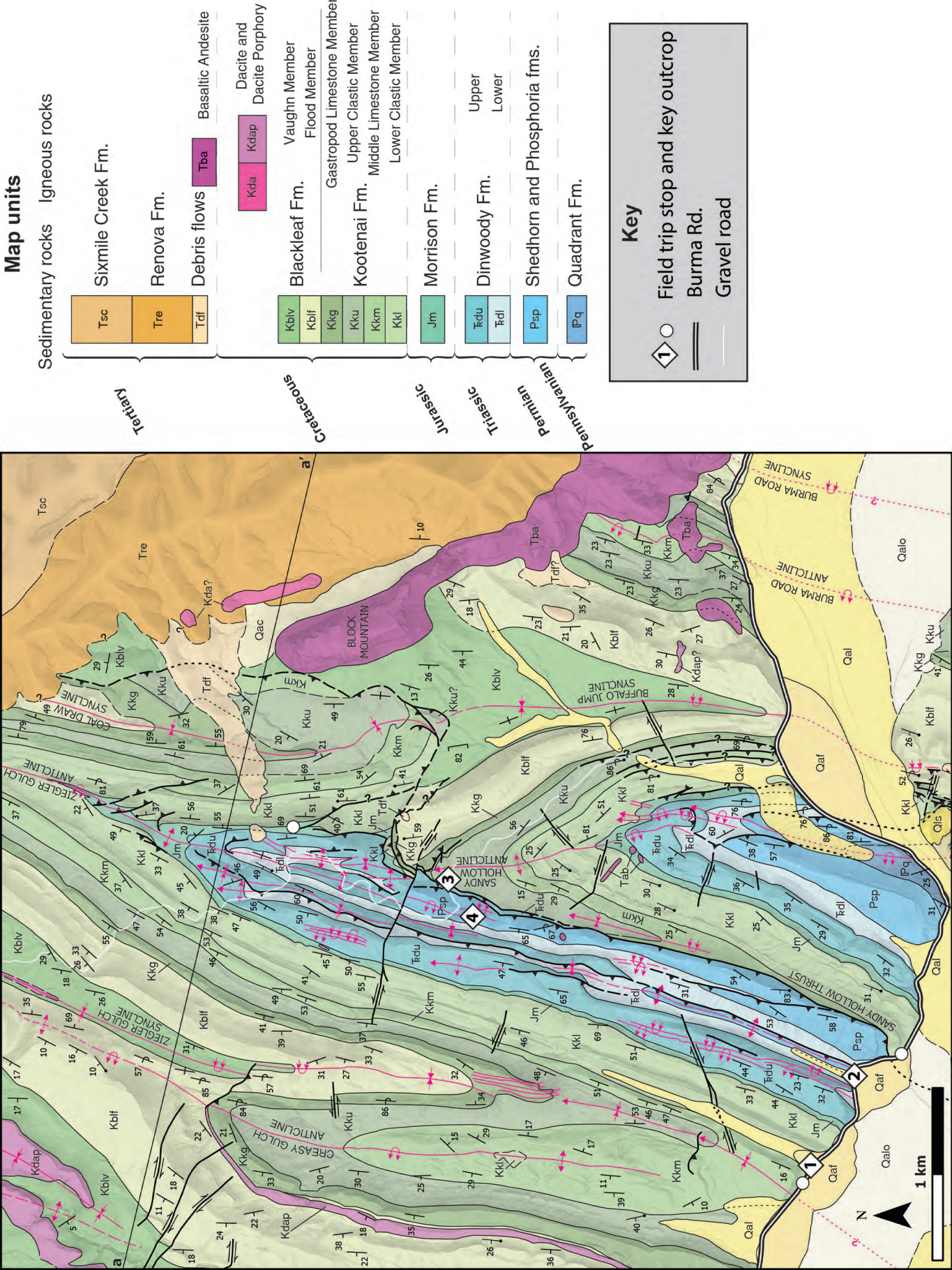


Figure 3. Simplified geologic map of the Sandy Hollow uplift, adapted from Parker and others (2025). See Parker and others (2025) for description of Quaternary surficial deposits.



Figure 4: Balanced and restored cross-section for the Sandy Hollow uplift (Parker and others, 2025). Circles show the approximate structural position of the three main outcrops of the field trip stops, shown in figure 3.

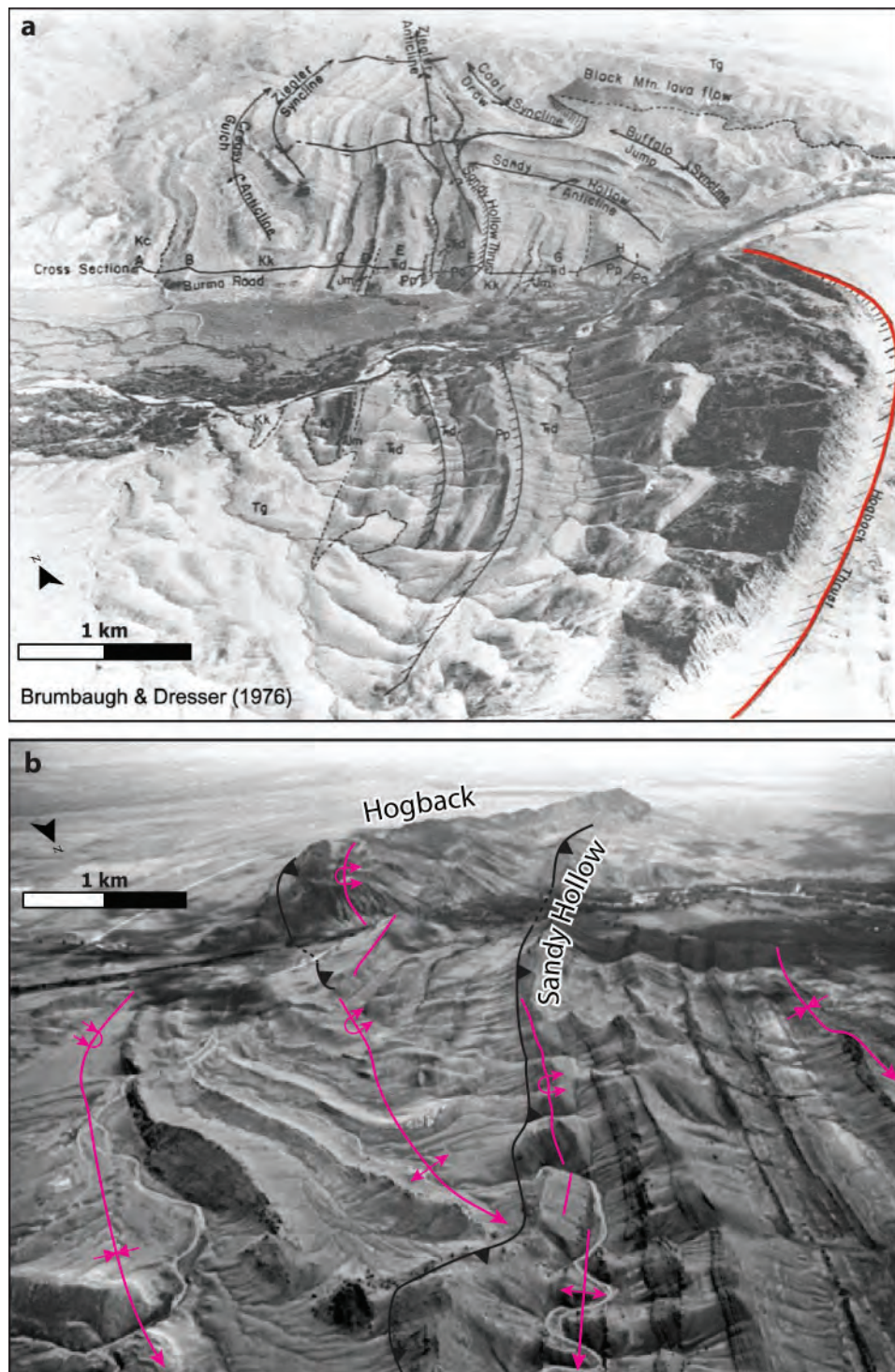


Figure 5. Two aerial views of the Sandy Hollow uplift. (a) Looking north, from Brumbaugh & Dresser (1976). (b) Looking south, taken by Boyer ca. 1979. Major thrusts labeled.

that envelops the fault tip (fig. 6). With continued thrust propagation and tectonic shortening, the fold amplitude grows and the fold tightens. Note the four fold axial planes, labeled A, A', B, and B' on figure 6. A' emanates from the fault tip and moves with the fault tip as it continues to propagate. B intersects the fault at the base of a ramp in the thrust and stays fixed throughout deformation. Between axial planes A and B' is a region of no deformation, forming a broad, low amplitude, flat-topped anticline as in the top figure. As deformation continues, A and B' move toward

one another and the anticline tightens and amplitude increases. Below the convergence point of A and B', the fold is a chevron (in yellow).

Laying out time sequences, like in figure 6, in map view defines the three-dimensional structure of a fault-propagation fold. The block model and structure contour map (fig. 7) show an example of this geometry, demonstrating how the surface expression of a fault-propagation fold may be mapped. Note that the flat-topped anticline resembles a kink fold (Faill, 1973), with a fold axis that is defined by the line of intersection of kink planes (C in fig. 7a). These kink planes are the boundaries of kink bands, and may be mapped at the surface as two separate fold axes (fig. 7b; Faill, 1973). Faill (1973) notes that the three-dimensional geometry of kink folds could only be explained if the intersection line of kink planes (C in fig. 7a) did not lie along the same bedding plane. The models of Suppe and Medwedeff (1990) and Wilkerson and others (1991) demonstrate how this can be. In the early fold, on the left of figure 7a, the flat-topped anticline is bounded by two kink planes (A, B'). These kink planes converge downward, intersecting at the stratigraphic layer that coincides with the fault tip (black dot in fig. 7a). As the fault-propagation fold continues to grow through time, or is viewed along strike (fig. 7a), the fault tip migrates to higher stratigraphic levels. As a result, the line of intersection of the two kink planes (C), which defines the fold axis, does not lie in bedding planes (Faill, 1973). In other words, the line of intersection (C) that separates the tight, lower chevron part from the more open, upper flat-topped

part of the anticline plunges down through the bedding, towards the active fault tip along strike and at depth. The plunging three-dimensional geometry defined by the active fault tip and the coincident intersection of these opposing kink planes (C in fig. 7) demonstrates how progressive fault-propagation folding can produce complex geometries that are inclined relative to bedding. This concept may be useful not only while mapping structures in the Sandy

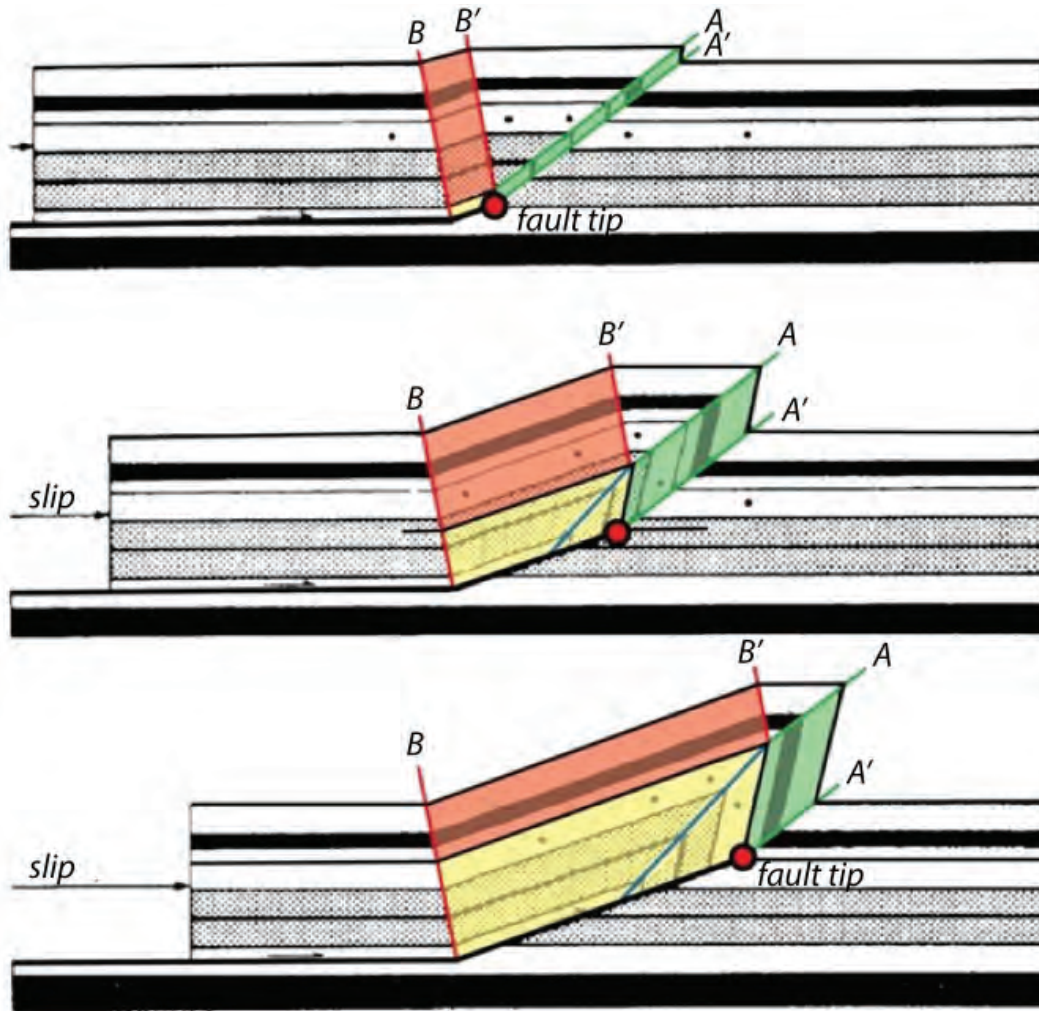
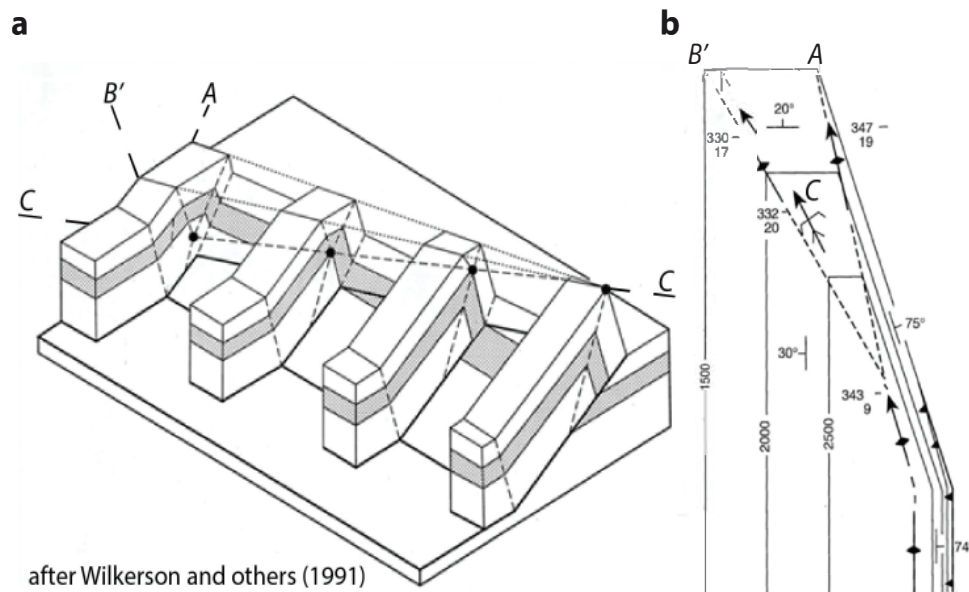


Figure 6. Fault-propagation fold model of Suppe and Medwedeff (1990). Red circle shows active fault tip. Red and green define the limbs of the upper fold. Yellow defines the extent of the lower chevron fold. Kink planes, also referred to as axial planes in the text, are shown and labeled (A, A', B, B').



after Wilkerson and others (1991)

Figure 7. Three-dimensional structure of a fault-propagation fold, showing the differing geometries of the fold that develop with progressive deformation. Kink planes (A, B') and the line of intersection of the two kink planes (C), labeled. (a) Block model showing the geometric relationship between the propagating fault and the overlying fold. See text for discussion. (b) Structure contour map showing distribution of folds and faults at the surface. Labeled structure contour lines show the elevations of a reference bed.



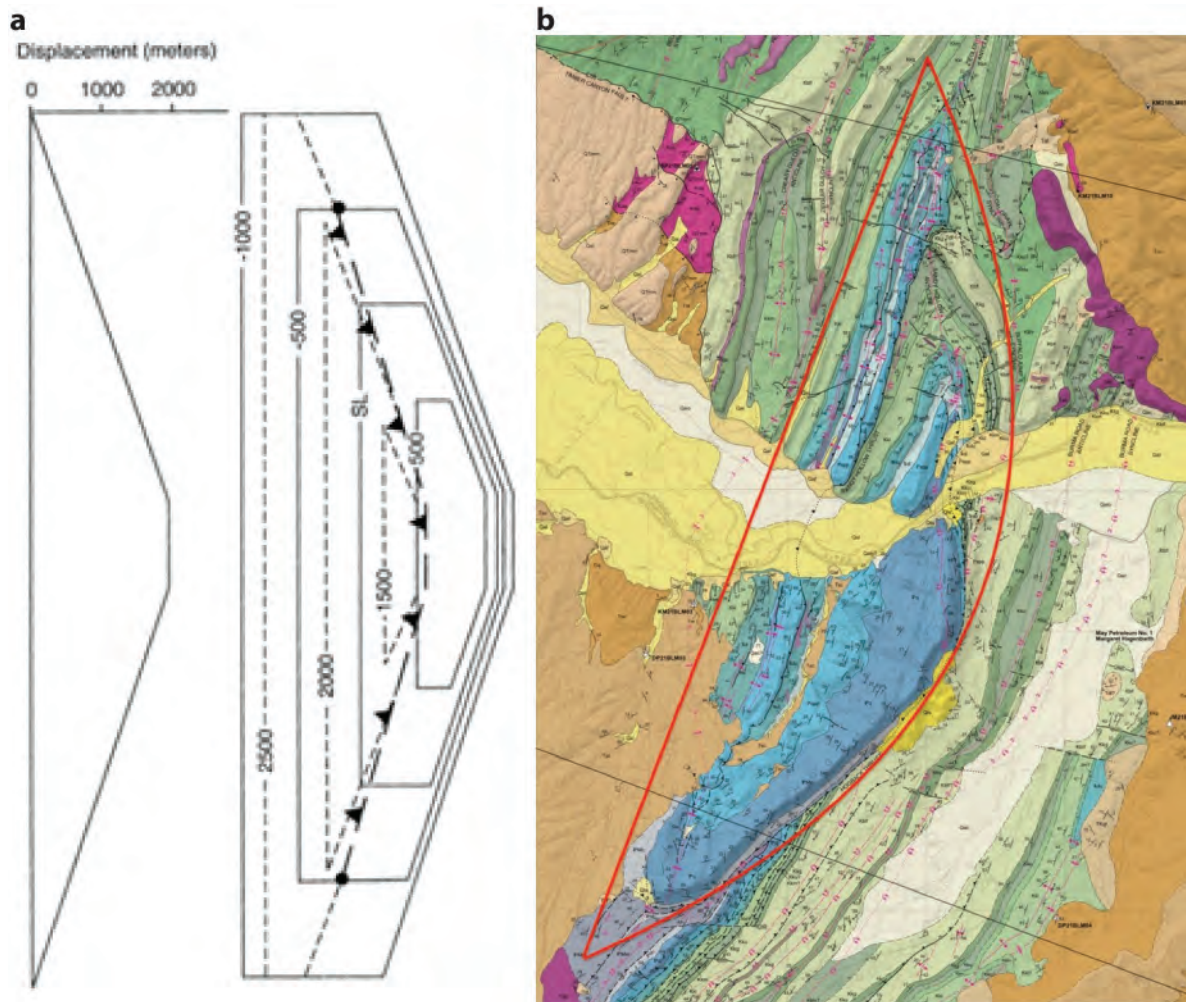


Figure 8. Fold geometry of the Sandy Hollow anticline, which we interpret as a fault-propagation fold formed above the Hogback thrust. (a) Comparison of displacement magnitude as plotted against strike and structure contour map of a deep layer deformed by the Hogback thrust. (b) General extent of Sandy Hollow anticline, outlined in red. Geologic map from Parker and others (2025).

Hollow uplift, but also while interpreting the geometry and kinematics of the smaller fault-propagation folds we will see throughout this field trip.

The Sandy Hollow uplift shows similar geometries, but on a larger scale. For instance, let's look at displacement along the Hogback thrust, the fault that formed the overlying Sandy Hollow anticline. A structure contour map of a deep layer deformed by the Hogback thrust shows that displacement along strike (fig. 8a) mimics the arcuate map pattern of the fault trace (fig. 8b), both of which resemble the letter “D.” This arcuate shape develops as the fault-propagation fold grows laterally during progressive shortening, resulting in a maximum amount of displacement in the center of the fault trace. Additionally, the contours show that the deep layer is folded into a tight chevron. Like in the block model (fig. 7a), mapped fold hinges plunge toward the fault tips, oblique to bedding. These similarities suggest that the Sandy Hollow uplift is composed of fault-propagation folds.

BRIEF TUTORIAL ON DUPLEX ZONES

While the overall structure of the Sandy Hollow uplift matches that of a series of fault-propagation folds, the first-order fold hinges and fault bends are composed of countless small-scale duplexes of all types. An understanding of duplexes is therefore prerequisite for understanding the outcrop-scale structures of the Sandy Hollow uplift.

The simple “flat-ramp-flat” geometry of thrust faults, often referred to as the “Rich model” (Rich, 1934), is the basic building block of many compressional structures, including duplexes. This flat-ramp-flat geometry (fig. 9) arises as thrust faults follow the path of least resistance toward the surface, following weaker layers along long flats, and ramping across stronger layers. As a consequence, “fault-bend folds” form in the translated hanging wall (Suppe, 1983). As shown in figure 9, a thrust may move from a lower detachment level to a higher one, transferring displacement upward through a single step or ramp.

In a duplex, displacement is transferred from a lower to an upper detachment, not through a single step, but through a zone of imbricates called a “duplex fault zone.” The

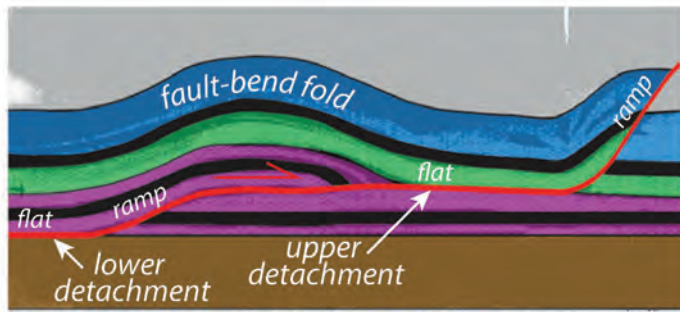


Figure 9. Labeled geometric features of the Rich model (Rich, 1934), with fault motion shown.

lower detachment is called the “floor thrust” and the upper detachment is the “roof thrust” (fig. 10). The term “duplex” first appeared in Dahlstrom (1970) and was later popularized by Elliott and Johnson (1980), and Boyer and Elliott (1982). Figure 10 shows how the displacement magnitudes (d_1 – d_4) of the individual imbricates (1–4) transfer all the slip from the floor to the roof thrust. Note that displacement decreases on the floor thrust as it increases on the roof thrust. The Waterton gas field in Alberta (fig. 11) is an example of such a fault duplex.

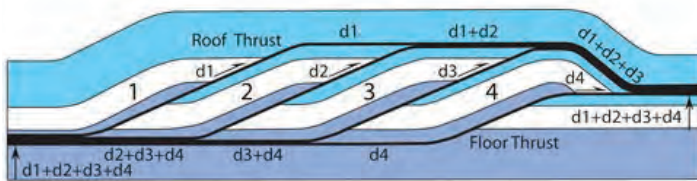


Figure 10. Labeled features of a fault duplex, from Boyer and Mitra (2019). Fault imbricates (1–4) and their corresponding displacement magnitudes (d_1 – d_4) are labeled. Thickness of the fault planes are proportional to the amount of displacement.

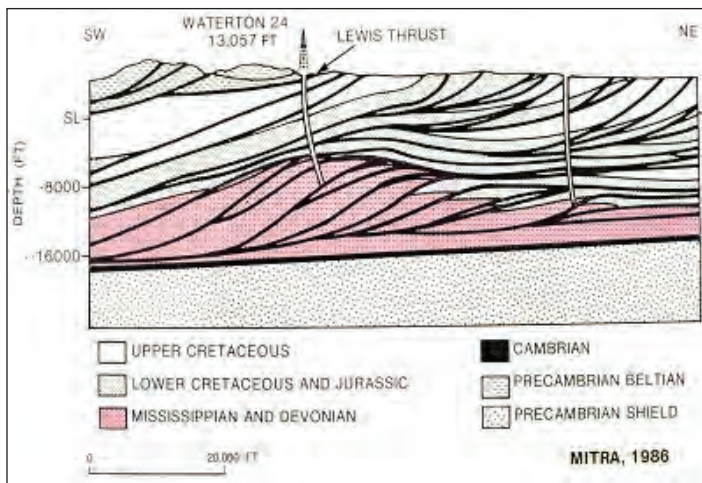


Figure 11. Schematic cross-section of the Waterton gas field, showing duplexes at depth. Reproduced from Mitra (1986).

Fault duplexes are not the only way to transfer slip between a floor and roof thrust. A fold duplex plays the same role as a fault duplex, transferring displacement from a lower detachment to an upper detachment, but in this case the transfer mechanism is a train of detachment folds (fig. 12). This model was first presented by Dahlstrom (1969), who argued that folds above a detachment required an upper detachment as well. The tight anticlinal cores require an underlying detachment. By the same argument the synclines, a mirror image of the anticlines, also require an overlying detachment. Dahlstrom (1969) termed this “the upper detachment in concentric folding”. Such a structure might also be called a “fold duplex” (Boyer and Mitra, 2018, 2019). Such fold duplexes are common throughout the Rocky Mountain region. Figure 13 shows an example of a fold duplex from the northern part of the Sawtooth Range thrust belt in northwestern Montana. In this example, a roof thrust separates vertical isoclinal folds in Cambrian carbonates from more open folds in the overlying Devonian rocks. Note that the roof thrust itself is folded, demonstrating additional local slip on the blind floor thrust (Singdahlsen, 1986).

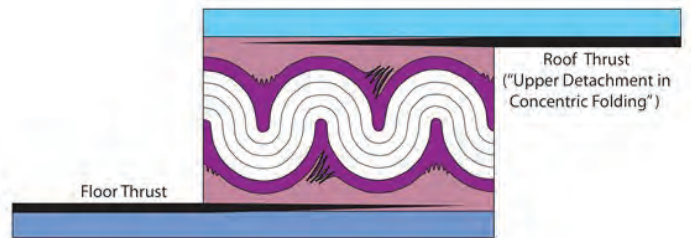


Figure 12. Conceptual model of a fold duplex, from Boyer and Mitra (2019), modified from Dahlstrom (1969). Thickness of the fault planes are proportional to the amount of displacement.

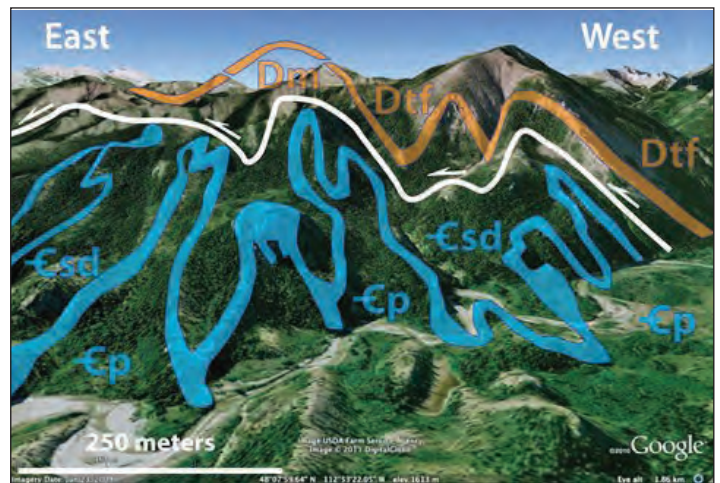


Figure 13. An annotated Google Earth image showing fold duplexes near Swift Reservoir. Outcrops of the Steamboat limestone are highlighted in blue. See Singdahlsen (1986) for more discussion and unit abbreviations.



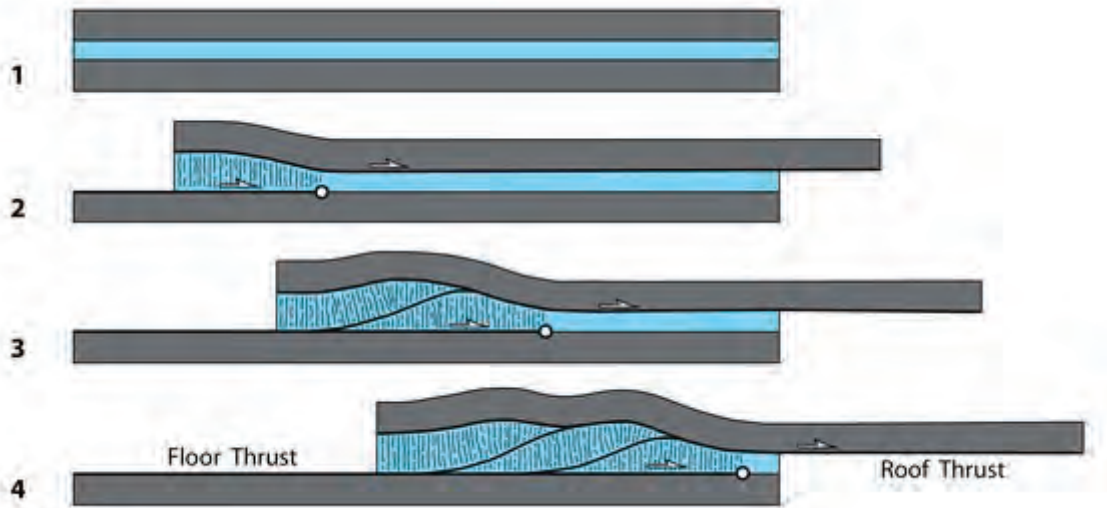


Figure 14. Formation of an active-roof hybrid cleavage-fault duplex. Small circles show the active fault tip. From Boyer and Mitra (2019), modified from Nickelsen (1986).

Pressure-solution cleavage can also transfer displacement between floor and roof thrusts, forming a cleavage duplex. Pressure-solution cleavage is common in argillaceous limestones and limy mudstones, where small grains and tiny fossil fragments act as stress concentrators, thereby dissolving calcium carbonate along grain boundaries and leaving behind insoluble residuum. The calcium carbonate solution often moves to strain shadows at the ends of the grains or fills tension fractures as they open. In an “open system” the fluids may migrate much farther, sometimes far out into the foreland. For some additional background on pressure-solution cleavage and the role it plays in tectonic shortening, see Nickelsen (1972), Ihle (1988), Mitra and others (1984), and Protzman and Mitra (1990).

Due to the strain rate dependance and strain hardening effects of pressure solution, hybrid cleavage-fault duplexes are common (fig. 14). During initial shortening, or at low strain rates, shortening may be entirely accommodated by pressure solution. Where a floor thrust loses displacement into a zone of cleavage, a cleavage duplex will develop (fig. 14.2), transferring slip to a roof thrust. Pressure solution selectively removes the relatively weaker, more soluble calcium carbonate from areas of concentrated stress, leaving behind relatively stronger material. As a result, the host rock becomes more brittle. When it is easier to break the cleaved rock, rather than to continue shortening by pressure solution, imbricate faults cut across the former cleavage duplex (fig. 14.3). A hybrid cleavage-fault duplex may develop as this strain cycle oscillates (fig. 14.4). In passive cleavage duplexes, excess bed length in the hanging wall of the roof thrust may be accommodated by faulting, folding, or pressure solution (Boyer and Mitra, 2019). As a result, natural examples of cleavage-fault duplexes may be quite complicated (fig. 15).

Hybrid duplexes may link slip from a floor to a roof thrust by any combination of faults, folds, and cleavage, depending on the material properties of the deformed rocks. Figure 16 shows an example of a cleavage-fold duplex, adapted from Coogan and Royse (1990). At the base of the section is a fault duplex involving the Paleozoic clastic and carbonate rocks, as well as Triassic strata (Tr) and the Jurassic Nugget sandstone (Jn). The roof thrust for this lower fault duplex lies above the Jurassic Nugget sandstone (Jn) and the Twin Creek limestone (Jt). Between the Meade thrust and the lower fault duplex is a cleavage-fold duplex of the Twin Creek limestone (Jt). When the Twin Creek (Jt) was flat-lying, prior to any deformation, it experienced layer-parallel shortening via the formation of pressure-solution cleavage (Protzman and Mitra, 1990). This may have occurred as the Meade thrust began to move. As the Meade thrust sheet overrode the cleaved Twin Creek limestone (Jt), the limestone continued to deform by folding, forming a hybrid cleavage-fold duplex. This example

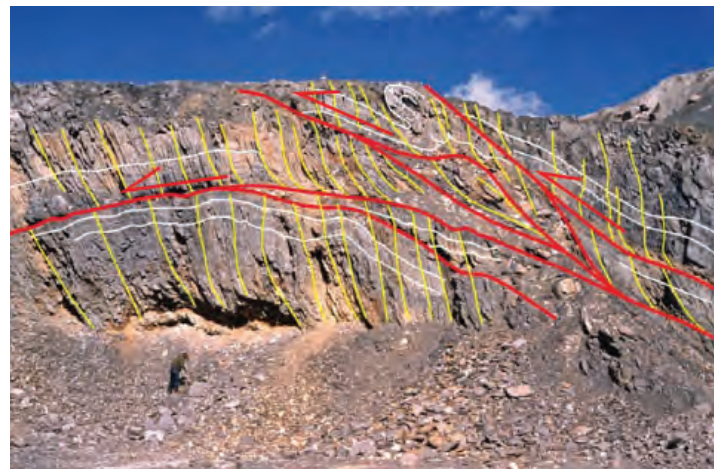


Figure 15. An example of a cleavage-fault duplex in Silurian–Devonian limestones in the Appalachian fold-thrust belt of central Pennsylvania. Yellow lines highlight cleavage trajectories. White lines highlight bedding. Red lines show inferred thrust faults.

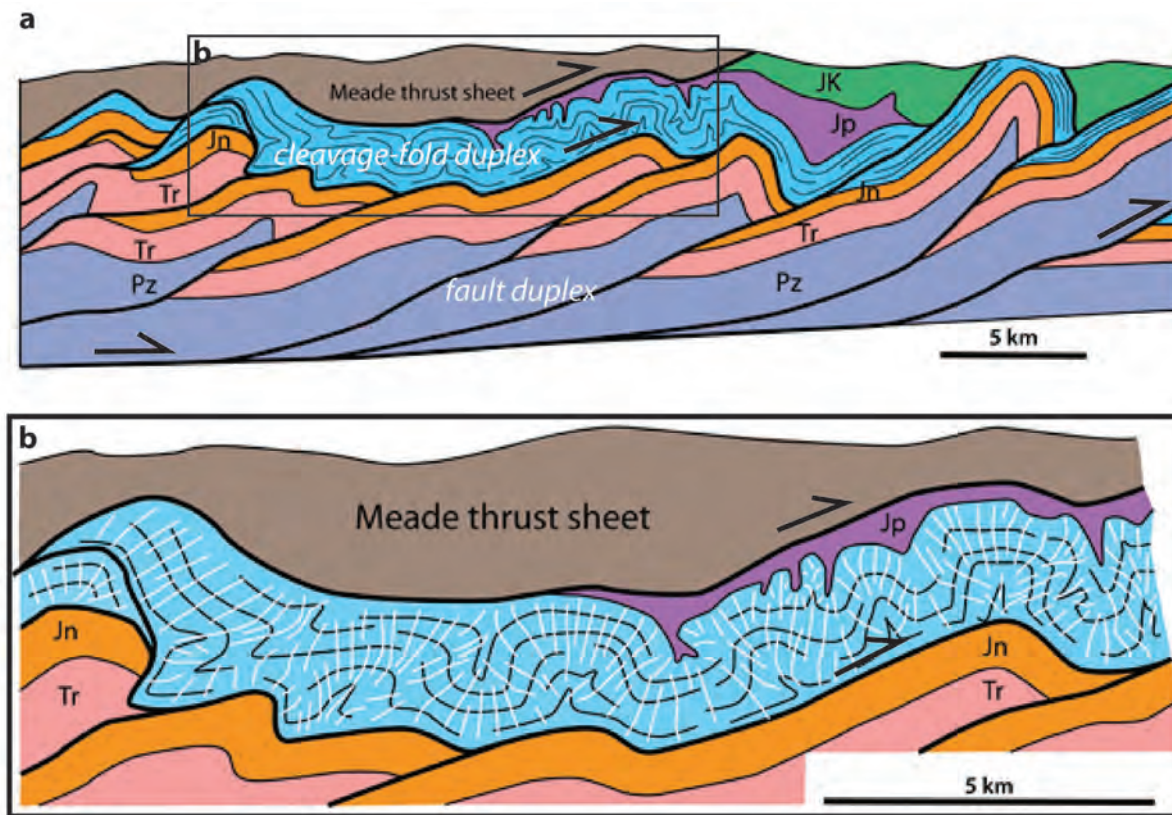


Figure 16. Examples of multiple hybrid duplexes beneath the Meade Thrust of the Wyoming Salient, adapted from Coogan and Royse (1990). White lines show cleavage trajectory. Arrows highlight the main detachment horizons that define the duplexes.

demonstrates how duplexes of different types and scales may work together to accommodate shortening and transfer strain forward and up-section in mechanically stratified rocks.

With this background, we will now begin our field trip to duplexes along the Sandy Hollow thrust system.

ROAD LOG

From Bannack State Park, take MT-278 east and I-15 north to exit 85 at Glen, Montana. It is a 46 mi (74 km), about 45 min drive. At the end of the offramp, set your odometer to zero. See figure 3 for an overview map of the field trip stops.

0.0 mi (0.0 km) From the exit, take a right (east) onto Brownes Bridge Rd.

0.1 mi (0.2 km) Turn right (south) onto MT-91.

4.3 mi (6.9 km) Turn left (east) onto Burma Rd. Pass the Glen Campground and the Big Hole River. Follow Burma Rd. along the bench that marks the dissected toes of the older alluvial fans on your left (north). On the left, you will pass a gulch with powdery white outcrops, followed by resistant dip panels of brown and red sandstones. This marks the faulted Tertiary/Cretaceous contact at the western edge of the Sandy Hollow uplift.

9.6 mi (15.4 km) After passing a small ranch on the right and an orange outcrop on the left, turn left onto the red dirt road and park at the mouth of Ziegler Gulch. This is STOP 1 (fig. 3).

STOP 1—Structural style within the hinge of the Ziegler Gulch Syncline

45.442002°, -112.589373°

From the parking spot, hike back (west) along Burma Rd. to view the orange outcrop of deformed limestones just north of the road.

Outcrop 1a: Cross-section view of the deformed Middle Limestone member of the Kootenai

45.442525°, -112.591095°

At this outcrop, we invite you to participate in a warm-up exercise. Standing back along Burma Rd., take 15–20 minutes to sketch the outcrop. As you scan the outcrop, do you notice any faults that break the continuity of the layers? Sketch these faults, and the different characteristic layers on either side. Pay close attention to the color, outcrop habit, and deformation style of the different layers. Try to find correlative layers on each side of the fault that can be used to constrain the sense of offset. After you have identified the main faults, describe the stratigraphy, and note how each unit deforms. What stratigraphic contacts act as detachment horizons? How are different detachment horizons linked?



Our interpretation is shown in figure 17. Kiorpes (1990) studied this outcrop in much more detail, meticulously restoring each bed set (fig. 18). They found that bed lengths do not restore, indicating syndeformational volume loss by pressure solution in unit *c* (fig. 18a). Cleavage in unit *c* accommodated about 22 m (70 ft) of preferential shortening in this bed. Their restoration suggests that the early structure consisted of fault duplexes in units *d* and *a*, and a cleavage duplex in unit *c* (fig. 18b). Pressure solution and strain hardening of unit *c* promoted out-of-sequence thrusts that cut across

all units, linking detachments above and below the middle limestone member (fig. 18c). The final product is the duplex seen in outcrop.

Once you are satisfied with your interpretation, walk back to the mouth of the canyon, and view the continuation of this outcrop as it wraps into Ziegler Gulch. If you'd like, climb the short, steep ridge to the east to get a better view.

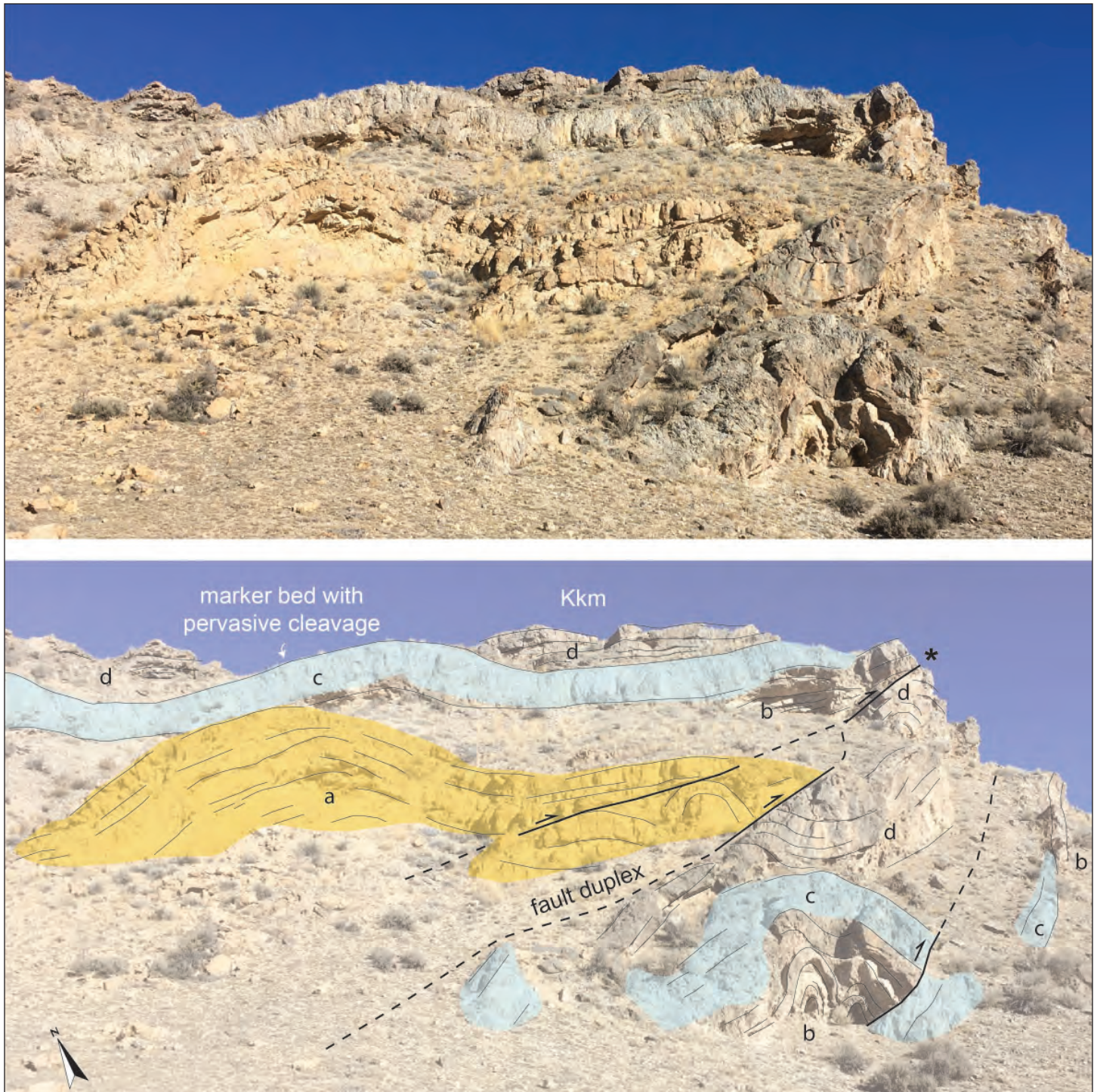


Figure 17. Outcrop photo and our interpretation of Outcrop 1a. Recognizable units of the Middle Limestone member of the Kootenai Formation (Kkm) are labeled a–d. *Reference point shared with figure 19.



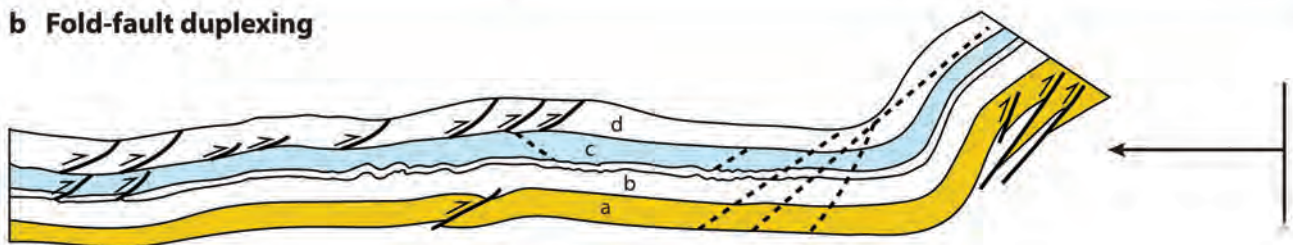
a Early cleavage duplexing**b Fold-fault duplexing****c Late fault duplexing**

Figure 18. Balanced and restored cross-section of Kiorpes (1990). In each time step, shortening is shown as an arrow. Dashed lines show trajectory of future thrusts. Colors and unit labels correspond to those in the previous figure.

Outcrop 1b: Oblique view of the Ziegler Gulch syncline

45.442135°, -112.588601°

Viewing the long continuation of the roadside outcrop and using your previous sketch as a reference, see if you can trace the thrust faults into the along-strike view. Using the generalized stratigraphic model that you developed in the previous sketch, see if you can find any repetitions of the units. Once again, what contacts act as detachment horizons? Is the shortening magnitude consistent along strike?

Looking north, notice the small cliffy knob of tightly folded limestone. This is the Gastropod Limestone member of the Kootenai Formation. The red slope below is the recessive upper clastic member, and you are standing on the lower clastic member (see fig 2. for reference). From this vantage point, you have a nice oblique view of the NNE-plunging, ESE-verging Ziegler Gulch syncline (fig. 19).

This overview illustrates the typical range of deformation styles within the various siliciclastic and carbonate members of the Kootenai Formation. The siliciclastic members contain thick packages of resistant, relatively strong sandstones and are folded with large wavelengths of about a kilometer (1/2 mi). The carbonate members are weaker and are folded, faulted, and

cleaved at much smaller scales. As a consequence of this mechanical stratigraphy, the folded members of the Kootenai Fm. within the Ziegler Gulch syncline are decoupled from one another. As is evident in oblique views like this one, and from map relationships, strain is concentrated within the weaker carbonate members, in the tight hinges of the folds. Fold, fault, and cleavage duplexes accommodate shortening within the thickened fold hinge, transferring strain between detachments with flexural slip between the siliciclastic and carbonate members of the Kootenai Formation.

Return to the cars and drive to STOP 2.

From STOP 1, continue east on Burma Rd., passing a small gulch, and wrapping around a large yellow-brown outcrop on the left.

10.0 mi (16.1 km) Park on the left, in the small unnamed gulch, across from the ranch entrance and before the road bends hard to the right.

STOP 2—Overview of the Sandy Hollow thrust

45.440363°, -112.583063°

From the parking spot, hike 0.2 mi (0.3 km) along the Burma Rd. (southeast), to view the impressive outcrop of dark resistant bedded rocks above the red recessive slopes, just north of the road. For those that want a closer look at the fault zone, there is a sheep trail within the footwall.



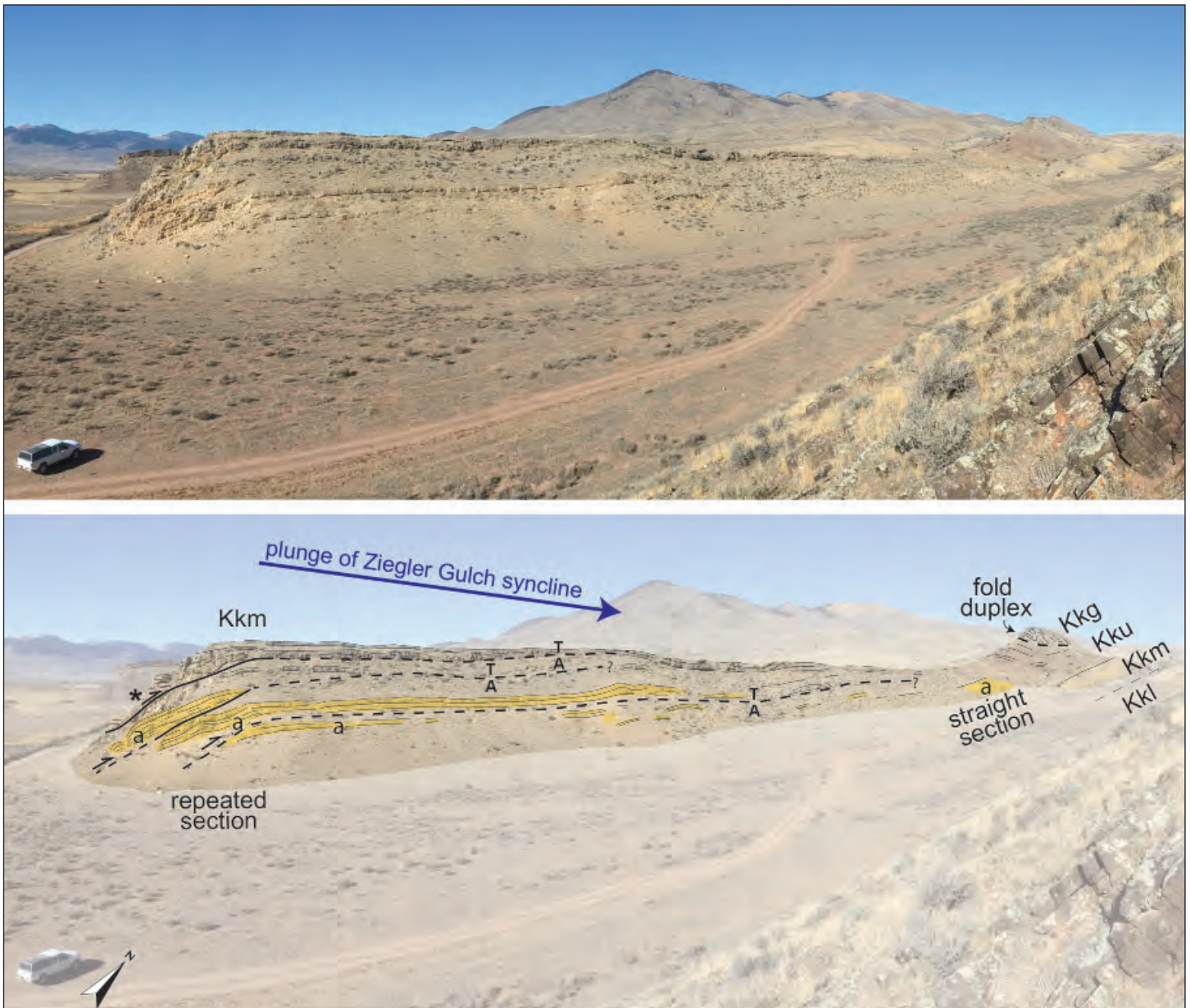


Figure 19. Oblique view of the Ziegler Gulch syncline hinge, showing fault duplexes of the Middle Limestone member (Kkm) and a fold duplex in the Gastropod Limestone member (Kkg) of the Kootenai. Thrust traces and kinematics labeled: T, towards; A, away. *Reference point shared with figure 17.

This outcrop gives a great cross-sectional view of the Sandy Hollow thrust (fig. 20). From the previous stop, we have driven down-section. Below the Kootenai Fm., we passed the covered interval of the Jurassic Morrison Fm., in the small unnamed gulch. To the northeast, you can see an overturned fold in the brown-yellow lime-wackestones and siltstones of the Triassic Dinwoody Formation. The dark gray rocks of the skyline belong to the Permian Phosphoria and Shedhorn Formations. The red sandstones and shales of the footwall belong to the lower clastic member of the Kootenai Formation. At this outcrop, there is about 1,400 ft (427 m) of stratigraphic displacement across the fault, the maximum amount anywhere along the Sandy Hollow thrust (Brumbaugh and Dresser, 1976).

The Sandy Hollow thrust dips moderately to the west-northwest. Measured fault planes in the footwall approximate the general dip of the main thrust, at about 20° towards an azimuth of 315°. Bedding on either side of the fault is apparently parallel, dipping to the west-northwest at about 30–35°, suggesting a flat-on-flat relationship. Slickenlines near the main fault plane and within a small synthetic fault in the footwall suggest a thrust sense of slip towards the east-southeast (105°). This shortening direction is perpendicular to the trace of mapped and inferred folds defined by plotted poles to bedding planes, and is also consistent with other available kinematic data (ex. Brumbaugh and Hendrix, 1981). About 10° of clockwise vertical-axis rotation has been documented in the lower clastic member of the Kootenai Formation in the footwall of the Sandy

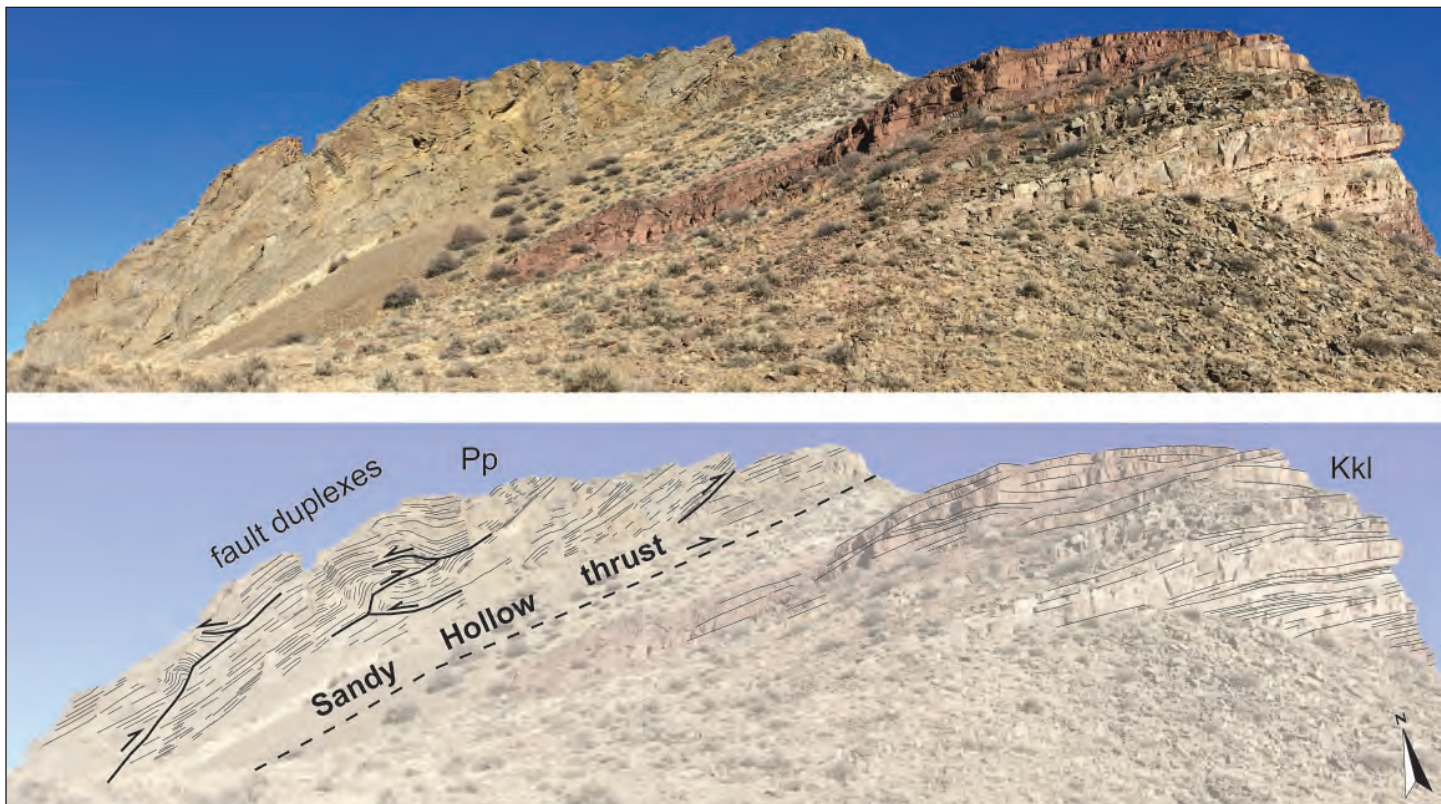


Figure 20. View of the Sandy Hollow thrust from Burma Rd.

Hollow thrust, suggesting that apparent shortening directions were rotated during formation of the later Sandy Hollow anticline (Eldredge and Van der Voo, 1988). Like the structurally lower Hogback thrust, the Sandy Hollow thrust cuts ESE-verging folds in both the hanging wall and footwall, suggesting it is a broken fault-propagation fold.

While the hanging wall rocks look fairly coherent at first glance, they are actually quite deformed. Careful tracing of apparent layers in the hanging wall defines several boxy folds and numerous truncated bedding planes. Several small thrusts with moderately dipping ramps branch off of inferred flat detachments. We interpret these geometries as fault-bend folds within several small fault duplexes (fig. 20), which internally thicken the Phosphoria and Shedhorn Formations in the hanging wall of the Sandy Hollow thrust. The duplexes continue to the north, ultimately leading to the tight hinge of the Ziegler Gulch anticline, which we will see at the next stop.

In the footwall, a careful line tracing also reveals a few apparent truncations of the visible layering. The traced layers defined gently curved surfaces that are truncated by overlying strata. We interpret these geometries not as fold-thrust structures, but as large fluvial bedforms, as are typical in the classic members of the Kootenai Fm. (Rosenblume and others, 2021). Like we've seen at the previous stop, the strong sandstones of the Kootenai Fm. have negligible internal deformation and rarely host outcrop-scale structures.

Return to the cars and drive to STOP 3.

From STOP 2, Continue driving east on Burma Rd., through the notch carved by the Big Hole River.

11.1 mi (17.9 km) Any two-wheel-drive vehicles may want to park at the Tony Schoonen fishing access parking lot and car pool to STOP 3.

11.6 mi (18.7 km) Turn left (north) onto the BLM road, into Sandy Hollow. Follow the main dirt road north.

12.2 mi (19.6 km) Bear left at the fork, to follow the main road.

13.1 mi (21.1 km) Park at the large parking lot on the left. There are two entrance/exits to this large pullout, but be sure not to park the group in.

STOP 3—Hike along the folded hanging wall of the Sandy Hollow thrust

45.461558°, -112.567719°

From the parking spot, follow the game (geologist) trail north. Hike about 0.7 mi (1.1 km) north, to low, exposed green outcrops of the Morrison Formation.

As we hike north (structurally up) along the fault contact, take note of the deformation styles of the different units in the hanging wall. This is the northern continuation of the high strain zone that we viewed at the last stop, within the hinge of the Ziegler Gulch anticline. Based on our earlier stops, try to locate yourself within the stratigraphy as we walk along. On the way back, we



will take a closer look at outcrop-scale structures and try to gain a more complete understanding of the fault and fold geometries in the hanging wall.

STOP 3a—Morrison/Kootenai contact

45.470258°, -112.564207°

These small outcrops offer rare exposures of the Jurassic Morrison Formation. Feel free to explore, and notice the many carbonate and siliciclastic lithologies present in this thin unit. Looking to the west and north, you'll see some oblique views of parasitic folds within the Dinwoody Formation. These tight folds are in the hinge of the Ziegler Gulch anticline, in the hanging wall of the Sandy Hollow thrust. Figure 21 shows a clearer, down-plunge aerial view and interpretation of this high strain zone within the core of the anticline. Note that the second- and third-order folds in the Dinwoody do not involve the overlying Morrison or underlying Shedhorn and Phosphoria Formations. This suggests that roof and floor thrusts bound the Dinwoody. Particularly in the west limb of the anticline, you can see some small thrusts that clearly cut the upper Dinwoody Formation and its second- and third-order folds. These thrusts generally merge into the inferred floor and roof thrusts near the upper and lower contacts of the Dinwoody. From the right vantage point, you may

also notice that the underlying Shedhorn and Phosphoria are repeated, like we saw at STOP 2. Unlike at STOP 2, the Shedhorn and Phosphoria here are apparently folded into a tight isocline. This isocline formed at depth during initial fault propagation folding, as the Sandy Hollow thrust broke upward from near the base of the Phosphoria Formation. This isocline is analogous to the lower chevron fold shaded yellow in figure 6.

The clifty outcrops stratigraphically above (east of) the Morrison consist of the resistant basal conglomerate of the Kootenai Formation. From this vantage point, try to follow the Morrison/Kootenai contact across the skyline (refer to fig. 21 for an aerial view). You'll notice that the contact is mostly continuous, defining the geometry of the Ziegler Gulch anticline (* in fig. 21). Above this contact, an "upper package" forms a continuous fold, like the Ziegler Gulch syncline observed at STOP 1. Below this contact, a "lower package" is more pervasively strained, with duplexes on many scales. A detachment is inferred between these two packages, essentially acting as a roof thrust to a fault-fold duplex. This detachment also separates the deeper, isoclinal part of the hinge from the upper, more open part of the fold. In reference to figure 7, this detachment approximately coincides with the line of intersection (C). The

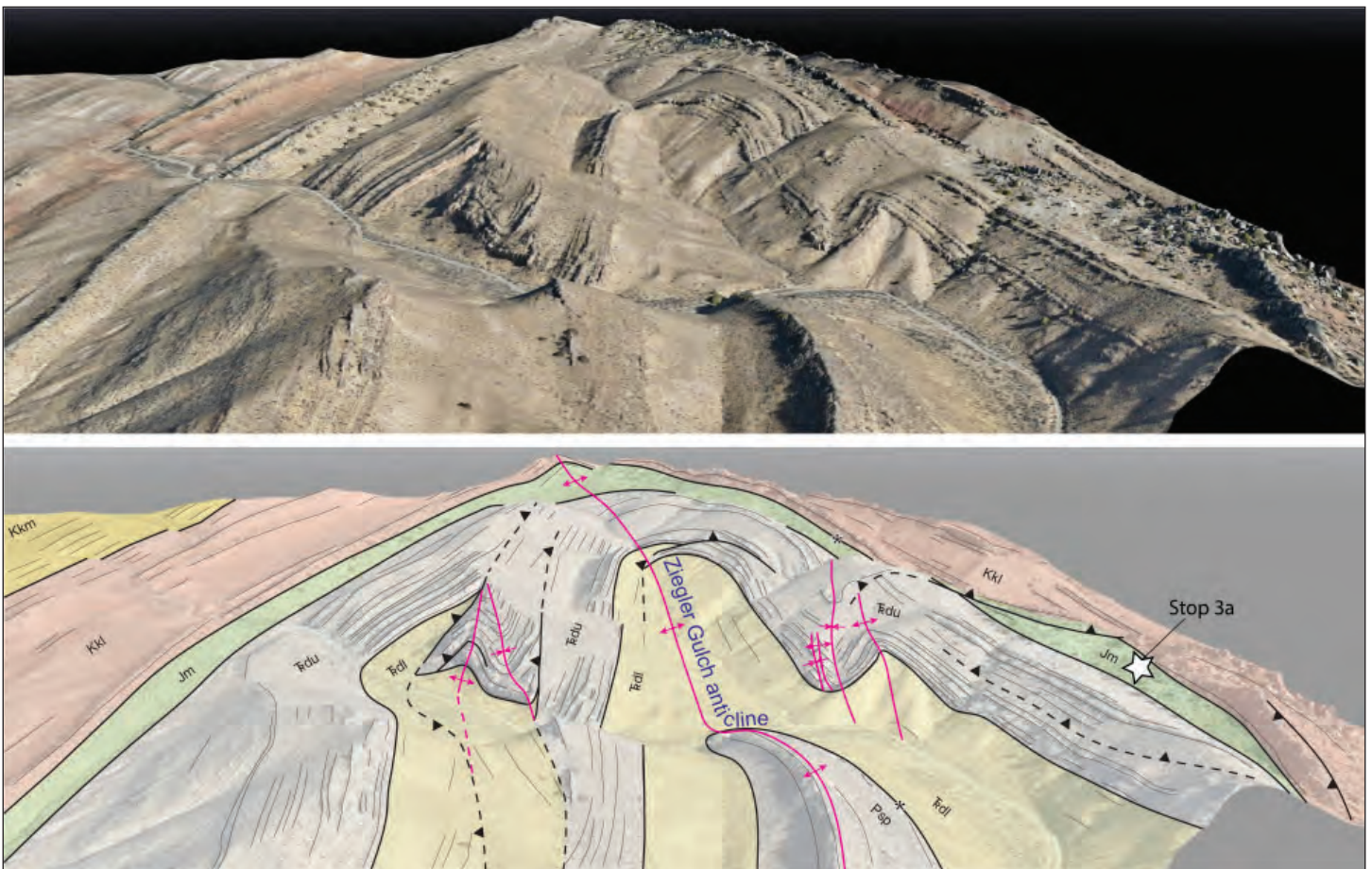


Figure 21. Down-plunge aerial view and interpretation of the high strain zone within the lower package in the hinge of the Ziegler Gulch anticline. Drone imagery is from Chris Langhoff and Celine Beaucamp-Stout. *Inferred detachment between upper and lower package.



structure of the Ziegler Gulch anticline is more complicated because of the several detachment horizons that shared flexural slip during fault-propagation folding, similar to the multi-mode fault-propagation model of Erslev and Mayborn (1997; fig. 18).

Just to the south, you can see one of the rare small faults that cuts the contact, offsetting the cliff band of the basal Kootenai conglomerate. We will discuss this small structure briefly at the next stop. Try to get a sense of the geometry of the broken Ziegler Gulch anticline as we walk back (south) along the contact.

Walk back to the parking spot, generally following the basal Kootenai contact until you hit the thrust contact. The next stops mark particular points of interest along the hike back.

STOP 3b—Outcrop-scale folds in the upper Dinwoody Formation

45.467104°, -112.565430°

From this vantage point, the faulted basal conglomerate of the Kootenai Fm. is visible to the north–northeast. There are also a few knobs of the same resistant basal conglomerate (Kootenai) visible to the south, which we will pass on our walk back. Notice how the northern exposures of the basal conglomerate abruptly stop just to the north. Which way do they dip? Are they upright or overturned?

Here, the Dinwoody Fm. is apparently thrust over the Kootenai, concealing the overturned basal conglomerate in the footwall. Look carefully at the surrounding Dinwoody, noticing the small, outcrop-scale fold hinges. Follow the trace of these fold axes, to try to get a sense of which way is stratigraphic up, and how repeated the section is. Notice the irregular trace of these fold axes. Rare exposures show decoupling of the tabular beds, often with pervasive deformation approximating granular flow within the fine-grained lithologies that form the cores of the folds. You will find that many of the folds are isoclinal and recumbent. Here the eastern limb is faulted, with an overturned recumbent fold in its hanging wall. In contrast, to the north (structurally higher), the same eastern limb of the Ziegler Gulch anticline is upright with a moderate to steep dip.

STOP 3c—Fault plane of a hanging wall imbricate of the Sandy Hollow thrust

45.465331°, -112.565778°

This small outcrop gives a rare chance to stand on the fault plane of an imbricate thrust within the Sandy Hollow thrust zone. This thrust plane is a continuation of the fault trace discussed at the previous stop, that places the Dinwoody on the Morrison/Kootenai. The thrust dips to the east–northeast, at a moderate angle

of about 20°. Notice that the slickenlines on the fault plane are oblique to the dip of the fault plane. The older on younger relationships signify thrust motion, towards the southeast (130°). Recall our kinematic indicators near the Sandy Hollow thrust along Burma Rd. (STOP 2), which were more eastward (105°).

Notice how the outcrops of the basal conglomerate in the footwall end abruptly, with gray limestones beneath. Beneath the gray limestones, you can see the continuous, northeast-dipping east limb of the Sandy Hollow anticline. The block we are standing on is a large horse (fault-bounded block), composed mostly of the competent basal conglomerate of the Kootenai. Beneath us is another large horse, of the Gastropod Limestone member of the (upper) Kootenai Fm., thrust on the younger Blackleaf Formation. The outcrops of the basal conglomerate of the Kootenai, as viewed from north to south, show progressively increasing strain; from first-order folding above duplexes in the underlying Dinwoody Fm., to a faulted and tightly overturned anticline, to completely isolated horses. This illustrates how strain proceeded, as the fault tip propagated through the verging fold pair and coalesced into the Sandy Hollow thrust. The 25° disparity between the orientation of the slickenlines above this horse as compared to those observed at STOP 2 may suggest clockwise rotation of this horse during slip on the overlying Sandy Hollow thrust.

This spot coincides with an important bend or “step” in the Sandy Hollow thrust (Brumbaugh and Dresser, 1976). To the south and west, duplexes of the lower package (Phosphoria–Dinwoody Formations) are thrust over the upper package (Kootenai–Blackleaf Formations). To the east, the upper package is duplicated, with progressively less stratigraphic offset to the north. At the step, the thrust abruptly cuts across the lower package and into the upper package, isolating large horses of the upper package. In the hanging wall, the lower package is much more tightly folded than the upper package. These observations suggest that the Sandy Hollow thrust is a broken fault propagation fold that has subsequently been folded by the structurally lower Sandy Hollow anticline.

Slip first occurred on several weak horizons in the Phosphoria and Dinwoody Formations, defining a lower package of duplexes (fig. 2). This lower package represents the triangular shear zone or chevron fold that formed at depth as the fault tip propagated. The duplexes of the lower package that make up this triangular shear zone transferred slip from the decollement near the base of the Phosphoria Fm. to the base of the upper package during initial fault propagation folding. The relationships north of the step (fig. 21) show how, during



intermediate shortening, slip was partitioned between a floor thrust that partly cut across the lower package, and a roof thrust that isolated a continuous upper package. The step area shows how the thrust ultimately cut across the lower package of duplexes, linking the floor and roof thrusts to form a single continuous thrust.

STOP 4—Overlook of the “Sandy Hollow collision structure”

45.460180°, -112.570720°

From the parking spot, look to the northeast, towards the highly deformed carbonates in the footwall of the thrust. For a better view, hike up the steep hill to the immediate southwest. This overlook gives a nice over-view of the core of the Sandy Hollow uplift.

The impressive series of folded gray-brown limestones in the immediate footwall of the Sandy Hollow thrust have long captivated geologists and horrified students. Many field camp students and instructors have referred to this general area of high strain, near the intersection of the Sandy Hollow anticline hinge and the Sandy Hollow thrust, as the “rat’s nest.” Hendrix and Porter (1980) referred to this specific outcrop as the “Sandy Hollow collision structure” (fig. 22).

On the left (northwest) side of the outcrop, chocolate brown limestones of the Triassic Dinwoody Formation are thrust over the Gastropod Limestone of the Kootenai. At this contact, the Sandy Hollow thrust coincides with the hinge of the north-plunging Sandy Hollow anticline (in the footwall). On the scale of this outcrop, the collision structure resembles an asymmetric wedge, with its thickest portion near the thrust, similar to the classic “sand box” model that recreates a fold-thrust wedge in front of a sliding backstop. The upper 29 m of the Gastropod Limestone have been scraped off the lowest 7 m (23 ft; Hendrix and Porter, 1980). Above this floor thrust, the Gastropod Limestone deforms mostly as a passive-roof fold duplex (Boyer and Mitra, 2019), within the forelimb of the Sandy Hollow anticline. Outcrops of the overlying Blackleaf Formation

are apparently continuous, suggesting a roof thrust in the covered interbedded shale and limestone that makes up the gradational contact. Strain decreases away from the hinge, eventually dying out about 450 m (1/4 mi) away from the thrust (Hendrix and Porter, 1980). Kinematic indicators from small faults within the collision structure suggest a top-to-the-east–southeast (099°) sense of slip, parallel to the inferred slip of the Sandy Hollow thrust (Hendrix and Porter, 1980).

Near the middle of the collision structure, a small thrust cuts across the entire wedge-shaped fold duplex. This late-stage thrust suggests that with increasing strain the floor thrust was abandoned and a new detachment was initiated lower in the section, similar to what we saw at STOP 1. It is unclear whether this late-stage thrust merges into the roof thrust, dies out beneath it, or cuts it. Regardless, this late-stage thrust is “caught in the act” on its way to isolating a large horse like those observed earlier (STOP 3c). The small-scale detachments within the fold duplex apparently locked up, and larger-scale detachments emerged with increasing strain. Regardless of scale or timing, these structures effectively act as duplexes, linking strain between the various weak stratigraphic layers that acted as detachment horizons.

In detail, the Sandy Hollow collision structure contains countless smaller fault, fold, and cleavage duplexes. Hendrix and Porter (1980) painstakingly measured and described the stratigraphic units of the Gastropod Limestone and defined structural units based on deformation style (fig. 22). We invite you to walk throughout the collision structure and notice the range of structural styles that define these different structural units. See what scale and style (fault, fold, cleavage) duplexes you can find. See if you can locate the floor thrust, and the several intermediate thrusts within the overall fold duplex. This will give you a sense of how these structures collectively transferred slip from a floor thrust near the base of the Gastropod Limestone to a passive roof thrust near the top of the formation.

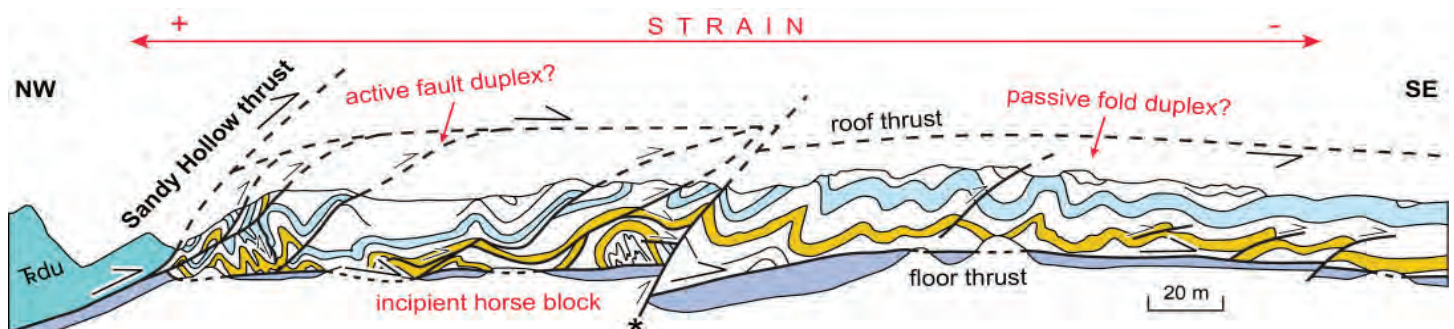


Figure 22. Schematic cross-section of the “Sandy Hollow collision structure,” modified from Hendrix and Porter (1980). Colors denote correlative layers. *Largest latest stage fault.

INTERPRETATION OF THE SANDY HOLLOW COLLISION STRUCTURE

We interpret the Sandy Hollow collision structure as a hybrid fold-fault duplex developed in the Gastropod Limestone, in the footwall of the Sandy Hollow thrust. The along-strike strain gradient illustrates how deformation progressed through time, as it advanced into the footwall of the active Sandy Hollow thrust. A fold duplex formed during the initial phase of deformation as slip was transferred between a floor and roof thrust bounding the Gastropod Limestone (Hendrix and Porter, 1980). Slip also occurred along several internal detachment horizons, resulting in many scales of fold duplexing. With increasing strain, late-stage faults cut folds, possibly linking the floor and roof thrust by small ramps across the former fold duplex. The largest late-stage thrust, in the middle of the outcrop (* on fig. 22), is apparently the youngest fault. With continued offset, this latest thrust would completely isolate the duplex as a large horse, similar to those observed at STOP 3c. Together, these observations show that within the duplex: (1) slip was shared between multiple stratigraphically controlled detachment horizons, (2) strain generally progressed from folding to faulting, and (3) the scale of structures generally increased with increasing strain. The complicated fold-fault duplex of the Sandy Hollow collision structure developed during progressive shortening, as strain was transferred between the many available weak layers. Strain hardening resulted in the gradual abandonment of some detachment horizons, giving rise to the larger out-of-sequence thrusts.

On a larger scale, the floor thrust of the duplex branches off a footwall-flat segment of the Sandy Hollow thrust. This short flat connects to the larger footwall ramp, or “step” of Brumbaugh and Dresser (1976). With continued slip the earlier duplex was passively translated in the hanging wall of the Sandy Hollow thrust. These relationships suggest that the final ramp-flat-ramp geometry of the Sandy Hollow thrust was established by progressive deformation of a fold-fault-duplex. Slip likely ceased as the underlying Sandy Hollow anticline grew, rotating and folding the thrust out of a favorable orientation.

SUMMARY

The Sandy Hollow thrust may be broadly understood as a broken fault-propagation fold that developed within rocks with a pronounced mechanical stratigraphy (e.g., Erslev and Mayborn, 1997). A lower package of fault and fold duplexes within Permian to Jurassic rocks defines the high strain core of the fold, which formed at depth near the propagating fault tip. In the upper package, small fold duplexes developed within weak carbonates of the Kootenai Formation. These duplexes accommodated flexural slip within the hinges of the fault-propagation fold, well above the active fault tip. With progressive shortening, the Sandy Hollow thrust propagated through the lower pack-

age. At this point, duplexes transferred slip between the detachment horizons that would ultimately become the flat segments of the Sandy Hollow thrust. Flats developed in the weak carbonates and shales, while ramps developed in the stronger sandstones. Once the ramp-and-flat geometry of the throughgoing thrust was established, the Ziegler Gulch anticline was passively translated. The Hogback thrust continued to propagate, folding the overlying Sandy Hollow thrust. This essentially stacked the two broken fault-propagation folds on top of one another, creating the Sandy Hollow uplift.

This field guide highlights how duplexes not only accommodate folding of a mechanically stratified section, but also how they facilitate progressive growth of thrust faults by efficiently linking segments of detachment ramps and flats. The outcrops in this field guide are particularly important for adapting general concepts of fault-propagation folds to sedimentary rocks (e.g., Erslev and Mayborn, 1997). STOP 1 demonstrates how small duplexes in the hinge of the Ziegler Gulch syncline accommodate folding at high structural levels. STOP 2 displays numerous fault duplexes in the hanging wall of the Sandy Hollow thrust that formed over a flat detachment horizon near the base of the Phosphoria Formation. STOP 3 demonstrates how fold and fault duplexes in the lower package accommodated tight folding near the propagating fault tip. Finally, the complicated rat’s nest viewed from STOP 4 shows a snapshot of how early fold duplexes evolved to larger fault duplexes and were ultimately incorporated into the thrust zone once a throughgoing fault coalesced.

We hope this field guide demonstrates that in thin-skinned fold-thrust belts, duplexes of all styles and on many scales play a crucial role in transferring slip between detachment horizons. Understanding fault, fold, and cleavage duplexes is a prerequisite for understanding the complicated but common geometries found in many fold-thrust belts. Not only do duplexes highlight which stratigraphic units act as detachment horizons, they also record the evolving mechanics that developed during progressive shortening. We hope participants can apply these concepts to other localities in order to describe their geometry more precisely and gain a more complete understanding of their deformation history.

ACKNOWLEDGMENTS

Work in Sandy Hollow is indebted to the countless geologists who came before. We thank Rob Thomas for use of his unpublished field maps, which highlight the inferred detachment at the Morrison contact. David Pearson and Katie McDonald did much of the recent mapping, especially near the Hogback thrust. David Pearson also helped develop many of the cross-section interpretations. Chris Langhoff and Celine Beaucamp-Stout provided the unpublished drone imagery used in figure 21. Darrel Cowen provided a



copy of the hard-to-find Kiorpes thesis, which helped shape this field trip.

Sandy Hollow is a public outdoor classroom for budding geologists. We ask future guests to Sandy Hollow to respect the outcrops and tread lightly on this public land. Because Sandy Hollow is a popular map area for university field camps, we ask readers to please respect the instructors by not sharing this field guide or published geologic map with future or current field camp students.

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RECONSTRUCTING THE TIMING OF QUATERNARY GLACIATION IN THE PIONEER MOUNTAINS, MONTANA

Spruce W. Schoenemann

Environmental Sciences Department, The University of Montana Western, Dillon, Montana

INTRODUCTION

(excerpts from Schoenemann and others, QSR, 2023)

Located south of the Cordilleran and Laurentide Ice Sheets (LIS), alpine glaciers occupied numerous mountain ranges across the western Cordillera during the latest Pleistocene, spanning a wide range of climatic zones that likely controlled glacier behavior (Laabs and others, 2020; Licciardi and others, 2004; Thackray, 2008). The pattern of glacial retreat during the last glaciation in the northwestern United States reveals regionally variable responses to shared climatic forcings (Laabs and others, 2020; Licciardi and others, 2004; Thackray, 2008), however, variability in the timing of deglaciation between regions appears to be related to climatic setting, differences in glacier mass balance characteristics (precipitation versus temperature), and changing atmospheric patterns driven by ice sheet growth and decay (Laabs and others, 2020; Licciardi and others, 2004; Licciardi and Pierce, 2018; Oster and others, 2015; Ullman and others, 2014, 2016; Wong and others, 2016).

Because glacier extent is controlled by both temperature and precipitation (Leclercq and Oerlemans, 2012; Oerlemans, 2005; Roe, 2011; Roe and O'Neal, 2009), it is often difficult to disentangle which of these parameters controlled glacier response to past climate fluctuations (Oerlemans, 2005). To evaluate which climatic factor might explain the significant timing differences among glacier maxima of the interior northwestern U.S., Schoenemann and others (2023) focused on a once-glaciated Pioneer Mountain range with an arid climate whose glacier response is thus likely to be temperature-dependent and controlled by solar insolation (Huybers and Roe, 2009; Roe and O'Neal, 2009; Rupper and Roe, 2008). Schoenemann and others (2023) used cosmogenic ^{10}Be exposure dating of glacier moraines in the Pioneer Mountains of southwestern Montana to explore ice retreat timing and hence climate sensitivity in an arid region.

The Pioneer Mountains are well located to record responses of alpine glaciers to climate forcings. Due to the proximity of the LIS margin (~250 km north of the Pioneer Mountains), the region's glacial climate would likely have been precipitation-starved due to the dry, cold, katabatic winds descending off the ice sheet during the Last Glacial Maximum (LGM; Oster and others, 2015). As deglaciation of the LIS commenced, the formation of pro-glacial melt-

water lakes may have altered local precipitation patterns, supplying additional moisture relative to more distal alpine regions. Hence, these glaciated landforms archive valuable records of changes in temperature and regional precipitation, providing insight into the onset and drivers of deglaciation in the continental northwestern North America.

Previously published chronological data from the interior northwestern U.S. suggest that the glacial maxima observed in continental ranges (i.e., Wallowa, Wind River, and Big Horn mountains) followed insolation pacing, initiating deglaciation between 23 and 22 ka (Dahms and others, 2018; Laabs and others, 2020; Licciardi and others, 2004; Licciardi and Pierce, 2018; Phillips and others, 1997). Exposure ages from the northwestern (17.6 ka) and eastern (19.8 ka) greater Yellowstone glacial system (GYGS) do not show such an early response, and interpretation of its retreat timing is complicated by a migrating ice cap (Licciardi and others, 2001; Licciardi and Pierce, 2008, 2018).

PREVIOUS WORK IN THE PIONEER MOUNTAINS

In southwest Montana, glacial moraines, outwash, and erosional landforms indicate that many of the mountain ranges with land area above 2,500 m contained glaciers during the LGM (Locke, 1990; Porter, 1983). However, ice extent and glacier history are known only generally. Ice extents in the region are broadly delineated based on glacial geomorphic field mapping by Alden (1953). Glacial deposits mapped by Pearson and Zen (1985) and Ruppel and others (1993) helped to further establish glacial extents, although slightly different limits were outlined by Locke and Smith (2004), who derived ice limits from map and aerial photo interpretation.

During Pleistocene glaciations, ice extent expanded in southwest Montana, producing small ice fields with outlet glaciers in some ranges and alpine glaciers that carved and filled mountain valleys, preferentially in north-northeast aspects. Evidence for previous glaciers includes bowl-shaped cirques in the valley headwaters, which now contain small tarns or meadows, striations on bedrock, U-shaped valleys with obvious trimlines, erratic boulders, and moraines of bouldery till (Alden, 1953; Locke and Smith, 2004; Pearson and Zen, 1985). The East and West Pioneer Mountains are one of the largest ranges in the region (fig. 1), and during

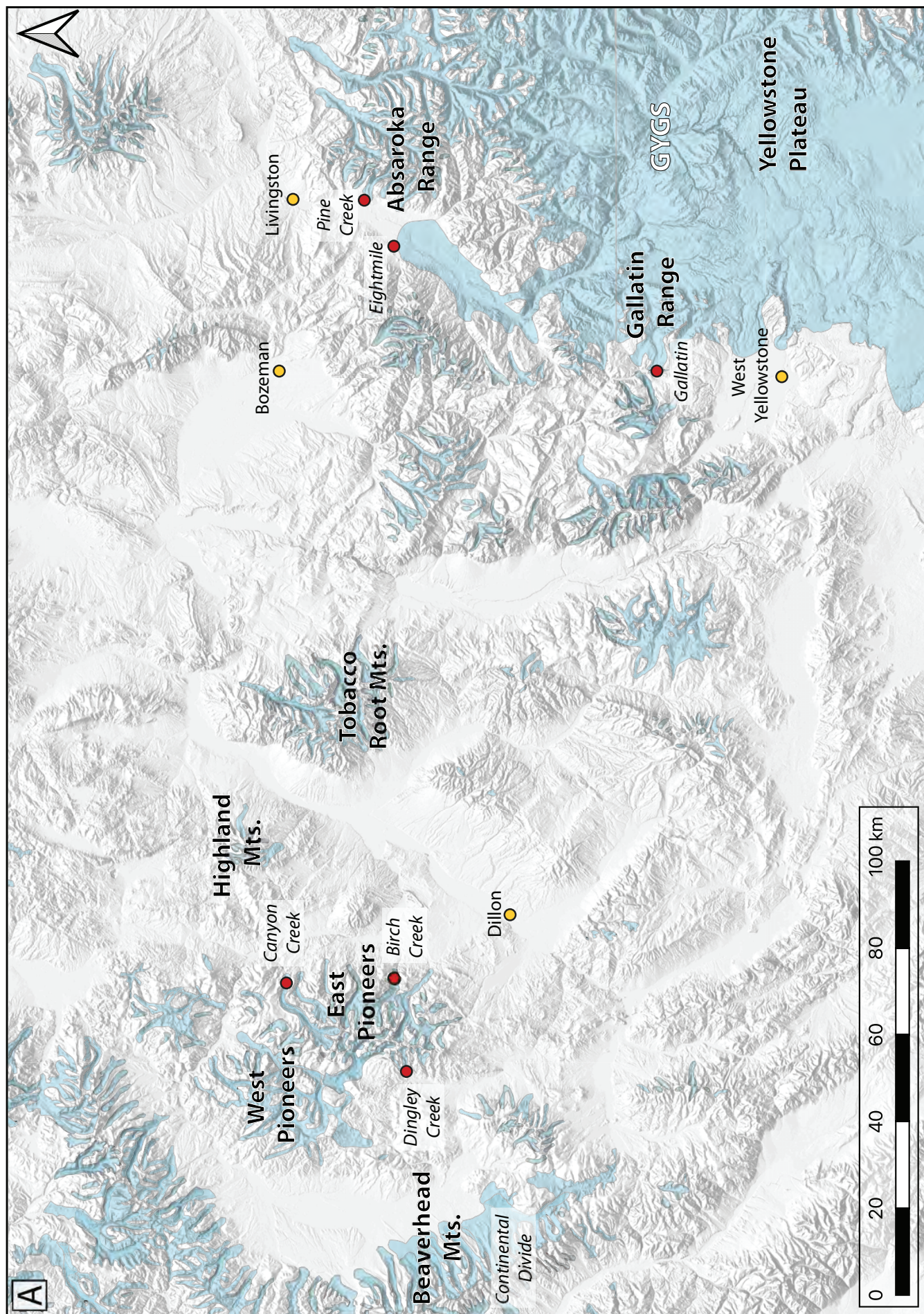


Figure 1A. Extent of Pinedale glaciation (outlines from Western U.S. Paleoglaciators) of southwest Montana and northwest Greater Yellowstone Glacial System (GYGS) overlain on a USGS shaded relief basemap. Relevant mountain ranges (black), glacial valleys (italics) and sample locations (red circles) included.

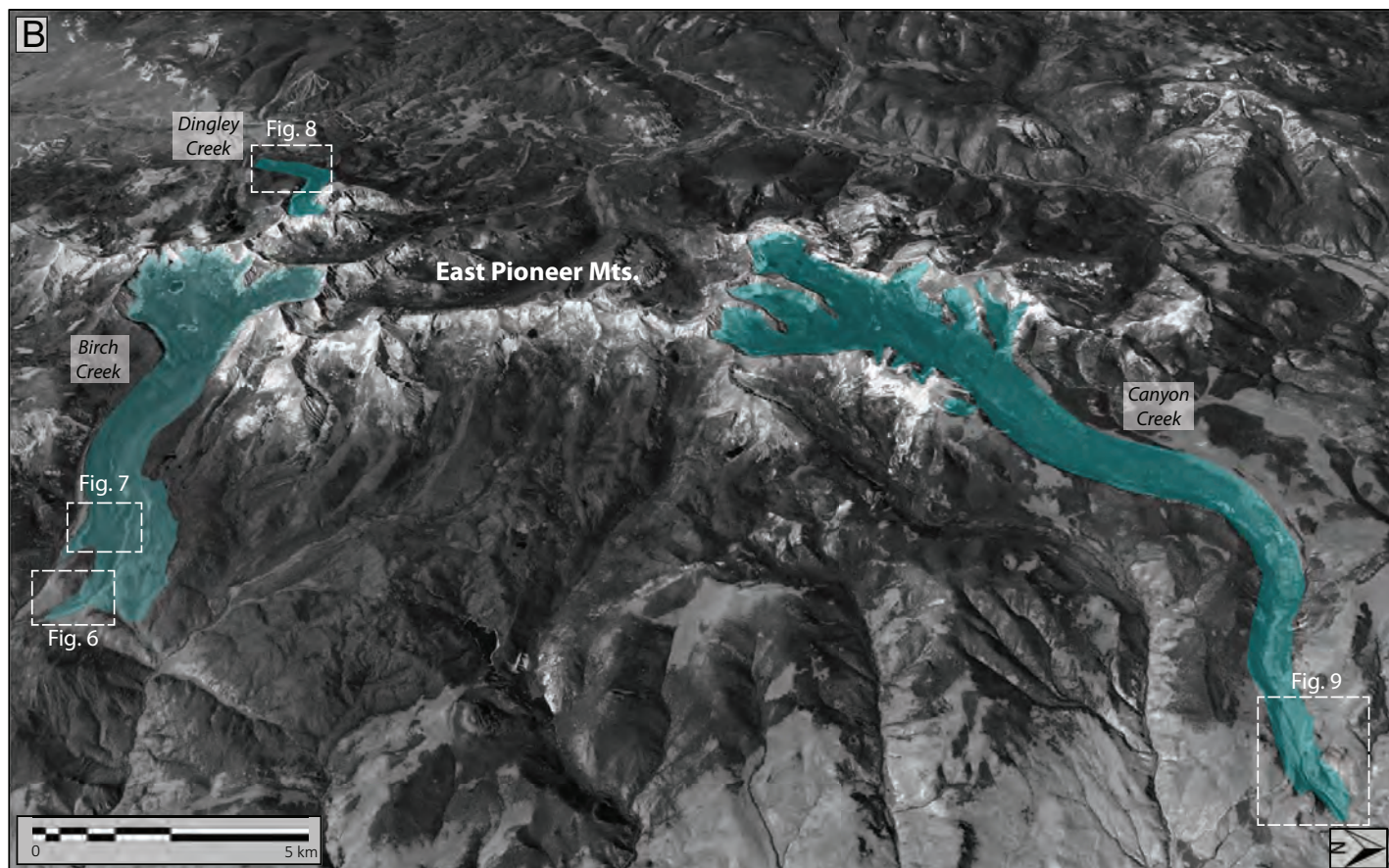


Figure 1B. Oblique aerial view of the East Pioneer Mountains looking west (from Google Earth). Erosional and depositional glacial landforms are visible along the north–south spine of the range. The blue outlines show the maximum glacial extent during the LGM of the three valleys sampled. Canyon Creek, Birch Creek, and Dingley Creek study areas outlined in dashed white. (Figure numbers refer to originals in Schoenemann and others (2023). From Schoenemann and others (2023).

the Pinedale glaciation [i.e., the last glaciation, as termed by Blackwelder (1915) based on the type locale in Wyoming), ice on the ranges merged in the Wise River/Grasshopper drainages to form a small ice cap (Smith, 2007). According to Smith (2007), the east Pioneer ice cap may have been thick enough to form outlet glaciers that spilled across drainage divides near the headwaters of Wise River, Birch Creek, and Grasshopper Creeks (see fig. 1 in Smith, 2007).

Geomorphic descriptions of glacial landforms by Alden (1953) provide the earliest documentation of ice extent in western Montana. Alden explored many localities in the Beaverhead, Madison, Tobacco Root, and Pioneer Mountains between 1927 and 1938, and described both younger and older moraines and outwash plains that extended down-valley of the glacial ice extent. In some valleys of the Pioneer Mountains, he identified older till with boulders containing large surface pits and moraines with subdued relief. Alden hypothesized that these landforms were formed during an earlier glaciation, and indeed, the differences in till characteristics are similar to those differences identified on confirmed Bull Lake and Pinedale moraines in the nearby West Yellowstone area by Pierce (2003). These early mapping efforts provided initial locations of

Quaternary glacial deposits and have since been improved upon by Pearson and Zen (1985) and Locke and Smith (2004), although neither separated moraines from multiple glaciations. More recent regional surficial geologic mapping efforts by McDonald and others (2012) at 1:100,000 scale include the northeast portion of the Pioneer Mountains, and by McDonald and Yakovlev (2019) at 1:24,000 scale of Birch Creek valley provide more detailed glacial outlines. However, even these most recent maps of the Pioneer Mountains have not delineated between Bull Lake and Pinedale till deposits, nor glacial outwash, with some boundaries mistaking hillslope colluvium with till.

STUDY AREA: PIONEER MOUNTAINS, MONTANA

The Pioneer Mountains are located in southwestern Montana in a region characterized as arid-steppe-cold summer (BSk) based on the Köppen–Geiger climate classification (Peel and others, 2007). When the LIS was present, between ~35 and ~13 ka (Dyke, 2004; Fullerton and others, 2004; Stokes and others, 2012), the Pioneer Mountains were located within the anti-cyclonic katabatic winds that descended off the ice sheet (Oster and others, 2015; Wong and others, 2016). The isolated dome-shaped range is



composed of Cretaceous (~75 Ma) granodiorite of the Pioneer Batholith, forming the north–south spine of the East Pioneer Mountains that rise to ~3,200 m asl and Paleozoic metasedimentary rocks forming the more subdued peaks of the West Pioneers (Hyndman and Thomas, 2020).

The Pioneer Mountains are located east of the Continental Divide, in the orographic shadow of the Beaverhead Mountains, which deplete moisture from the prevailing westerly winds. Today, the Pioneer Mountains receive the bulk of their precipitation in the winter and spring, with an average of 775 mm/yr of water equivalent precipitation and a mean summer (June, July, August) temperature of 11°C (1981–2021, Mule Creek station, 2530 m asl, NRCS). Major valleys along the eastern flanks of the Pioneer Mountains contained large valley glaciers (10–20 km long) during the Late Pleistocene. Although no ice remains, the semi-arid climate of southwest Montana has preserved many geomorphic features of the last glaciation (Smith, 2007).

Dingley Creek

Dingley Creek is located on the west side of the East Pioneer Mountains (figs. 1B, 2). The west-facing valley has well-preserved glacial trimlines up-valley and lateral moraines down-valley, and it extends ~5 km from the cirque headwall to the terminus. The elevation ranges from ~2,900 m to 2,150 m. The upper valley is characterized by one cirque flanked by steep peaks to the east and south and a col to the north. The quartz-rich granodiorite-dominated valley contains an abundance of large, semi-rounded boulders along the moraines. The glacier terminus is composed of a steep end-moraine incised by the creek, with extensive hummocky ground till, a well-preserved, right-lateral moraine crest, and a more subtle, broad, left-lateral ridge.

In the Dingley Creek valley, Schoenemann and others (2023) sampled a total of 11 boulders (figs. 3C, 4). They collected four samples from the terminus (DC02–04, 11), three samples from the left lateral moraine (DC12–14) that overlapped the bedrock valley wall, and four samples (DC05, 06, 08, and 10) from the well-defined right lateral moraine crest.

Dingley Creek ^{10}Be Ages

In the Dingley Creek terminus, 11 samples marking the outermost extent yield cosmogenic-exposure ages ranging from 14.4 ± 0.3 ka to 49.0 ± 0.6 ka, with a median age of 17.9 ka (figs. 3, 4). Of the 11 exposure ages in Dingley Creek valley, two were identified by CRONUS as outliers. The older outlier, DC10 (49.0 ± 0.6 ka), was a very large boulder (~2 m wide) located ice-proximal of the right-lateral moraine crest and appeared to be in place. Based on its geomorphic position, it was likely reworked, with the significantly older age indicative of inheritance of ^{10}Be from a prior period of exposure followed by minimal ero-

sion before deposition. The younger outlier, DC06 (14.4 ± 0.3 ka), was sampled from a lower bench, ice-distal of the steep-sided, right-lateral crest where the moraine bends sharply (figs. 3C, 4). Although DC06 was located ~10 m below the moraine ridge, it was sampled as a possible older Bull Lake boulder based on its extremely weathered and oxidized appearance. The younger age suggests the boulder was exhumed and rolled from above to the bench. The remaining nine cosmogenic-exposure ages have a mean age of 17.8 ± 0.4 ka (average, 1 SD; table 1, figs. 3, 4), suggesting that glacier ice in Dingley Creek valley persisted at the terminal moraine until this time.

SUMMARY

In the Pioneer Mountains, the maximum Pinedale glacier extent occurred around 18 ka, with all three glacial valleys overlapping in their retreat timing, suggesting a synchronous deglaciation throughout the mountain range. The specific timing of the onset of deglaciation is essential for understanding both the large-scale climatic drivers causing glacier retreat (e.g., insolation, greenhouse gases, albedo, and ocean-atmospheric circulation), as well as the regional factors (e.g., elevation/extent of LIS, locally sourced moisture from pro-glacial lakes, prevailing winds) that influence the spatial pattern of retreat across the northern Rocky Mountains and inland northwest.

Synchronous LGM Glacial Retreat among Pioneer Mountain Valleys

Compilation of the cosmogenic-exposure ages from all three glacial valleys indicates synchronous timing of maximum glacier extent in the Pioneer Mountains. The individual mean ages for each valley, from the northeast to southwest, are Canyon Creek (18.5 ± 0.4 ka, $n = 6$), Birch Creek (18.2 ± 0.5 ka, $n = 5$), and Dingley Creek (17.8 ± 0.4 ka, $n = 9$), yielding a range-wide average of 18.2 ± 0.9 ka ($n = 22$). The exposure ages from the different valleys are indistinguishable from one another in their maximum extent timing based on their 1SD age ranges. This overlap indicates that, despite the different orientations (northeast, east, and west-facing) and differing paleo-valley characteristics, the Pioneer Mountain glaciers responded synchronously to the large-scale climatic forces within the resolution of the ^{10}Be chronometer. The consistent timing of moraine abandonment among the three distinct glacier valleys provides confidence that ice retreat among the valleys reflects a broader, regional response, unlike the transient ages derived from the greater Yellowstone glacial system.

For more information on the glaciation of the Pioneers, see Schoenemann and others (2023).



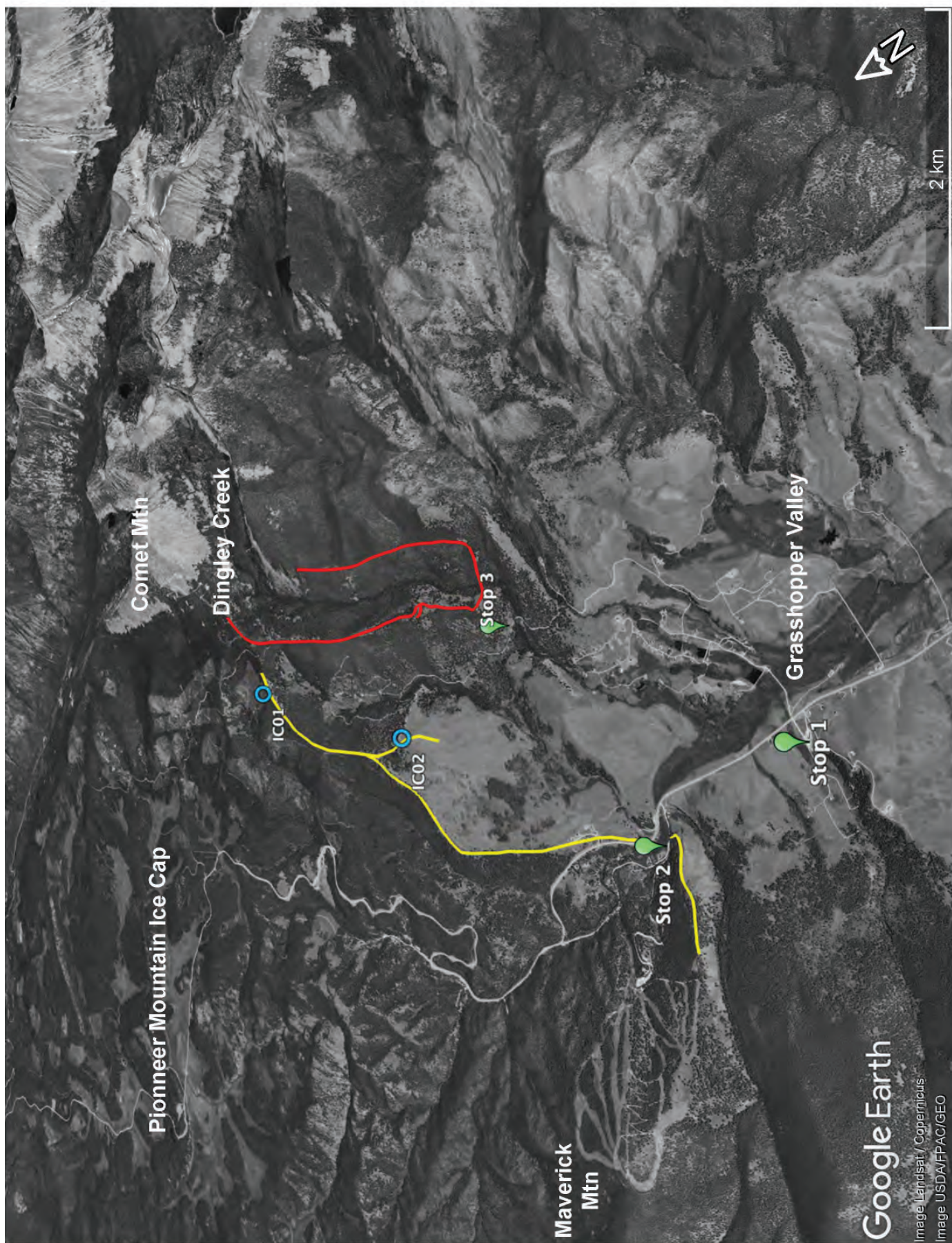


Figure 2. Grasshopper Valley and East Pioneers, with the locations of STOPS 1, 2, and 3. Aerial view from Google Earth Pro, 2013 timestamp.



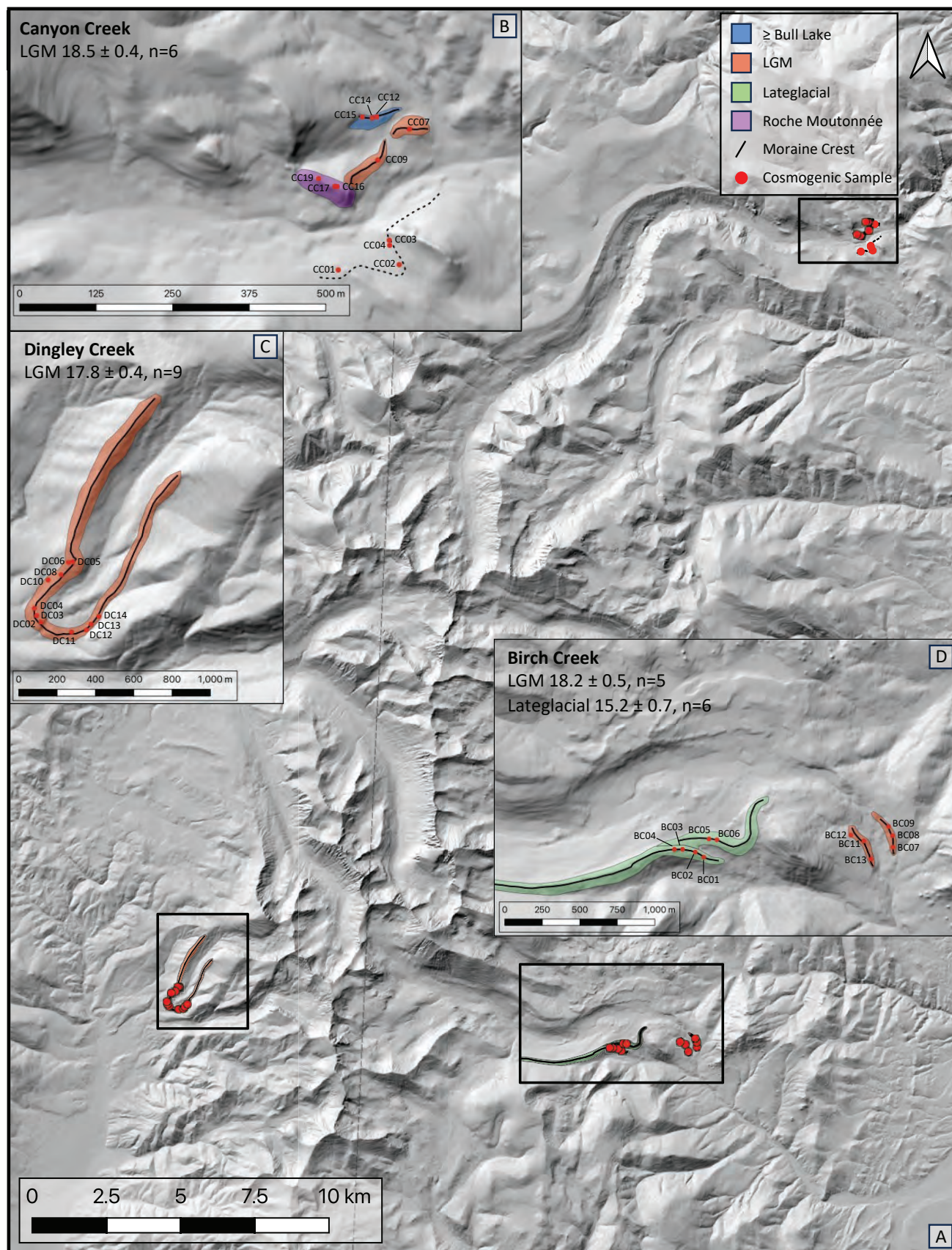


Figure 3. (A) Hillshade-derived image from 10 m lidar from the USGS map of the Pioneer Mountains showing mapped Last Glacial Maximum (LGM) and late glacial deposits. Cosmogenic samples are shown as red circles (refer to table 1 and fig. 4 for sample names and ages). Boxes include LGM/Pinedale averages and 1 SD deviation, excluding outliers. Insets (B, C, D) same as (A) but close-up view of Canyon Creek, Dingley Creek, and Birch Creek, respectively, showing the moraines and sampling sites, corresponding to the black boxes in (A). From Schoenemann and others (2023).



Table 1. Calculated exposure ages based on *in situ* ^{10}Be ($n = 35$) concentrations

Sample Name	Latitude (°N)	Longitude (°E)	Elevation (m a.s.l.)	^{10}Be Exposure Age (ka) ^a	^{10}Be Internal Uncertainty (ka) ^a	^{10}Be External Uncertainty (ka) ^a
Birch Creek Recessional (4 km up-valley)						
<i>BC-01</i>	<i>45.4230</i>	<i>-112.8877</i>	<i>2221</i>	<i>16.0</i>	<i>0.4</i>	<i>0.7</i>
<i>BC-02</i>	<i>45.4236</i>	<i>-112.8891</i>	<i>2266</i>	<i>16.0</i>	<i>0.4</i>	<i>0.7</i>
<i>BC-03</i>	<i>45.4238</i>	<i>-112.8913</i>	<i>2293</i>	<i>15.3</i>	<i>0.4</i>	<i>0.7</i>
<i>BC-04</i>	<i>45.4238</i>	<i>-112.8927</i>	<i>2319</i>	<i>14.3</i>	<i>0.5</i>	<i>0.7</i>
<i>BC-05</i>	<i>45.4252</i>	<i>-112.8869</i>	<i>2271</i>	<i>15.0</i>	<i>0.4</i>	<i>0.7</i>
<i>BC-06</i>	<i>45.4251</i>	<i>-112.8856</i>	<i>2267</i>	<i>14.5</i>	<i>0.4</i>	<i>0.7</i>
Recessional ($n = 6$) avg $\pm$ st dev				15.2	0.7	
Birch Creek Terminus						
BC-07	45.4252	-112.8556	2050	17.7	0.5	0.8
BC-08	45.4265	-112.8557	2075	18.3	0.5	0.9
BC-09	45.4276	-112.8564	2083	18.1	0.4	0.8
BC-11	45.4258	-112.8605	2057	17.9	0.3	0.8
BC-12	45.4264	-112.8628	2085	19.0	0.4	0.8
*BC-13	45.4236	-112.8592	2042	15.7	0.3	0.7
($n = 6$) avg $\pm$ st dev				17.8	1.1	
LGM (no outliers, $n = 5$) avg $\pm$ st dev				18.2	0.5	
Canyon Creek Terminus						
*CC-01	45.6681	-112.7993	1908	16.1	0.4	0.8
<i>CC-02</i>	<i>45.6686</i>	<i>-112.7942</i>	<i>1891</i>	<i>68.2</i>	<i>1.3</i>	<i>2.9</i>
CC-03	45.6700	-112.7952	1872	18.7	0.5	0.9
CC-04	45.6697	-112.7951	1873	18.0	0.3	0.8
CC-07	45.6766	-112.7939	1806	18.9	0.4	0.8
CC-09	45.6747	-112.7964	1830	18.8	0.4	0.8
CC-16	45.6730	-112.7997	1904	18.7	0.4	0.8
*CC-17	45.6730	-112.7999	1900	20.9	0.4	0.9
CC-19	45.6735	-112.8013	1899	18.2	0.4	0.8
($n = 8$) avg $\pm$ st dev				18.5	1.3	
LGM (no outliers, $n = 6$) avg $\pm$ st dev				18.5	0.4	
Canyon Creek pre-LGM Composite Moraine						
CC-13	45.6772	-112.7966	1808	142.5	1.5	5.8
CC-14	45.6772	-112.7970	1820	90.6	1.1	3.7
CC-15	45.6772	-112.7979	1831	222.5	2.1	9.1
Dingley Creek Terminus						
DC-02	45.4305	-113.0832	2221	17.2	0.3	0.7
DC-03	45.4311	-113.0838	2229	17.9	0.3	0.7
DC-04	45.4318	-113.0842	2230	18.1	0.4	0.8
DC-05	45.4363	-113.0793	2354	18.3	0.3	0.8
*DC-06	45.4362	-113.0799	2342	14.4	0.3	0.6
DC-08	45.4350	-113.0808	2311	17.5	0.3	0.7
<i>DC-10</i>	<i>45.4345</i>	<i>-113.0825</i>	<i>2279</i>	<i>49.0</i>	<i>0.6</i>	<i>2.0</i>
DC-11	45.4297	-113.0791	2241	17.8	0.4	0.8
DC-12	45.4299	-113.0771	2260	18.3	0.3	0.8
DC-13	45.4305	-113.0765	2285	17.4	0.6	0.9
DC-14	45.4313	-113.0755	2298	18.0	0.4	0.8
($n = 10$) avg $\pm$ st dev				17.5	1.2	
LGM (no outliers, $n = 9$) avg $\pm$ st dev				17.8	0.4	
Pioneer Mountains ALL Samples ($n = 35$) median Pioneer				18.0		
Mountains LGM/Pinedale ($n = 22$) avg $\pm$ st dev				18.2	0.9	

Note. Original data from Schoenemann and others (2023; see their table 3). Exposure ages were calculated using version 3.0 of the Online Exposure Age Calculator (Balco and others, 2008) and the Promontory Point production rate (Lifton and others, 2015) with LSDn scaling. Calculations assumed no erosion, elevation/pressure flag = standard atmosphere, and 07KNSTD.

Italics indicate samples not included in the LGM/Pinedale (30–16 ka) averages

*Possible outliers as identified by the OEAC landform outlier detection using the standard deviation, chi-squared and p-value of each valley.



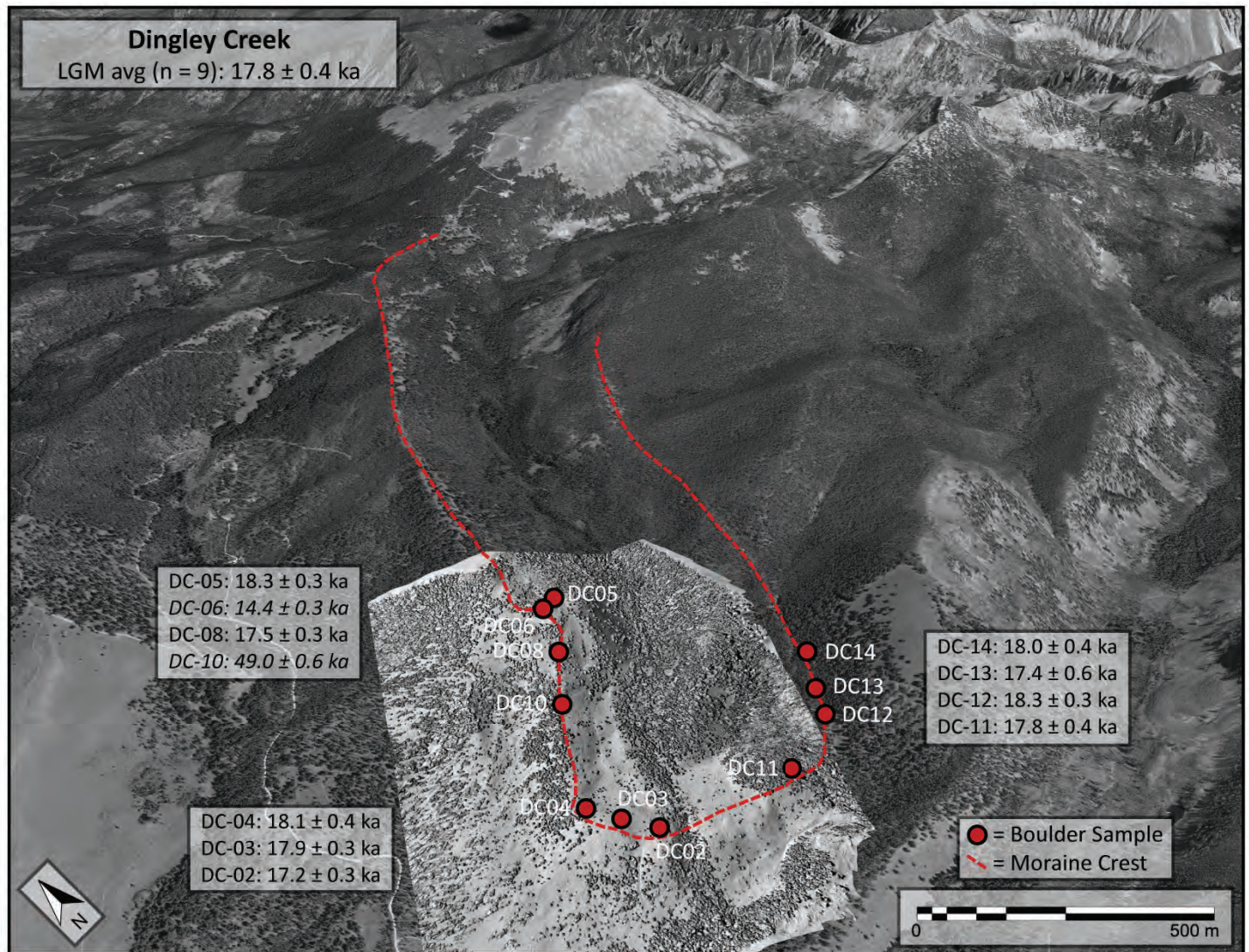


Figure 4. Aerial imagery (Google Earth) with shaded relief map (from drone survey/photogrammetry) of Dingley Creek terminus with boulder sample locations (red circles) and ages (boxes). Red lines indicate moraine crests—maximum ice extent. Italics indicate ages not included in the LGM (30–16 ka) average. From Schoenemann and others (2023).

FIELD TRIP

The field trip consists of three stops, with the final stop involving a half-day hike on uneven terrain with some moderate to steep sections along the terminus/moraine front. The round-trip distance is ~2 mi. Figure 2 shows the stop locations.

STOP 1

This field trip guide begins in the Grasshopper Valley, Montana, at the turnoff to the Grasshopper Inn on Shale Rd., which provides a scenic overlook of the East Pioneers, and the once glaciated alpine valleys that show well-preserved moraine complexes reaching the main valley floor. The Grasshopper valley is filled with glacial till deposits comprised mostly of granitic boulders, along with remnants of broad recessional moraines that stem from the edges of the valley. A prominent moraine crest trends southwest from Comet Mountain, indicating the terminus of the Pioneer Ice Cap, and

neighboring it to the south is a smaller but well-defined right-lateral moraine crest bounding Dingley Creek. The dominant mode of boulder ages measured in the Dingley Creek valley corresponds to the late Pinedale (ca. 18 ka), but several boulders date to ca. 50 ka, with a few younger outliers as well.

Directions

From Dillon, MT, head west on US Hwy 278 until you reach the turnoff for the Pioneer Mountain Scenic Byway (~29 mi, signs for Polaris, MT and Maverick Mountain Ski Area to the east). Head north on the Scenic Byway for 9.5 mi, then turn west on Shale Rd., parking at the Grasshopper Inn. Look northeast and east for glacial landforms, including the moraine complex exiting Sawtooth Canyon, the distinct lateral moraines of Dingley Creek, and the massive ridge former of the Pioneer Ice Cap.



STOP 2

(Optional—provides close-up of well-preserved Pioneer Ice Cap terminal moraine)

At the turn off to Maverick Mountain Ski Area, the Pioneer Mountain Scenic Byway climbs steeply over one of the well-preserved terminal moraines from glacial ice that expanded southward into the Grasshopper valley, originating in the high plateau between the West and East Pioneers, forming a modest ice cap (~250–300 m thick; Smith, 2007). A single boulder (IC02) located midway up the terminal moraine of the ice cap provides a ^{10}Be cosmogenic age of 33.6 ± 1.4 ka. A lack of reliable boulders along the ice cap edge made it difficult to constrain the timing of its maximum extent, but the well-formed lateral moraines around Dingley Creek provide excellent age constraints in the neighboring drainage.

Directions

Heading north on the Pioneer Mountain Scenic Byway, travel 0.6 mi to Ski Hill Rd., turning left (west) and park at the pullout at the intersection of Ski Hill Rd. and Old Canyon Rd. Walk back towards the Scenic Byway, and cross the bridge to get a better view of the steep-faced moraine and granitic boulders.

STOP 3 (Primary Stop, half-day hike, round trip ~2 mi, on uneven terrain/game trails)

The terminus of the Dingley Creek paleoglacier begins here. The steep and well-preserved terminus of Dingley Creek forms a classic horseshoe shape, with a deep notch carved out by glacial meltwaters through the center of the terminus. ^{10}Be cosmogenic ages from the lateral and terminal moraine ridges date to ~17.8 ka ($n = 9$, table 1), and the lack of obvious recessional moraines up-valley suggests consistent glacial retreat into the cirque basin by 14 to 12 ka (GSA presentation, Schoenemann and others).

Directions

Return to the Pioneer Mountain Scenic Byway, heading south towards the Grasshopper Inn. If you skipped STOP 2, leave the Grasshopper Inn and cross over the Byway onto Clark Creek Rd. Follow Clark Creek Rd. and signs to Sawtooth Lake Trailhead for 2.1 mi, turn left (north) at an unmarked junction on Wellman Creek Rd. If you reach the Sawtooth Lake Trailhead, you've gone too far. Travel north on Wellman Creek Rd. for 0.5 mi, crossing Dingley Creek, and pull off on a logging road to the right. Drive along the logging road for ~0.2 mi until reaching a clear meadow to park.

From here, the main hiking portion of the field trip begins. Head southeast along the logging road for ~800 ft, where the road cuts left. Head east, up the steep logging road, hiking up through the glacial outwash

towards the middle of the terminal moraine crest (towards the notch). Once in the notch, head north following game trails along the moraine crest. Here you will find the first dated boulder, DC02, dated to ~17.2 ka (see table 1). Head north along the terminus, passing boulders DC03, DC04, moving towards the intersection of the right lateral moraine, which curves towards the northeast up-valley. There are numerous granitic boulders along the lateral moraine crest, including DC05 through DC10 (see table 1 for all exposure ages). The highest dated boulder, DC05, is about 0.7 mi up from DC02, located at a jog in the moraine crest (45.4363 N, 113.0793 W, 2354 m). This general location provides some nice views both up-valley towards the cirque and down-valley at the left-lateral moraine that hugs the base of the bedrock ridge. Here, you can either head further up-valley along the moraine crest or return the way you came to the parking area.

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FIELD GUIDE TO THE NEOGENE SIXMILE CREEK FORMATION IN THE UPPER RUBY VALLEY, MONTANA

Daniel T. Brennan

Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

INTRODUCTION

In southwestern Montana, the relationship between the Yellowstone hotspot system, extensional intermontane basin evolution, and development of the modern Snake and Missouri drainage networks is debated (see Staisch and others, 2022; Thomas and Sears, 2020). The Sixmile Creek Formation of southwestern Montana, as defined by Kuenzi and Fields (1971), is the Neogene strata above the middle Miocene (Hemingfordian) unconformity and consists of a heterogenous assemblage of fluvial and alluvial conglomerates, reworked tuffaceous sediments, and volcanic flows

that constitutes much of the intermontane valley-fill deposits in southwestern Montana (fig. 1). Thus, understanding the depositional architecture, age, and tectonostratigraphic framework of the Sixmile Creek Formation is key to deciphering the interrelationships between Neogene tectonism, drainage network evolution, and the Yellowstone hotspot. Over the 2020–2024 field seasons, Montana Bureau of Mines and Geology researchers conducted 1:24,000-scale geologic mapping and geochronological sampling in the upper Ruby Valley, including five 7.5' quadrangles (Belmont Park Ranch, Brennan and others, 2024; Red Canyon, Mossolf and Sears, 2024; Home Park Ranch, Brennan, in press;

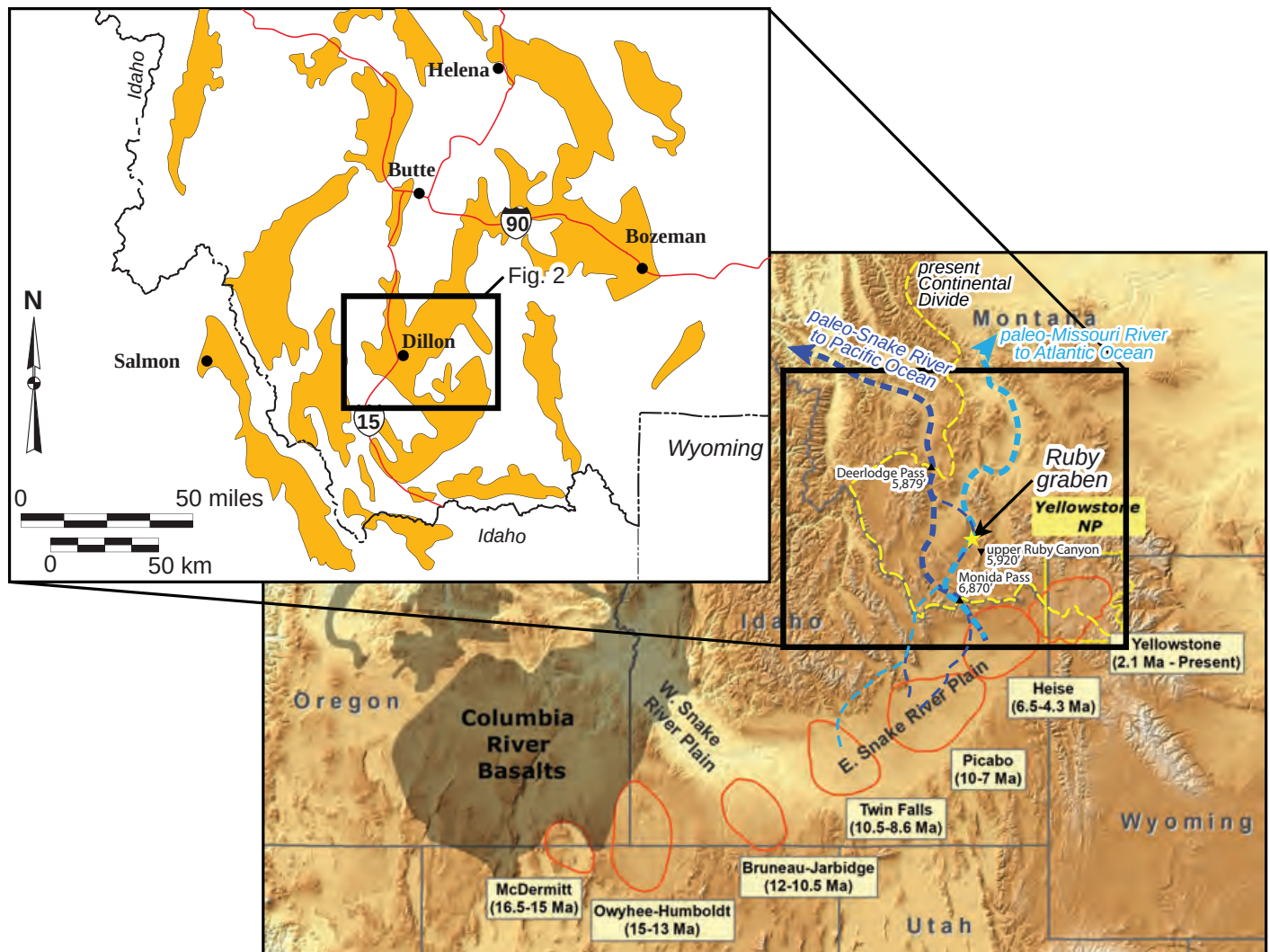


Figure 1. Regional map of the Yellowstone hotspot track and the time progression of associated volcanic fields (modified from Barry and others, 2013). In southwest Montana, the Ruby graben is starred and the recently proposed drainage networks for southwestern Montana are shown as dashed blue lines (Staisch and others, 2022; Thomas and Sears, 2020). Inset map shows the Tertiary basins of southwest Montana (in orange) and the extent of the simplified geologic map in figure 2.

Metzel Ranch and Ruby Dam 7.5' quadrangles, Brennan, in prep.) for compilation into the Dillon 1:100,000-scale 30' x 60' quadrangle (Mosolf and others, in prep.) and to explore these foundational geologic relationships. The upper Ruby graben is an approximately 500 km² basin and is one of the most proximal intermontane basins to the current Yellowstone hotspot to contain notable exposures of Neogene Sixmile Creek strata, including several key member type-localities, and important volcanic-age controls.

REGIONAL GEOLOGIC BACKGROUND

The upper Ruby River Valley is flanked by predominantly northeast–southwest-trending mountain ranges, the Ruby Range to the northwest and the Snowcrest–Greenhorn Ranges to the southeast (fig. 2). These ranges are cored by Archean and Paleoproterozoic metamorphic rocks containing a principally northeast–southwest-trending fabric likely associated with late Paleoproterozoic deformation during the Big Sky Orogeny (Harms and Baldwin, 2023; Berg, 1979; Tysdal, 1976). Metamorphic basement rocks are overlain by Paleozoic to early Paleocene sedimentary rocks. Both Precambrian metamorphic and Paleozoic–Mesozoic sedimentary rocks are deformed by overlapping “thick- and thin-skinned” thrust structures of the Blacktail–Snowcrest uplift associated with crustal shortening and related exhumation from ca. 100–64 Ma during the Cordilleran Orogeny (Parker and Pearson, 2023; Kaempfer and others, 2025). Cretaceous to early Paleocene igneous rocks are rare within the ranges directly flanking the upper Ruby Valley, but igneous activity of this age is more widespread in the western portion of the Dillon 100:000-scale map region, including rocks of the ca. 80–64 Ma Pioneer Batholith, ca. 78–72 Ma Bannack Volcanic Group, and ca. 50–0 Ma Dillon Volcanic Group (Mosolf, 2021, 2024; Scarberry and others, 2021).

Eocene and younger extensional structures crosscut and locally reactivate older shortening structures, and a younger period of post-Cordilleran orogeny igneous activity and sedimentation is recorded in the Eocene to early Oligocene Renova Formation and mid-Miocene to Pliocene Sixmile Creek Formation. Within the upper Ruby Valley, rhyolitic flows dated at ca. 34 Ma are interbedded with volcanic–lithic sandstones, and finer-grained deposits of the Eocene to earliest Miocene Renova Formation, which are unconformably overlain by mostly coarser-grained deposits of the Miocene to Pliocene Sixmile Creek Formation. In the Ruby Valley, the Renova Formation has been subdivided into three formal mappable members: the (a) Climbing Arrow member, (b) Dunbar Creek member, and (c) Passamari members (fig. 3; Monroe, 1976, 1981). A regionally extensive middle Miocene, Hemingfordian-age unconformity separates the Renova Formation from the younger Sixmile Creek Formation, the latter formation being the focus of this field guide. Previous workers have indicated that this unconformity is regionally extensive and coeval with out-break of the Yellowstone hotspot and the onset of Basin and Range extension (Hanneman and Wideman, 1991; Fritz and others, 2007; Gerritsen and others, 2025). In the portions of the upper Ruby Valley that encompass this field guide, Tertiary valley-fill deposits of the Renova and Sixmile Creek Formations are generally tilted east–northeast into the Snowcrest–Greenhorn Range-front fault, indicating this range-front fault is the main structural control on the valley geometry, except where Tertiary units are locally tilted by young northwest-trending faults like in the vicinity of Timber Hill. Consequently, for the remainder of this field guide, the term Ruby graben will be used interchangeably with upper Ruby Valley (fig. 1).

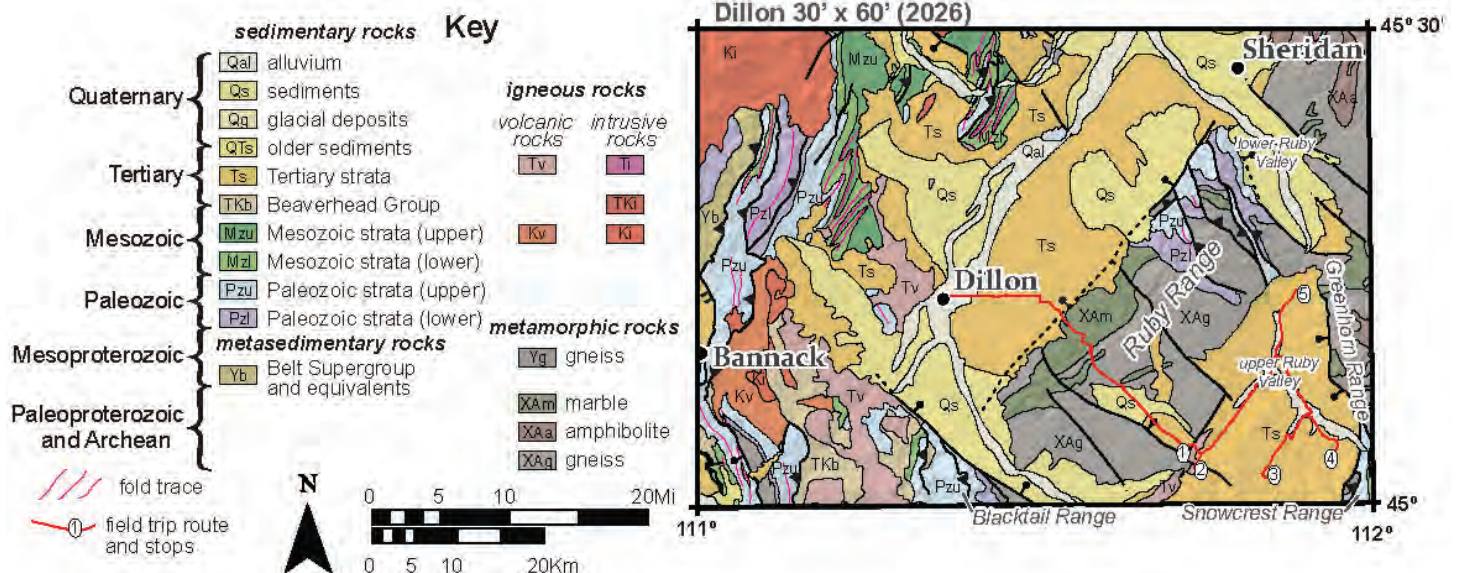


Figure 2. A simplified geologic map of the Dillon 30' x 60' sheet including the numbered locations of field trip stops (adapted from Ruppel and others, 1993; Mosolf and others, in prep.).

Monroe (1976) was the first to map the Ruby graben in detail and delineated separate, informal members of the Sixmile Creek Formation. He subdivided the formation into three primarily lithologically defined members: the metamorphic fanglomerate member; feldspathic sandstone member; and quartzite pebble conglomerate member. Fritz and Sears (1993) and Fritz and others (2007) redefined these lithological members into stratigraphic members they called the Sweetwater Creek, Anderson Ranch, and Big Hole River members (fig. 3). Fritz and others (2007) report the observations below from the Ruby graben.

The Sweetwater is defined as the basal member, with a type locality along the canyon of Sweetwater Creek in the southern Ruby Range (fig. 4, STOP 2). Here, the Sweetwater member is described as a heterogenous mix of feldspathic, matrix-supported, poorly sorted, poorly rounded, locally derived metamorphic and igneous rock. Fritz and others (2007) report the Anderson Ranch member as overlying the Sweetwater member. The Anderson Ranch member is described as prominent ash beds interlayered

with siltstone, sandstone, and channels of well-rounded pebble conglomerate, with a type-locality defined along the East Fork of Blacktail Creek in the Price Creek NE 7.5' quad. This region has only been mapped at 1:100,000 scale in the Lima 30' x 60' quad (Lonn and others, 2000). Finally, Fritz and others (2007) describe the Big Hole River member as overlying the Anderson Ranch member, and composed of a thick sequence of roundstone conglomerate, siltstone and sandstone, named for excellent exposures located along the Big Hole River, southwest of Twin Bridges. Although Fritz and others' (2007) exact type locality is not clear, recent geologic mapping along the Big Hole River southwest of Twin Bridges has identified ca. 48–45 Ma tuffs within coarse rounded conglomerate channels, which correlates some exposures previously considered Sixmile Creek Formation (Yakovlev, 2022) with the Renova Formation (Parker and others, in press).

Fritz and others (2007) also note that the Sixmile Creek Formation contains several volcanic flows likely related to the ca. 6.5–4.3 Ma Heise volcanic field (fig. 1). In the Ruby

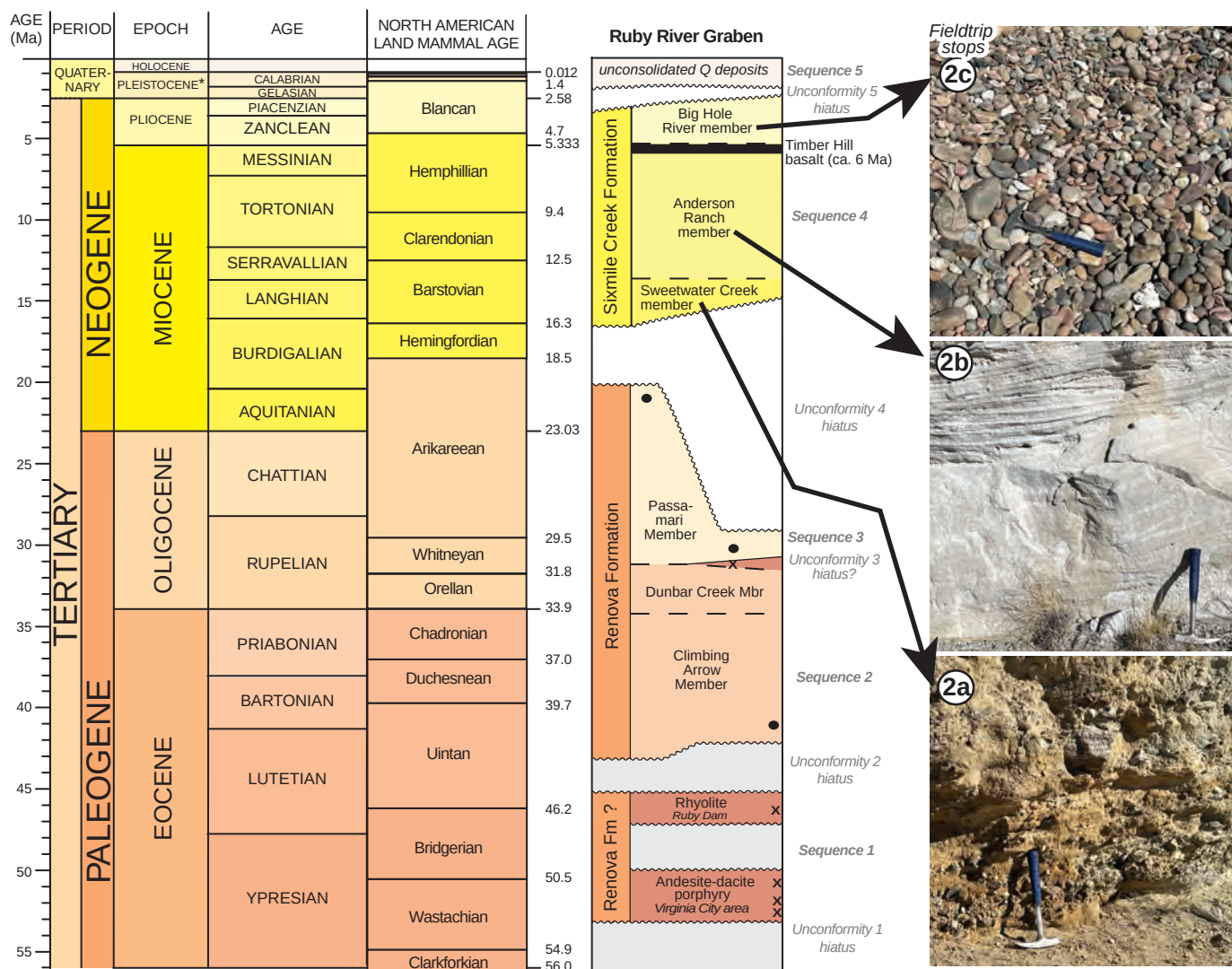


Figure 3. Generalized stratigraphy of Renova and Sixmile Creek Formations (adapted from Vuke, 2020 and Thomas and Sears, 2020) including field photos and generalized stratigraphic position of the lithologies viewed at STOP 2.



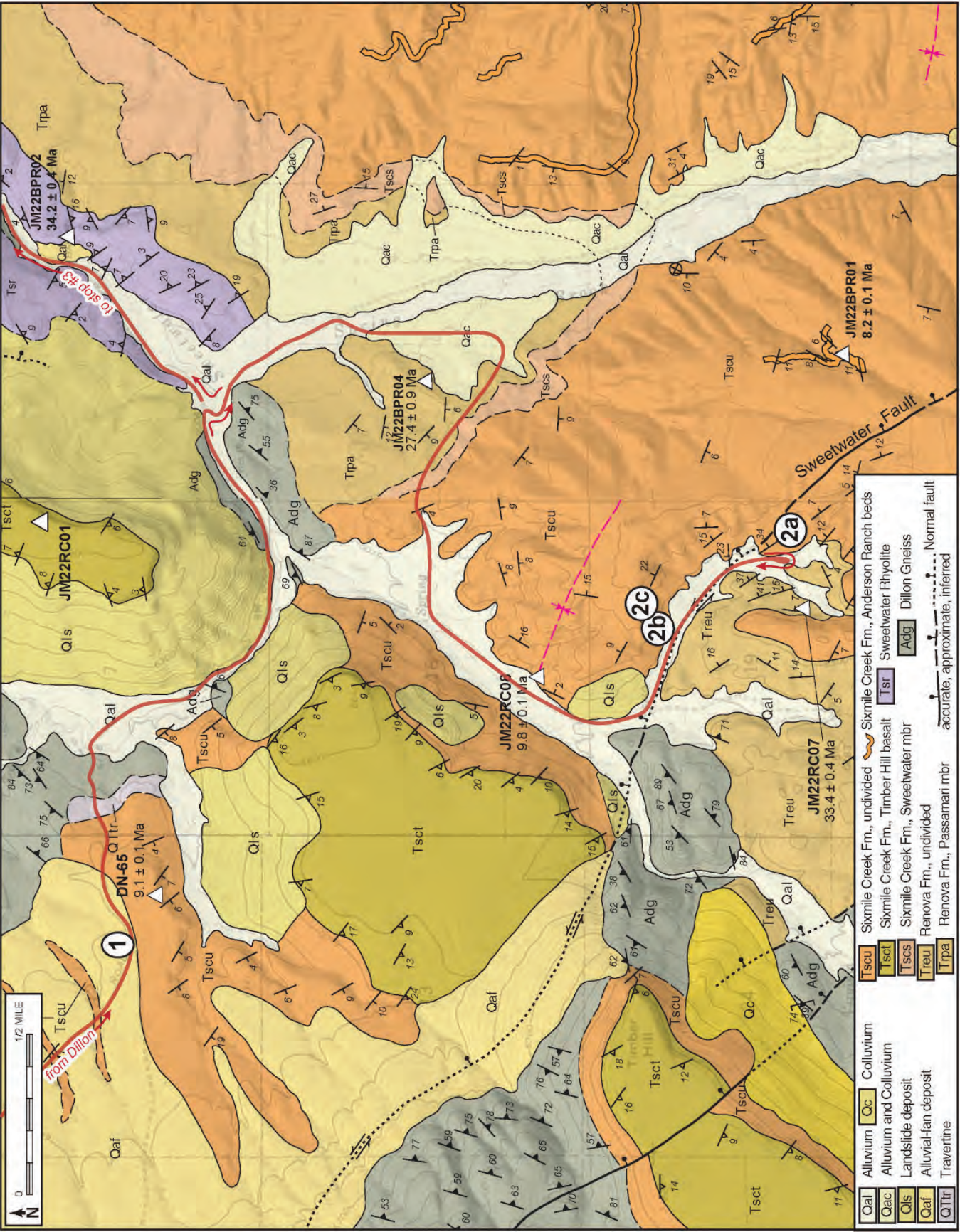


Figure 4. 1:24,000-scale geologic map of the Timber Hill area including field trip STOPS 1 and 2 (adapted from Mosolf and Sears, 2024; Brennan and others, 2024).



graben, this includes the ca. 6 Ma Timber Hill basalt. In the Ruby Range and Ruby graben the 6 Ma Timber Hill basalt can be traced along a southwest to northeast flowpath and is offset by a series of northwest-trending normal faults. Fitz and others (2007) describe the Timber Hill basalt in the vicinity of Timber Hill, as overlain by Sixmile Creek Formation gravel (presumably of the Big Hole River member) that is capped by a ca. 2.0 Ma tuff (the Robb Creek tuff) correlative to the Huckleberry Ridge tuff of the Yellowstone volcanic field.

In more recent review work on the Sixmile Creek Formation, Thomas and Sears (2020) emphasize the temporal and spatial overlap of these Sixmile Creek members, and note the difficulty in mapping these members at regional scale. This difficulty has resulted in a poorly defined internal stratigraphy, and a somewhat debated depositional and tectonic history for the Sixmile Creek Formation. In this field guide, I advocate for a revised facies-based approach to mapping the internal stratigraphy of the Sixmile Creek Formation and share some new zircon U-Pb geochronological constraints that help refine the Sixmile Creek Formation's depositional age, provenance, and tectonic history in the Ruby graben. All geologic relationships and geochronology results discussed in the following field trip stops are the product of 1:24,000-scale geologic mapping of the Red Canyon (Mosolf and Sears, 2024), Belmont Park Ranch (Brennan and others, 2024), Home Park Ranch (Brennan, *in press*), Metzel Ranch, and Ruby Dam 7.5' quadrangles (Brennan and others, *in prep.*). Geochronology results are discussed in the corresponding maps. Plates with raw datasets are available for download in Mosolf and others (2023), Brennan and others (2025), and Brennan and others (*in prep.*).

ROAD LOG

Road log mileage starts at the intersection of North Montana and East Virginia streets in Dillon, Montana. Set odometer to zero and begin travel east/southeast on East Virginia St.

0.2 mi Keep slightly left as East Virginia St. changes into Kentucky Ave, and continue east as Kentucky Ave changes to Sweetwater Road. As you leave Dillon you drive across low-relief surfaces of Renova and Sixmile Creek strata, with some overlying Quaternary alluvium, colluvium, and fan deposits (McDonald, 2022).

9.5 mi The road crosses the Ruby Range western border fault, which is concealed beneath Sixmile Creek strata dated at ca. 10.5–10.7 Ma (McDonald and Mosolf, 2023). As the topography increases, you are heading into the Ruby Range. The Ruby Range consists primarily of Archean and Paleoproterozoic metamorphic rocks that generally exhibit northeast-trending metamorphic fabrics.

11.5 mi Sweetwater Road crosses the Regal Mine, which is an active Talc ($\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$) mine that is hosted within a synform of the Paleoproterozoic Christensen Ranch Metasedimentary Suite. Continue east/southeast along Sweetwater Road to STOP 1, at ~23.2 mi. From the Regal Mine to STOP 1, Sweetwater Road essentially parallels the Sweetwater fault, a northwest-trending normal fault that is mostly concealed by Quaternary alluvial fan deposits. Map relations indicate the fault offsets underlying Sixmile Creek Formation strata and Paleoproterozoic to Archean metamorphic basement rocks of the Christensen Ranch Metasedimentary Suite, Dillon Gneiss, and Elk Gulch Suite (Mosolf and Sears, 2024).

STOP 1—Timber Hill Basalt Offset by the Sweetwater Fault

23.2 mi
(45.06523°N, -112.28042°W)

STOP 1 is an overview stop, and STOP 2 is a stratigraphic transverse. Both stops are located along the border of the Red Canyon and Home Park Ranch 7.5' quadrangles mapped by Mosolf and Sears (2024) and Brennan and others (2024). STOP 1 is located on top of Quaternary alluvial fan deposits (fig. 4). These deposits are forming along the relief generated by movement on the Sweetwater fault. Off to the southeast, about 1 mi away, is a segmented mesa made by the resistant nature of the Timber Hill basalt, a ca. 6.0 Ma basalt flow named after the Timber covered mesa that is visible in the distance to the south. Just to the north of Timber Hill, the mesa is downdropped approximately 700 ft (~213 m), reflecting movement along the Sweetwater fault. Approximately 213 m of throw on the Sweetwater fault over the past 6 Myr indicates an estimated slip rate of ~3.6 cm/kyr.

The smaller mesas visible ~1.5 mi to the northeast are the northernmost traced extent of the Timber Hill basalt flow. To the southwest, “shoestring” outcrops of Timber Hill basalt can be traced approximately 30 mi to near Lima, Montana (Lonn and others, 2000). Based on this outcrop pattern, geochemistry, and geochronology similarities, prior workers (Fritz and Sears, 1993; Sears and others, 2009) propose the Timber Hill basalt flow originated from the Heise volcanic field on the Snake River Plain over 50 mi to the south–southwest (fig. 1). This correlation would require that the lava flowed down a paleo-channel that traversed the present sites of the Tendoy, Blacktail, and Ruby Mountains. Both the Tendoy and Blacktail Ranges are bounded by northwest-trending, down-to-the-northeast Quaternary normal faults that parallel the Sweetwater fault.

MBMG geologic mapping and geochronology (Mosolf and Sears, 2024; Brennan and others, 2024; fig.



3) indicates that in the Red Canyon and Belmont Park Ranch 7.5' quadrangles, the Timber Hill basalt primarily overlies fluvial gravels with interbedded, reworked, tuffaceous sandstone deposits of the Sixmile Creek Formation dated to ca. 16–7.5 Ma, and forms a mesa, notably with no Sixmile Creek strata preserved stratigraphically above it (fig. 4). This mapping relationship indicates that the sedimentation (and preservation of intervening volcanic units) from ca. 16 to 6 Ma was generally continuous within the upper Ruby graben. Sometime after eruption of the ca. 6 Ma Timber Hill basalt, the Ruby graben underwent exhumation, resulting in the inverted topographic relationship visible in the Timber Hill basalt mesa. In back-tilted (southwest-tilted) blocks of the Timber Hill basalt, to the south of this study area, some undated fluvial gravels are preserved (Rob Thomas, oral commun.). These gravels suggest that limited preservation of sedimentation may have occurred post-Timber Hill basalt.

From STOP 1, continue east on Sweetwater Road. As you pass through Sweetwater Canyon, where Sweetwater Creek turns ~90° from flowing southeast to northeast. Note the landslide complexes that surround the Timber Hill basalt mesa (fig. 4).

25.3 mi Turn right (south) onto Springbrook Creek Road, just past the Fries Place cabin ruins.

25.8 mi Stay slightly right (west) onto the unnamed BLM road. This road will head west, through a gate, and turn abruptly south.

28.1 mi Stay right and proceed to mile 28.8 to park for STOP 2A (fig. 4).

STOP 2—The Three Members of the Sixmile Creek Formation

This stop will consist of two short hiking traverses to view the three members of the Sixmile Creek Formation in relative stratigraphic order (see figs. 3, 4).

STOP 2A—Sweetwater Member of the Sixmile Creek Formation

28.8 mi

Park (45.0356°N, -112.2549°W),

Hike (45.0357°N, -112.2535°W)

On the basis of lithology, stratigraphic and geographic distribution, and depositional environment, Monroe (1976) informally defined three local members of the Sixmile Creek Formation in the Ruby graben. The stratigraphically lowest and oldest member he defined, the metamorphic fanglomerate member, later renamed by Fritz and Sears (1993) and Fritz and others (2007) as the Sweetwater Creek Member, is described as:

“A thick sequence of fanglomerate which has as its dominant constituent subangular pebbles, cobbles and boulders of pre-Belt gneiss and schist is designated as the metamorphic fanglomerate member. It overlies the Renova Formation with angular unconformity... laterally the metamorphic fanglomerate member grades into the feldspathic sandstone member and the lower part of the quartzite pebble conglomerate member. Lithologically, the member is very distinct, and can be differentiated from any other stratigraphic unit in the basin. Typically, it is composed of conglomerate or sandstone with boulders up to six feet in diameter floating in a sand matrix, and conglomerate filled channels are locally common. Other lithologic types in the member include siltstone, limestone, and tuff in subordinate amounts.” (Monroe, 1976, p. 70–71).

Lithologies consistent with this member are observable at STOP 2A, including coarse, angular, primarily metamorphic-cobble channels, and immature lithic sandstones (fig. 4).

Return to the cars, turn around and drive back the way you came to mile 29.2; park along road. From here, complete a short hike up the ridge to the south to STOP 2B and then continue another ~50 m (164 ft) up-section to STOP 2C. Note that the Sweetwater fault likely occurs in the drainage between STOPS 2A and 2B/2C and is concealed beneath Quaternary alluvium.

STOP 2B—Anderson Ranch Member of Sixmile Creek Formation

29.2 mi

Park (45.0405°N, -112.2598°W),

Short hike to (45.0415°N, -112.2592°W)

Fritz and others (2007) describe the Sweetwater Creek member in the upper Ruby Valley as overlain by approximately 50 to 100 m (164 to 328 ft) of prominent ash beds and interlayered siltstone, sandstone, and conglomerate, which they named the Anderson Ranch member. The ash beds form spectacular light gray cliffs that present a striking contrast to the enclosing members of the Sixmile Creek Formation. Numerous individual ash beds are exposed at the type locality at the old Anderson Ranch, along the East Fork of Blacktail Deer Creek. The thickest is up to 30 m (98 ft) thick. The ash beds consist of crisply angular glass shards, rounded pumice fragments up to a few centimeters in diameter, pumice rafts up to several centimeters long, and relatively minor, well-rounded pebbles of pre-Sixmile Creek rocks.



STOP 2C—Big Hole River Member of Sixmile Creek Formation

29.2 mi

Short hike to (45.0415°N, -112.2587°W)

Fritz and others (2007, p. 104) also write:

“The Anderson Ranch Member is overlain by a thick sequence of roundstone conglomerate, siltstone, and sandstone that we name the Big Hole River member, from excellent exposures along the Big Hole River southwest of Twin Bridges. This member is commonly poorly consolidated, degrades into rounded hills littered with boulders, cobbles, and pebbles, and is sparsely vegetated with bunchgrass and small cactus. The Big Hole River member is 200–300 m thick and occupies broad channels that are faulted and tilted....Outcrops that reveal the primary bedding characteristics of the Big Hole River member are rare but show that the conglomerates are channeled, are clast-supported and commonly have well imbricated pebbles and cobbles.”

Fritz and others (2007) report that many of the roundstone cobbles and boulders in the Big Hole River member are distinct and can be matched with bedrock in southwestern Montana and central Idaho. These varieties include: (1) red and purple, coarse, feldspathic, jasper-bearing, cross-bedded quartzites and finer-grained red clastic rock they correlate with the Proterozoic Belt Supergroup and Lemhi Group, (2) brown and white quartz arenite they correlate with the Pennsylvanian Quadrant Sandstone, and (3) other common type lithologies they correlate to Cretaceous granodiorite, Archean foliated metamorphic rocks, dark red conglomerate of the Cretaceous Beaverhead Group, Cretaceous hornfels, fossiliferous Paleozoic limestone, Paleozoic chert, and Eocene volcanic rocks. One particularly distinctive pebble type they describe is a black chert with sets of folded, cross-cutting, white quartz veins that resemble the greenschist-facies, Devonian-aged Milligan Formation of central Idaho.

These last two members of the Sixmile Creek Formation (the Anderson Ranch and Big Hole members) may be partially correlative to the Feldspathic sandstone and Quartzite pebble conglomerate members originally defined in the upper Ruby Valley by Monroe (1976). Monroe described the Feldspathic sandstone member as the portion of the Sixmile Creek Formation dominated by medium- to coarse-grained feldspathic sandstone, abundant conglomerate, and sparse tuff, which laterally grades into the Metamorphic fanglomerate (Sweetwater member equivalent) member. Monroe (1976) goes on to describe that both the Feldspathic sandstone and Metamorphic fanglomerate members may grade partially into the lower part of the Quartzite pebble

conglomerate. Monroe describes the Quartzite pebble conglomerate member as primarily consisting of well-rounded quartzite clasts with a significant component of light gray tuff.

Return to the cars and drive ~3.1 mi back to the Fries place cabin ruins.

32.3 mi At Fries place turn right (northeast) back onto Sweetwater Creek Road. From ~32.7 to 34.7 mi, you are driving through a canyon carved by Sweetwater Creek through a rhyolite flow, dated to ca. 34 Ma, with intercalated thin lithic sandstones. These units generally unconformably overlie Archean Dillon Gneiss basement (Brennan and others, 2024).

35.8 mi Cross another northwest–southeast-trending normal fault (the Stone Creek Fault), which is easy to identify where it downdrops Tertiary units against Archean basement but is difficult to trace southeast into the valley where, if present, it presumably offsets Sixmile Creek strata against Sixmile Creek strata. From ~39.3 to 41.3 mi, the cliffs to the west/left are composed of mostly angular cobble to boulder conglomerates with proximally sourced lithologies consistent with the Sweetwater Creek member of the Sixmile Creek Formation. Some clasts here are matrix supported, suggesting debris flow deposition. STOP 5 will contain exposures of similar lithofacies.

The final three stops (STOPS 3–5) are to view the variability of Sixmile Creek facies within the Ruby graben at a similar time interval, ca. 10 Ma.

40.5 mi Cross the Ruby River

40.9 mi Turn right and head southeast on upper Ruby Road. From ~41 to 47.4 mi, the visible stratigraphy on either side of the alluvial upper Ruby Valley is Sixmile Creek Formation that generally dips shallowly (~5 to 15°) to the northeast toward the Greenhorn Range front. A single tephra sample (JM20HPR01) collected along the road, at ~47.9 mi, returned a zircon U-Pb age of 9.3 ± 0.1 Ma (Mosolf and Kylander-Clark, 2023).

47.4 mi Turn right (southwest) onto Ledford Creek Road. Continue on Ledford Creek Road for 3.2 mi to the junction with Robb Creek Road.

50.6 mi Turn right (west) onto Robb Creek Road, and drive 3.4 mi to parking for STOP 3 (fig. 5).

STOP 3—2.0 Ma Robb Creek (Huckleberry Ridge) Tuff

54.0 mi

Park (45.0128°N, -112.1393°W),
Hike (45.0107°N, -112.1275°W)

STOP 3 is an ~1.8-mi round trip hike to the base of another prominent mesa in the upper Ruby Valley (fig. 5)



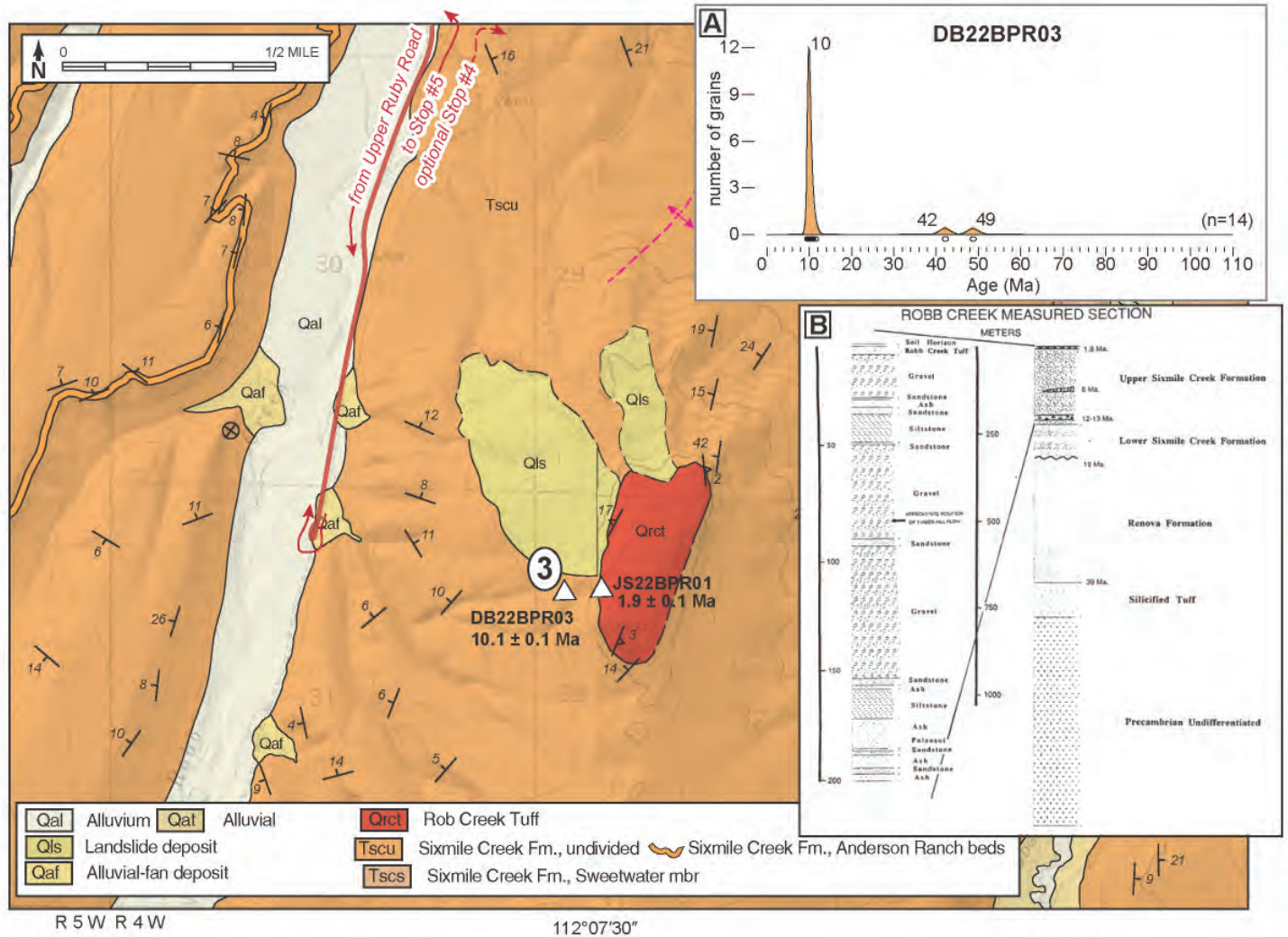


Figure 5. 1:24,000-scale geologic map of the Robb Creek Tuff area including field trip STOP 3 (adapted from Brennan and others, 2024). Inset A shows the geochronology results for sample DB22BPR03 sampled directly below the Robb Creek Tuff mesa. Inset B shows the suggested correlation and age of the Rob Creek measured section from Garson (1992).

near the border of the Belmont Park Ranch and Home Park Ranch 7.5' quadrangles, which were mapped by Brennan and others (2024) and Brennan (in press). This mesa is capped by ca. 1.9 Ma rhyolitic tuff, correlated with the 2.0 Ma Huckleberry Ridge Tuff of Yellowstone (Garson, 1992). The erosion-resistant welded tuff here overlies reworked, cross-bedded, tuffaceous sandstones interbedded with fluvial round cobble channels consistent with the lithologies that define the Anderson Ranch and Big Hole River members of the Sixmile Creek Formation. Hike to the location of sample DB22BPR03 (45.0107°N, -112.1275°W). The tephra bed at this location was the first tuffaceous horizon identified below the Robb Creek tuff and yielded a zircon U-Pb age of 10.1 ± 0.1 Ma (MSWD = 2.6), along with two older grains at ca. 42 and 49 Ma (fig. 5A; Brennan and others, 2024). Approximately 120 m (~400 ft) to the east, at the base of the mesa, the Robb Creek tuff (sample JS22BPR01, 45.0108°N, -112.1256°W) yielded a zircon U-Pb age of 1.9 ± 0.1 Ma (MSWD = 1.1; Mosolf and others, 2023a). The Robb Creek tuff contains 1- to

4-cm sanidine laths with similar sized basalt, tuff, and pumice lithics supported in an aphanitic groundmass.

Interestingly, these mapping and geochronology results suggest that minimal 10 to 2 Ma strata of the Sixmile Creek Formation are preserved here, and the ca. 1.9 Ma Robb Creek Tuff mesa shows a similar inverted topographic relationship to the Timber Hill basalt mesa. However, while the ca. 6.0 Ma Timber Hill basalt overlies strata with zircon U-Pb dates of ca. 16–7.5 Ma, suggesting relatively continuous late Miocene deposition, the Robb Creek Tuff overlies ca. 10 Ma strata (fig. 6), indicating basin exhumation prior to its eruption. This relationship differs from the interpretations of previous workers mapping the upper Ruby Valley (Garson, 1992), who lacked detailed geochronologic constraints but suggested a relatively complete ca. 16–2 Ma stratigraphic record within Sixmile Creek strata based mostly on the correlation of tuffaceous horizons above and below the Timber Hill basalt and Robb Creek tuff (fig. 5B).



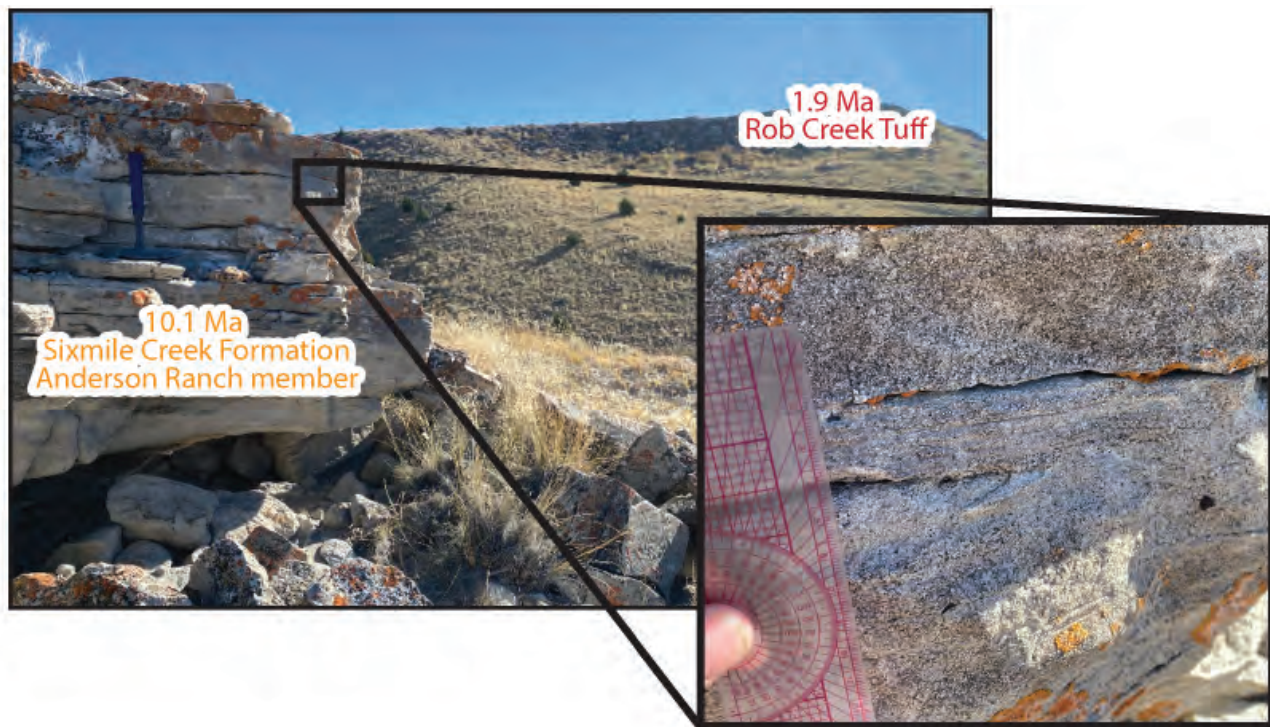


Figure 6. Field photo from STOP 3, showing the relationship between the ca. 1.9 Ma Rob Creek Tuff and ca. 10.1 Ma underlying Sixmile Creek Formation.

For optional STOP 4 (not included in Road Log mileage), turn around and drive back towards Upper Ruby Road.

6.0 mi Just before you cross the upper Ruby River riparian zone, turn right (southeast) onto Peterson Lane. This is a private road owned by the Snowcrest Ranch and permission must be granted before accessing it. Follow Peterson Road southeast.

7.9 mi Peterson Road turns southwest and con-

tinues up Cream Creek ~2.4 mi to optional STOP 4 (fig. 7). If access is granted, closed gates and bad road conditions may require hiking part of the way.

STOP 4 (optional)—Variations in Sweetwater Member of Sixmile Creek Formation

10.3 mi

Park (45.0364°N, -112.0644°W),
Tephra Bed Sample (45.0362°N, -112.0658°W)

STOP 4 is within the Home Park Ranch 7.5' quadrangle

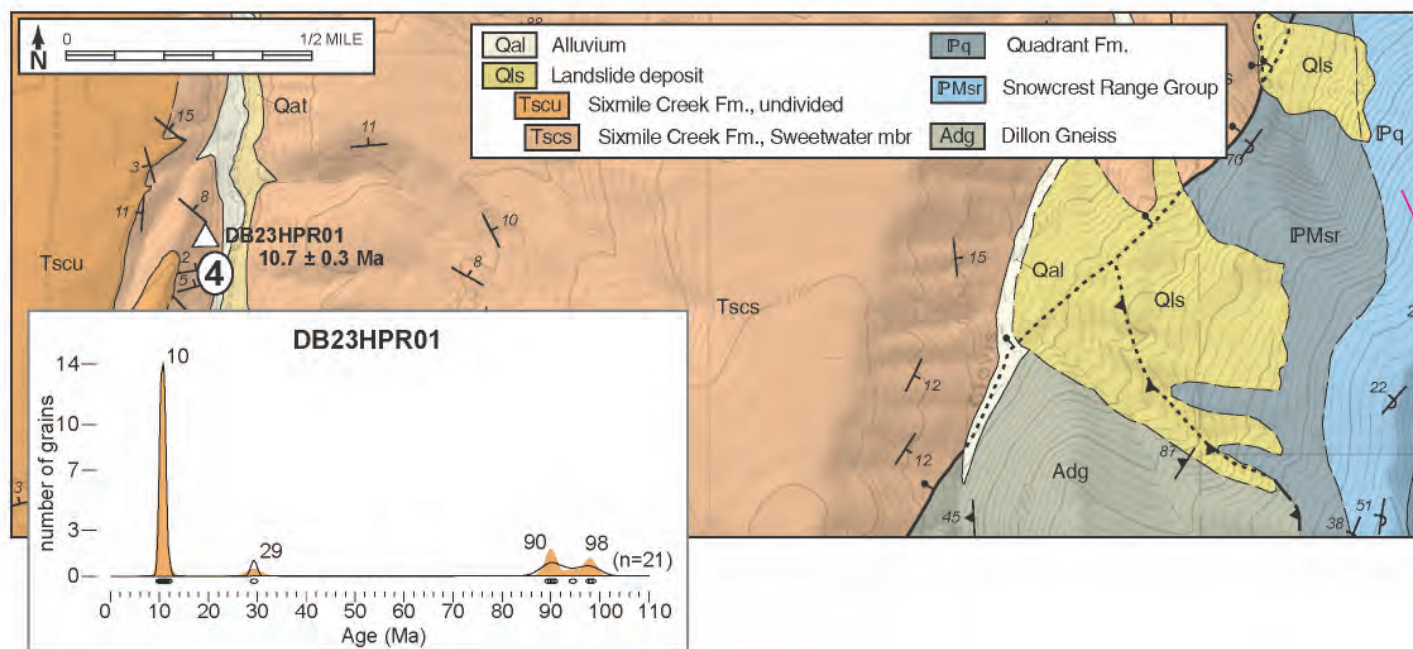


Figure 7. 1:24,000-scale geologic map of the Cream Creek area including field trip optional STOP 4 (adapted from Brennan, in press). Inset A shows the geochronology results for sample DB23HPR01.



(Brennan, in press) and contains an impressive ~10 m (33 ft) outcrop of poorly sorted, mostly angular, chaotic, cobble to boulder conglomerate that is consistent with lithologies that define the Sweetwater Creek member, and is interpreted as a fanglomerate, or ancient alluvial fan deposit. The clast compositions generally reflect the adjacent lithologies of the Snowcrest Range to the east, and include relatively common basement metamorphic clasts, some limestone and quartzites that resemble the Snowcrest Range Group and Quadrant Formation, and some large volcanic clasts that resemble a variety of volcanic units in the Snowcrest/Gravelly Range (Lonn and others, 2000; Kellogg and Williams, 2006). A thinner (couple of centimeter-thick) reworked tephra bed sampled from the base of the conglomerate cliff (sample DB23HPR01, figs. 8A, 8B) yielded a zircon U-Pb age of 10.7 ± 0.3 (MSWD = 9.4), indicating that this outcrop has a similar depositional age to the thicker reworked, cross-bedded tuffaceous sandstones and relatively well-sorted, well-rounded quartzite cobble conglomerates viewed at STOP 3.

If you continue to hike along Cream Creek, you can trace the fanglomerate outcrops along-strike for several miles. Along-strike, the predominant cobble and boulder lithologies vary in accordance with the proximally exposed bedrock in the adjacent Snowcrest Range. For example, ~2.5 mi (4 km) to the south, where Paleozoic limestones are the predominant lithology in the Snowcrest Range, the fanglomerate deposits are almost entirely limestone lithologies with some remarkably angular clasts (figs. 8C, 8D).

For STOP 5, turn around and drive back to the junction of Ledford Creek Road and Upper Ruby Road (6.6 mi).

60.6 mi At the junction of Ledford Creek Road and Upper Ruby Road, turn left (north) onto the Upper Ruby Road for drive for 12 mi to Barton Gulch Road.

72.6 mi Turn right (east) on Barton Gulch Road. Continue ~0.2 mi and park for STOP 5, the final stop of the field trip.

STOP 5—Barton Gulch: Overlapping Facies of the Sixmile Creek Formation

72.8 mi

Parking (45.2230°N, -112.1119°W)

Outcrop Traverse (45.2235°N, -112.1066°W) to (45.2217°N, -112.1008°W, Sample SC-3)

STOP 5 is a short (~0.3 mi) hike along outcrops on the north side of Barton Gulch Road, which is near the border of the Ruby Dam and Metzel Ranch 7.5' quadrangles. Prior mapping in this region was done by Monroe (1976) and St. Jean and Teeter (2004). Brennan and others revisited this area in the 2023 and 2024 field seasons with anticipated publication in 2026–2027 as an MBMG open-file report.

The upper Ruby Valley narrows from an approximately east–west width of ~15 mi (24 km) to the south near field trip STOPS 2 and 3, to a width of less than 4 mi (~6.5 km) here at STOP 5. Distinguishing between the members of the Sixmile Creek Formation in this narrower portion of the valley is not as easy as at prior stops of this field guide, perhaps reflecting depositional system complexities relating to the geometry of the Ruby graben. The traverse will consist of walking along-strike in the Sixmile Creek Formation from west to east, to the location of sample SC-3. Sixmile Creek strata exposed along this traverse exhibit remarkably variable depositional characteristics. The first, most western portion of the outcrop consists of matrix-supported, poorly sorted, angular clast conglomerate with lithologies consistent with Ruby Range metamorphic basement rocks exposed to the west–southwest (fig. 9A), where these basement rocks are overlain by mudstone and siltstone of the Renova Formation. This part of the Sixmile Creek Formation is interpreted to be debris flows sourced from Ruby Range basement and overlying, poorly indurated, finer-grained Renova Formation units, consistent with basin-lateral transport. To the east, coarser, angular basement-clast, matrix-supported conglomerates gradually shift to well-rounded,

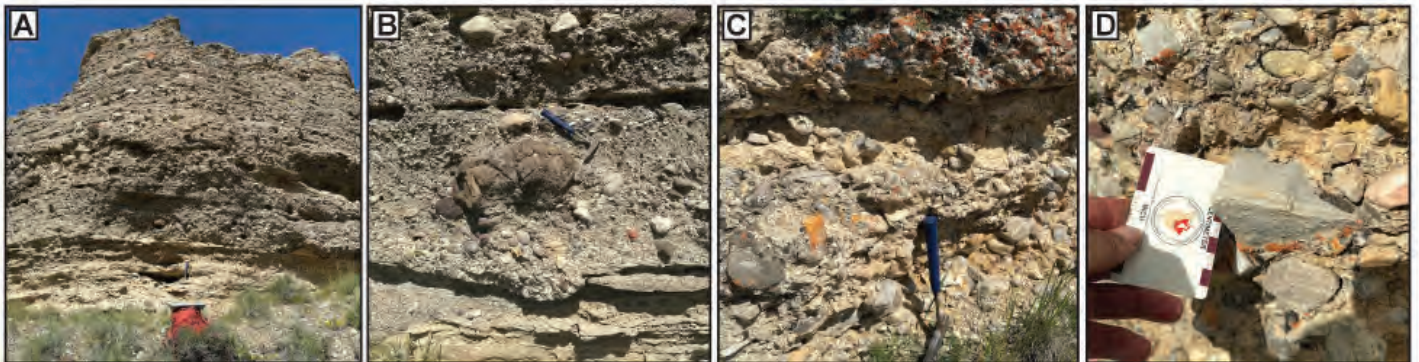


Figure 8. Field photos of the lithologies and facies present along the Cream Creek drainage at optional STOP 4. Photos A and B show the finer-grained tuffaceous bed collected for sample DB23HPR01 and the overlying angular, locally matrix supported, primarily metamorphic basement and volcanic clasts interpreted to represent a fanglomerate deposit. Photos C and D show a similar facies conglomerate, with almost entirely angular limestone clasts, found ~2.5 mi (4 km) along-strike to the south.



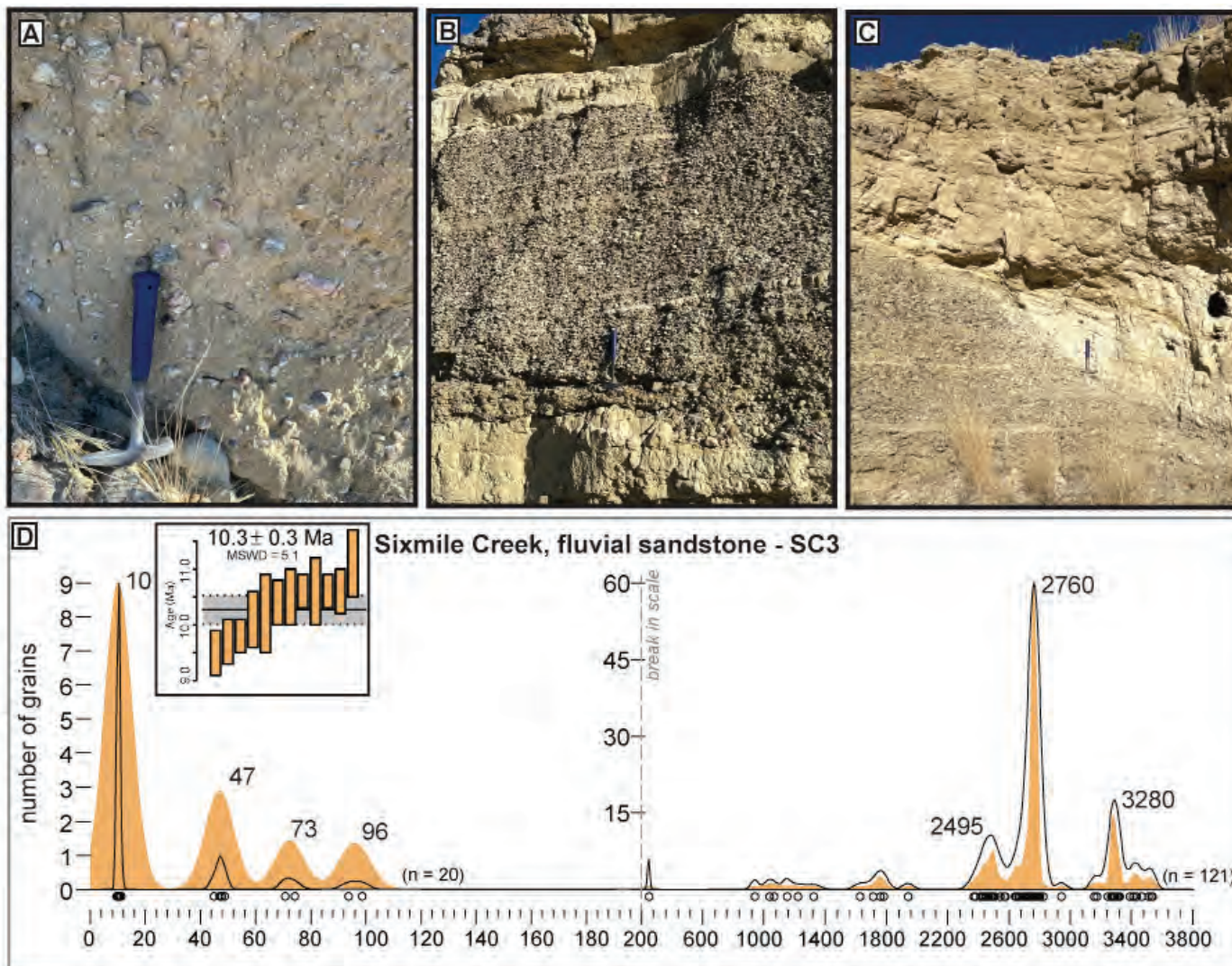


Figure 9. Field photos of the lithologies and facies present along Barton Gulch at STOP 5. Photo A shows the muddy/sandy matrix-supported, primarily angular metamorphic conglomerate interpreted to reflect a debris flow sourcing Archean basement and overlying fine-grained Renova Formation rocks to the southwest. Photos B and C show a clast-supported, round pebble to small cobble, primarily quartzite and volcanic clast conglomerate channel incised into and by finer-grained tuffaceous sandstones. Inset D shows the zircon U-Pb geochronology results from sample SC-3 collected from fluvial facies of the Sixmile Creek Formation along Barton Gulch, similar to that shown in photos B and C. Sample SC-3 was collected by Stuart Parker in 2016 (unpublished) and the data are available upon request.

better sorted, quartzite-clast-supported conglomerate channels incised in and by tuffaceous sandstone and finer-grained beds (fig. 9B and C). These lithologies are more consistent with far-traveled, extra-Ruby graben sources, brought in by basin axial, fluvial transport.

Sample SC-3 was collected from a tuffaceous sandstone near the end of the Barton Gulch traverse and yielded a 10.3 ± 0.3 Ma (MSWD = 5.1 Ma) maximum depositional age (fig. 9D), indicating this outcrop is similar in age to the Sixmile Creek strata observed at STOPS 3 and 4. However, the lithologies viewed along the Barton Gulch traverse include characteristics that define all three members, proximally sourced texturally immature debris flow conglomerates (Sweetwater member), fluvially reworked tephra beds (Anderson Ranch member), and fluvial round quartzite cobble conglomerates (Big Hole River member). This hetero-

geneity reflects the terrestrial basin depositional system of the Sixmile Creek Formation and exemplifies the difficulty in mapping out the Sweetwater, Anderson Ranch, and Big Hole River members as coherent stratigraphy instead of interfingering basin facies.

SUMMARY AND FUTURE WORK

This field guide includes stops demonstrating the traditional three-member view of the Sixmile Creek Formation (STOPS 1 and 2), followed by several stops (STOPS 3, 4, and 5) showing how the facies, lithologic characteristics, and preserved duration of Sixmile Creek Formation sedimentation varied within the basin at ca. 10 Ma. The last three stops demonstrate the problem of viewing the lithologically defined members of the Sixmile Creek Formation in a chronostratigraphic framework. It is the view of this author that the lithologic characteristics of the Sixmile



Creek Formation, at least in the Ruby Graben, vary primarily with position within the basin (i.e., space), as opposed to varying with stratigraphic position (i.e., time).

Mapping in the lower Ruby and Beaverhead basins, Petkewich (1972) took an approach to defining members of the Sixmile Creek Formation that is similar to how Brennan and others (2024) subdivided the formation. Petkewich (1972) mapped local, proximally derived, poorly sorted, angular conglomerates that fine towards the basin center as one member. This member is most similar to the Sweetwater Creek “fanglomerate” member of Fritz and Sears (1993) and Fritz and others (2007) and was mapped as such by Brennan and others (2024) and Brennan (in press). This proximally derived conglomerate member does not always occupy a basal position or consist of metamorphic clasts but rather represents basin-marginal facies, and thus records the basin margin geometry of the Sixmile Creek basins during deposition. The clast composition in this unit varies in accordance with the adjacent exhumed lithologies. For the more rounded, better sorted, fluvial conglomerate channels that primarily record basin-axial transport, Petkewich defined a different member that is most like the Big Hole River “round cobble conglomerate” member. However, instead of being the youngest or stratigraphically highest member, this member simply reflects a central, or axial, basin facies and is comprised of lithologies that reflect the up-catchment geology, including far traveled lithologies sourced from outside the Ruby graben. Both basin marginal and axial facies contain tephtras, reworked tuffs, and tuffaceous sandstones of varying thickness (although generally thicker near the basin center) and continuity. These tuffa-

ceous units are most consistent with the original definition of the Anderson Ranch “tephra” member and in locations may reflect lahar deposits (Rob Thomas, oral commun.). Given that without geochronology or fossil controls it is difficult to differentiate or correlate separate tephra and tuffaceous exposures within a single basin, let alone from basin to basin, it may be more useful in the field to initially map these tephra/tuffaceous intervals as “beds.” Eventually, subsequent geochronology can date these beds and allow for correlation across basins and regionally. Brennan and others (2024) and Brennan (in press) map the axial facies of the Sixmile Creek Formation as Sixmile Creek undivided (Tscu on figs. 4, 5, 7) as it predominantly contains lithologies consistent with both the Anderson Ranch and Big Hole River members. Brennan and others (2024) did delineate tephra and tuffaceous beds within “Tscu” where they were of mappable thickness and extent at 1:24,000 scale.

Mapping the Sixmile Creek Formation members primarily as facies, instead of stratigraphic members, and following up with subsequent dating of key volcanic beds has several advantages. For example, following this approach deposition of the Sixmile Creek Formation in the Ruby graben has been constrained to ca. 16–6 Ma with reworked tuffaceous units at ca. 16, 13, 10, 9, 8.5, and 7.5 Ma, and volcanic units at ca. 6 and 2 Ma. Sediments beneath the ca. 6 Ma Timber Hill basalt appear to be a complete basin record that spans ca. 16–6 Ma, coinciding with the McDermitt, Owyhee–Humboldt, Bruneau–Jarbridge, Twin Falls, and Picabo fields of the Yellowstone hotspot track (fig. 10). Notably, in the Red Canyon and Belmont Park Ranch quadrangles (Mosolf and Sears, 2024; Brennan and others,

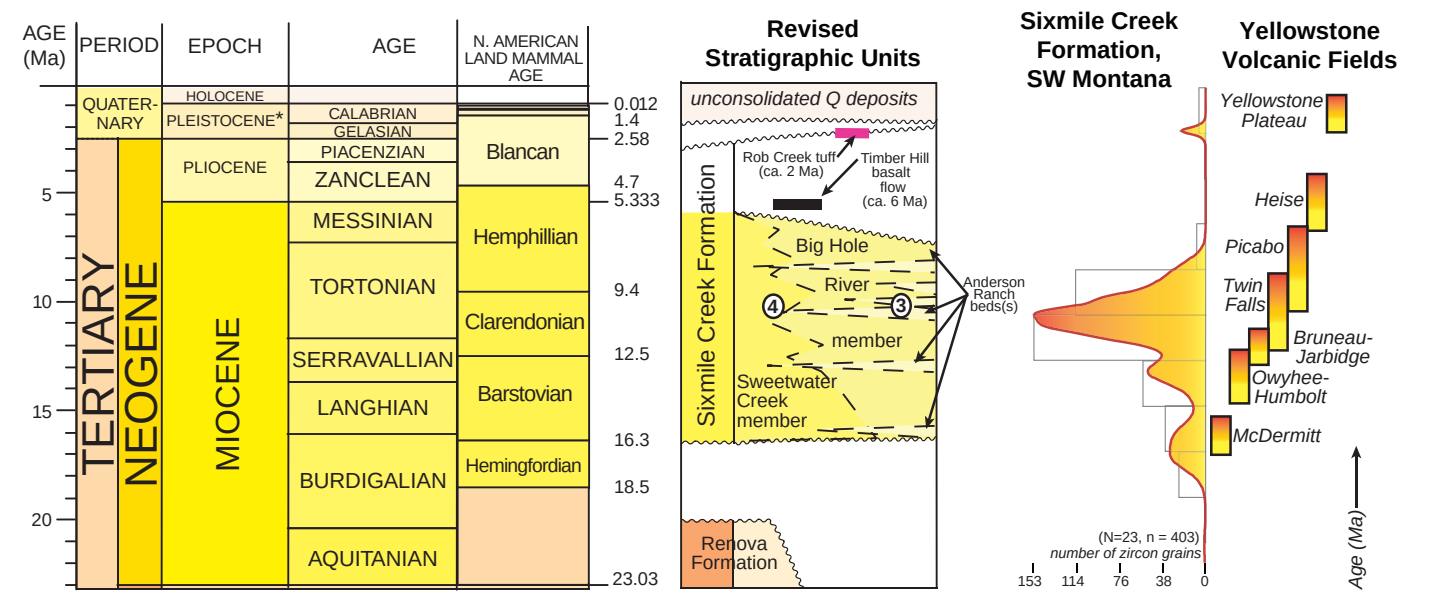


Figure 10. The Neogene and Quaternary timescale with my proposed revised subdivision of the Sixmile Creek Formation. Please refer to figure 4 for the conventional view of the Sixmile Creek Formation. The approximate stratigraphic locations of field trip STOPS 3 and 4 are shown to demonstrate how this approach could more accurately delineate variations with the Sixmile Creek Formation in the field. The right section of the figure shows a compilation of MBMG zircon U-Pb geochronology results from the Sixmile Creek Formation in the Dillon region, alongside a timeline of the Yellowstone hotspot track. Note how, at least in the Dillon region, the Sixmile Creek Formation appears to contain a stratigraphic record of the McDermitt, Owyhee–Humboldt, Bruneau–Jarbridge, and Twin Falls volcanic fields, but appears to “shut off” with the onset of the Heise volcanic field.



2024), there are no Sixmile Creek sediments overlying the Timber Hill basalt. However, some undated and thin gravel deposits do overlie the Timber Hill basalt to the southwest (Rob Thomas, oral commun.). Near the center of the basin, however, the ca. 2 Ma Robb Creek tuff overlies ca. 10 Ma Sixmile Creek strata, indicating erosion and removal of most ca. 9–6 Ma sediments except where they were protected by the resistant ca. 6 Ma, Timber Hill basalt “cap.” This incised relationship suggests that the proto-Ruby River and its tributaries were exhuming sediment rather than accommodating it after ca. 6.0 Ma. Interestingly, ca. 6 Ma coincides with the end of volcanic activity in the Picabo field and onset of activity in the Heise field (see figs. 1, 10).

As far as basin geometry, the occurrence of ca. 16–6 Ma basin marginal facies within the Sixmile Creek Formation near the modern basin margins suggests that the present geometry of the upper Ruby Valley and its bounding ranges is similar to those of the late Miocene, albeit with some alteration by active northwest-trending faults such as the Sweetwater fault (STOP 1). Additionally, the presence of marginal facies in some locations transitioning up-section to axial facies (like viewed at STOP 2) is also consistent with filling of the Ruby graben from ca. 16 to 6 Ma. Interestingly, the modern drainage network of the upper Ruby River, (i.e., modern basin axial facies) cuts right across several notable exposures of late-Miocene basin-margin facies. For example, the modern upper Ruby River makes a 90° bend near the upper Ruby Canyon where it cuts a ~900 m (3,000 ft) canyon through the Snowcrest–Greenhorn Range and late Miocene Fanglomerate deposits (like those viewed at STOP 4). Near the Ruby Reservoir, late Miocene basin axial fluvial facies of the Sixmile Creek Formation (observed at STOP 5) are also offset to the east from the modern basins axial drainage system through the lower Ruby River Canyon.

Collectively, these observations suggest that the Ruby graben accommodated Sixmile Creek sedimentation from ca. 16 to 6 Ma. Sometime after 6 Ma, here we suggest coincident with the onset of the volcanism in the Heise field, the Ruby graben began evacuating sediment as opposed to preserving it. After the Ruby graben transitioned from accommodating to evacuating sediment, the locus of basin-axial drainage appears to have shifted, with originally southwest to northeast flowing drainage systems, like that of the upper Ruby River and some of its notable tributaries like Sweetwater Creek, turning approximately 90° and incising significant topography. Consequently, at least some of the modern relief on mountain ranges like the Blacktail, Snowcrest, Ruby, and Greenhorn Ranges may reflect basin excavation. Facies mapping and systematic geochronology of additional Neogene basins across southwestern Montana may reveal how widespread this late Miocene and younger drainage pattern reorganization is and thus provide a field-based approach of evaluating large-scale models of

drainage divide migration associated with the Yellowstone Hotspot track (e.g., Staisch and others, 2022).

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FIELD GUIDE TO BANNACK STATE PARK, SOUTHWEST MONTANA

Jesse G. Mosolf,¹ Thomas Kalakay,² and Raymond Salazar¹

¹Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

²Rocky Mountain College, Billings, Montana

INTRODUCTION

This field guide explores the geology of Bannack State Park and the surrounding area. The historic ghost town holds an important position in Montana's mining and political history, most notably being the first territorial capital soon after the 1862 discovery of placer gold in Grasshopper Creek (see: Shenon, 1931; Sassman, 1941; Geach, 1972). While Bannack is best known for its colorful mining history, its natural history is equally colorful, culminating in a metallogenic geoframework supporting Montana's first major gold rush. Bannack's geology provides an exceptional record of Cordilleran thrust belt deformation and contemporaneous arc magmatism in southwestern Montana (fig. 1), detailing a chronology of folding and faulting, synorogenic volcanism and sedimentation, and post-kinematic pluton emplacement and mineralization.

Bannack's geology is spectacularly exposed in the Grasshopper Creek gorge east of the historic town center (fig. 2). The trip begins in Bannack and travels eastward by foot and car, ending near the southern trace of the ERMONT thrust fault. An optional stop visits the Muddy–Grasshopper extensional fault and associated Cenozoic sedimentary deposits near the Bannack town center. Two stops involve moderate hikes (<1.8 km, 1.1 mi), while the main stop

entails a strenuous 6.5 km (4 mi) hike, climbing 260 m (850 ft). Parts of the main hike include off-trail scrambling through cliffy exposures of the Beaverhead Group. Be prepared for extreme temperatures with little to no shade. Driving will be minimal but requires a sturdy chassis vehicle capable of traversing rough gravel roads. Participants should bring lunch, plenty of water, and off-trail hiking apparel.

STRATIGRAPHY

Basement Rocks and Phanerozoic Sedimentary Cover

The oldest rock near Bannack is Archean crystalline basement exposed in the Armstead anticline approximately 10 km (6.25 mi) south of Bannack (figs. 2, 3). The basement rock generally consists of quartzofeldspathic gneiss and amphibolite with minor metasedimentary intercalations, including metapelite, quartzite, and calc-silicate marble. The metamorphic assemblage resembles the Dillon Gneiss in the nearby Ruby Range (James, 1990; Mosolf, 2021; Mosolf and Sears, 2024). The basement rocks are overlain by a thick sequence of Phanerozoic sedimentary cover exceeding 2 km (1.25 mi) in stratal thickness, including Paleozoic marine and siliciclastic sequences of the Mississippian Madison Group, Mississippian–Permian

Snowcrest Range Group, Pennsylvanian Quadrant Formation, and the Permian Phosphoria Formation. Most Paleozoic rocks exposed in the Grasshopper Creek gorge are Madison Group carbonates that were variably metamorphosed to calc-silicate hornfels and skarn near contacts with granodioritic plutons, forming metallogenic deposits locally. South of Bannack, parts of the lower Mesozoic sequence (Dinwoody, Swift, and Kootenai Formations) are exposed in the western limb of the Armstead anticline. The rocks are dominantly siliciclastic, recording the eastward advance of the Cordilleran fold-thrust belt. Synorogenic volcano-sedimentary deposits of the Beaverhead and Bannack Groups unconformably rest on the older Phanerozoic sequences, signaling the onset of local thrust belt shortening. West of Bannack, coarse, unconsolidated Eocene–Miocene(?) sedimentary deposits filling the extensional Grasshopper Basin are likely time-correlative to the Tertiary Bozeman Group. Landslides and rock avalanches are common throughout the area.

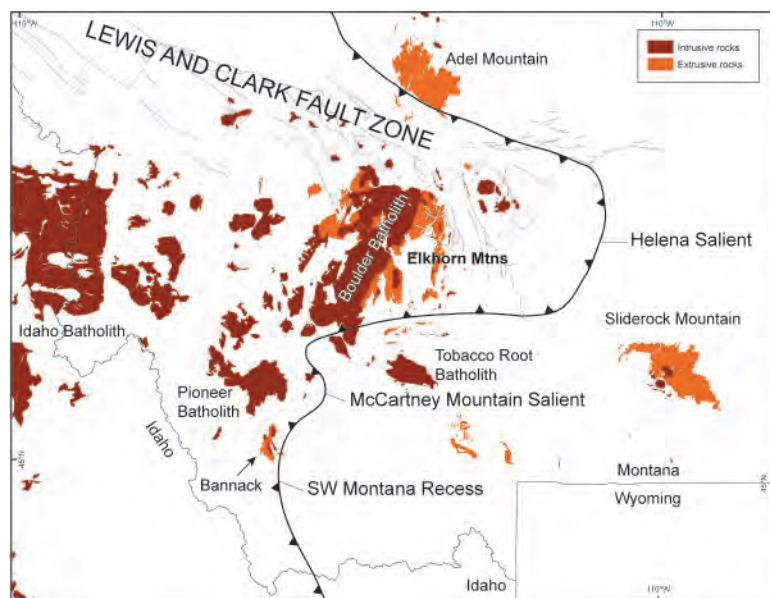
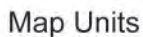


Figure 1. Regional tectono-geologic map of western Montana. The simplified overlay shows the extent of Cretaceous igneous rocks in Montana; major volcanic centers and batholiths are labeled. Compiled from Vuke and others (2007), Lewis and others (2012), and Montana Bureau of Mines and Geology (2025).





Phanerozoic igneous rocks

Tv	Tertiary volcanic
Kv	Cretaceous volcanic
Kg	Cretaceous granitoid intrusion

Proterozoic and Archean rocks

Xga	Gneiss and amphibolite
Yb	Belt Supergroup, undivided
XAm	Marble, schist and gneiss
XAqfg	Quartzofeldspathic gneiss
Am	Amphibolite
Aqfg	Quartzofeldspathic gneiss

_____ contact

	normal fault, certain, approximate, concealed ball and bar on downthrown block
	thrust fault, certain, approximate, concealed, teeth on upthrown block

Late Cretaceous Volcano-Sedimentary Deposits

unit of the Bannack Group, consists of highly silicified and zeolitized crystal–lithic tuff intervals interpreted to be pyroclastic flow and ash-fall deposits (Pearson and Childs, 1989). The overlying Cold Spring Creek Formation is a complex sequence of volcanic agglomerates, andesite–dacite lava flows, and pyroclastic flow deposits with U–Pb ages spanning 73.6–71.7 Ma. The waning stages of volcanism in the Bannack area were marked by the extrusion of the Bon Accord Porphyry lava dome complex (72.6–71.4 Ma) and associated Grasshopper Prospect rhyolite (approximately 72.5 Ma) and the emplacement of the Bannack–Blue Wing and McDowell Springs granodiorite ca.

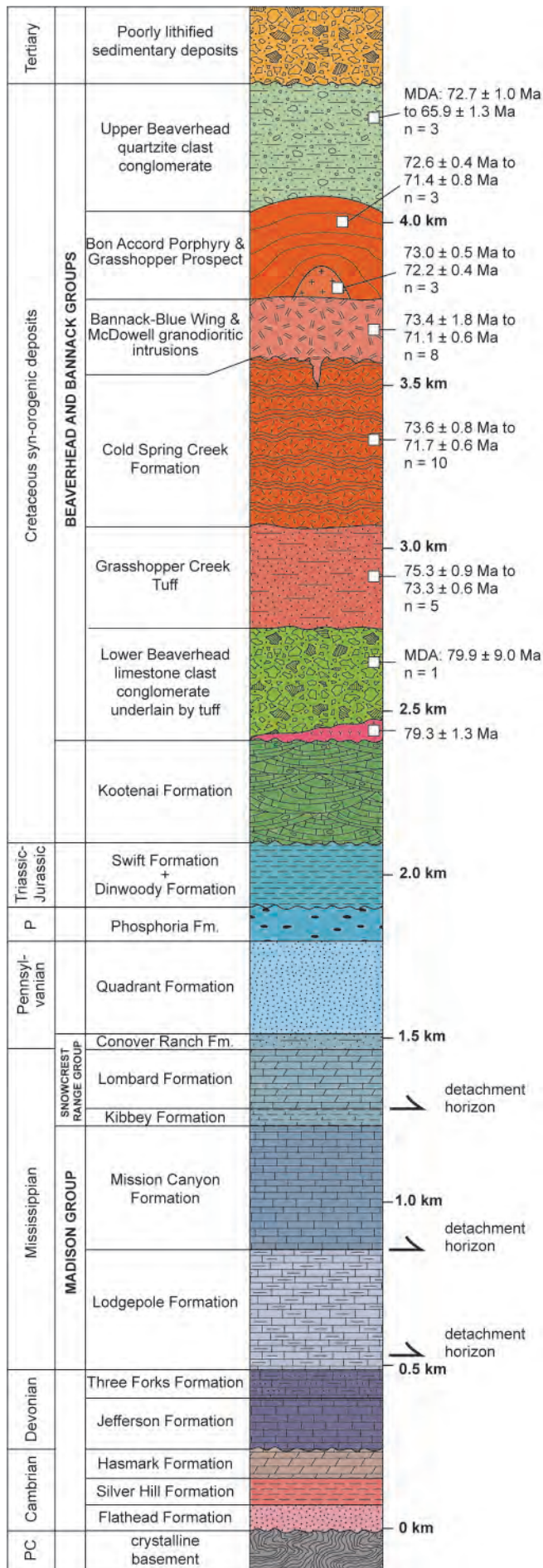


Figure 3. Generalized chrono-stratigraphic column for the Bannack area. Note that the thickness of the units varies significantly throughout the region. U-Pb zircon ages from the volcanic units are from Mosolf and McDonald (2022). The andesitic tuff age is from Murphy (2000). Max depositional ages from the Beaverhead Group are from Laskowski and others (2013), Garber and others (2020), and Mosolf (2021).



TABLE 1. LA-ICPMS U-Pb zircon geochronology data from the Dillon 30' x 60' quadrangle

Sample	Lithology	Unit or Formation	7.5' quadrangle	Latitude (°N)	Longitude (°W)	Method	Age (Ma)	2σ	MSWD*
JM21BK01	tuffaceous sandstone	Medicine Lodge beds?	Bannack	45.2324	-112.9947	MDA 206/238	30.0	0.2	0.6
TAM-2	granodiorite	Pioneer Batholith	Twin Adams	45.4876	-112.8663	WM 206/238	70.5	0.6	1.4
JM21TM06	granodiorite	Pioneer Batholith	Torrey Mountain	45.4823	-112.9454	WM 206/238	70.6	0.5	0.3
JM21TM01	tonalite	Pioneer Batholith	Torrey Mountain	45.4559	-112.9730	WM 206/238	71.0	0.9	1.7
JM21TM03	granodiorite	Pioneer Batholith	Torrey Mountain	45.4805	-112.9655	WM 206/238	71.0	0.4	0.9
TAM-1	granodiorite	Pioneer Batholith	Twin Adams	45.4933	-112.8611	WM 206/238	71.2	0.5	1.2
JM21TM04	aplite	Pioneer Batholith	Torrey Mountain	45.4783	-112.9501	WM 206/238	71.7	0.5	1.4
JM21TM02	luecogranite	Pioneer Batholith	Torrey Mountain	45.4450	-112.9653	WM 206/238	71.9	0.7	2.7
ES-14	sandstone	Upper Beaverhead Group	Eli Spring	45.1213	-112.8918	MDA 206/238	72.7	1	1.5
DN-60	diorite	Bannack-BlueWing	Bannack	45.1772	-112.9499	WM 206/238	71.1	0.6	0.5
DN-55	granodiorite	McDowell Springs	Bannack	45.1279	-112.9106	WM 206/238	71.8	0.5	0.6
DN-62	granodiorite	Bannack-BlueWing	Bannack	45.1967	-112.9416	WM 206/238	71.9	0.7	1.1
BK-02	trachyandesite	Bannack-BlueWing	Bannack	45.1556	-112.9679	WM 206/238	72.1	0.4	1.3
JM21BK04	granodiorite	Bannack Pluton	Bannack	45.1478	-112.9817	WM 206/238	72.4	1.0	1.3
JM21BK05	granodiorite	Bannack Pluton	Bannack	45.1511	-112.9814	WM 206/238	72.6	0.7	1.5
DN-52	diorite	Anniversary Pluton	Bannack	45.1635	-112.9805	WM 206/238	72.6	0.4	2.0
DN-61	granodiorite	Bannack Pluton	Bannack	45.1558	-112.9820	WM 206/238	73.4	1.8	1.2
JM24BNK03	altered porphyry	Grasshopper Prospect	Bannack	45.1578	-112.9121	WM 206/238	72.2	0.4	1.2
JM24BNK02	altered porphyry	Grasshopper Prospect	Bannack	45.1577	-112.9228	WM 206/238	72.6	0.5	0.8
DN-57A	altered porphyry	Grasshopper Prospect	Bannack	45.1578	-112.9120	WM 206/238	73.0	0.5	0.8
DN-57B	dacite	Bon Accord Porphyry	Bannack	45.1508	-112.9089	WM 206/238	71.4	0.8	0.4
JM24BNK04	dacite	Bon Accord Porphyry	Bannack	45.1582	-112.9087	WM 206/238	71.9	0.7	1.5
JM24BNK01	dacite	Bon Accord Porphyry	Bannack	45.1512	-112.9278	WM 206/238	72.6	0.4	0.8
DN-63	dacite	Cold Spring Creek	Bannack	45.1700	-112.9307	WM 206/238	71.7	0.6	1.4
DN-08	dacite	Cold Spring Creek	Eli Spring	45.1023	-112.8863	WM 206/238	71.8	0.6	1.0
DN-20	tuff	Cold Spring Creek	Dalys	45.1126	-112.8282	WM 206/238	72.0	0.6	0.7
ES-08	dacite	Cold Spring Creek	Eli Spring	45.0768	-112.8842	WM 206/238	72.0	0.5	0.7
ES-03	trachydacite	Cold Spring Creek	Eli Spring	45.1021	-112.9730	WM 206/238	72.5	0.8	1.1
ES-12	dacite	Cold Spring Creek	Eli Spring	45.0971	-112.8955	WM 206/238	72.6	0.5	1.0
DN-59	andesite	Cold Spring Creek	Bannack	45.1545	-112.9448	WM 206/238	72.9	0.6	1.8
DN-56	dacite	Cold Spring Creek	Bannack	45.1413	-112.9062	WM 206/238	73.1	0.8	0.7
DN-24	tuff	Cold Spring Creek	Dalys	45.1073	-112.8096	WM 206/238	73.3	0.5	1.1
DN-12	tuff	Cold Spring Creek	Dalys	45.1138	-112.8256	WM 206/238	73.6	0.8	0.9
DN-66	rhyolitic tuff	Grasshopper Creek Tuff	Bannack	45.2319	-112.9184	WM 206/238	73.3	0.6	1.8
DN-54	rhyolitic tuff	Grasshopper Creek Tuff	Bannack	45.1402	-112.9640	WM 206/238	73.4	0.6	0.6
ES-01	rhyolitic tuff	Grasshopper Creek Tuff	Eli Spring	45.1171	-112.9546	WM 206/238	74.7	1.2	2.0
DN-11	rhyolitic tuff	Grasshopper Creek Tuff	Dalys	45.0842	-112.8661	WM 206/238	74.7	0.6	0.4
JM19BM41	rhyolitic tuff	Grasshopper Creek Tuff	Burns Mountain	45.2197	-112.8685	WM 206/238	75.3	0.9	0.1

*MSWD is the Mean Square Weighted Deviation.

Method: WM 206/238 weighted mean of select $^{206}\text{Pb}/^{238}\text{U}$ datesMDA 206/238 max depositional age, weighted mean of youngest $^{206}\text{Pb}/^{238}\text{U}$ dates

Zircon separates were prepared at MBMG and analyzed by LA-ICPMS at the University of California, Santa Barbara.

Latitudes and longitudes are in the 1984 World Geodetic Survey (WGS84) datum.



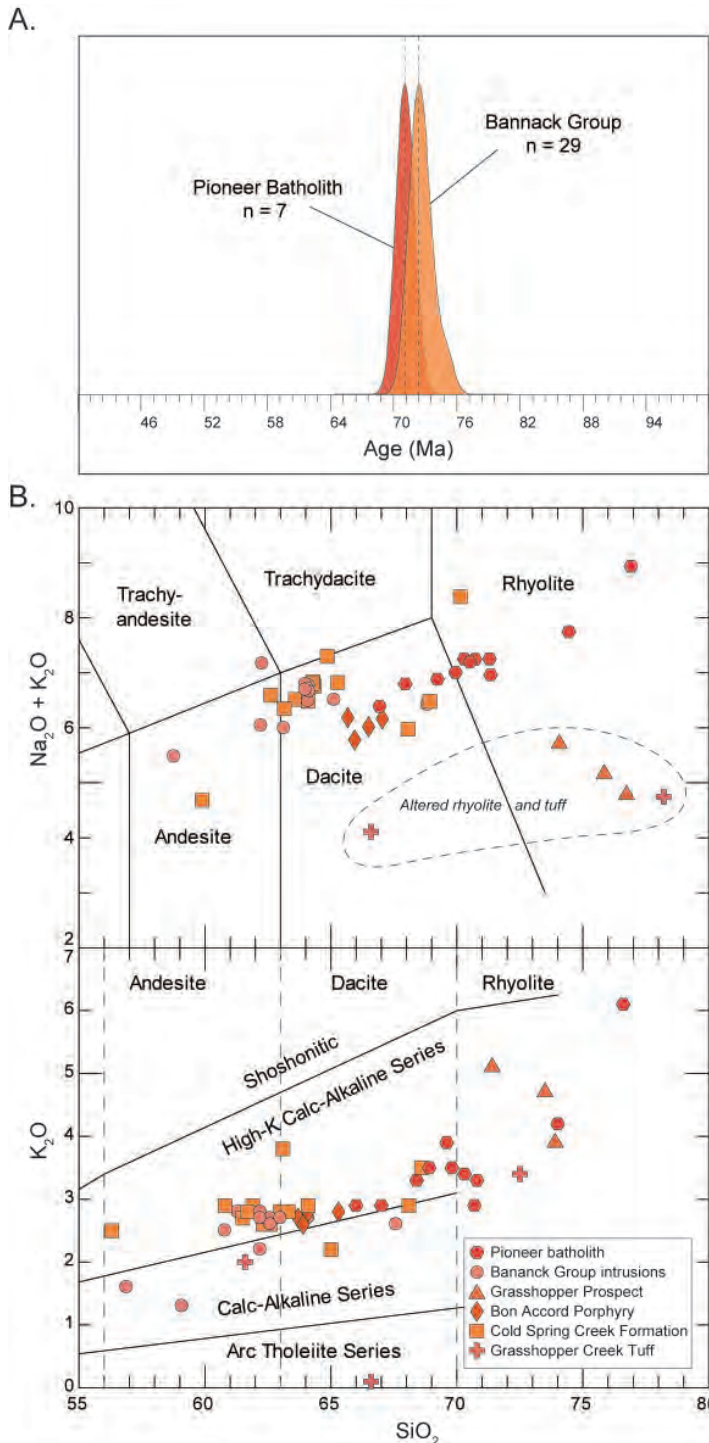


Figure 4. Age and geochemistry data from the Bannack Group and Pioneer Batholith in the Dillon 30' x 60' quadrangle (Mosolf and McDonald, 2022). (A) Kernel density plots of U-Pb zircon ages from volcanic and plutonic units. (B) Whole-rock geochemical data plotted on rock-type diagrams after Le Bas and others (1986) and Peccerillo and Taylor (1976).

73.4–71.1 Ma, including the Anniversary and Bannack plutons. The Upper Beaverhead Group unconformably rests on the volcanogenic sequence, consisting of poorly lithified conglomeratic deposits dominated by Proterozoic quartzite clasts; U-Pb max depositional ages span 72.7–71.6 Ma (Laskowski and others, 2013; Mosolf and McDonald, 2022).

The Bannack Group magmatic units generally have high-potassium, calc-alkaline geochemical trends (fig. 4) and are enriched in incompatible elements (Ivy, 1988). Most of the samples dated in this study contain significant zircon inheritance with ages spanning approximately 2.8–1.7 Ga (Mosolf, 2017; Mosolf and McDonald, 2022), indicating that parental melts assimilated evolved crustal rocks similar to the basement rocks exposed in the nearby Armstead anticline (fig. 2; Coryell 1983; Mueller and others, 2012).

STRUCTURAL GEOLOGY

Regional Setting

The North American Cordillera consolidated during Jurassic–Paleogene subduction and associated terrane accretion, forming an orogenic system stretching over 4,000 km (2,490 mi) from Alaska to Mexico. It is considered an archetype continental subduction orogen with several inter-related tectonic components comprising a forearc accretionary complex and basin, magmatic arc, retroarc hinterland, Sevier fold-thrust belt, foreland basin, and Laramide foreland belt (e.g., Yonkee and Weil, 2015). In southwest Montana near Bannack, many of the retroarc components were spatially condensed and possibly overlapping during the Late Cretaceous. Silicic arc magmatism extended eastward at this time and interacted with shortening structures in an area of presumed overlap between the Sevier and Laramide belts (Coryell, 1983; Johnson, 1986; Johnson and Sears, 1988; Lageson and others, 2020).

Bannack is within the southwestern Montana recess, bound by the Helena Salient to the north and the Snake River Plain to the south (fig. 1). Thrust belt structures locally involved Precambrian crystalline metamorphic rocks, which was previously attributed to buttressing of thin- and thick-skinned structures formed during separate orogenic events (e.g., Johnson and Sears, 1988). Subsequent studies suggest that local basement-cored structures formed in areas with a dramatically tapering pre-thrusting stratigraphic architecture or arch during a single progressive deformation rather than discrete orogenic events (Parker and Pearson, 2021; Parker and Pearson, 2023). Silicic plutonism and associated volcanism were generally synchronous with crustal shortening and thickening in Montana ca. 85–55 Ma, possibly utilizing major structures as melt conduits (Kalakay, 2001; Kalakay and others, 2001; Lageson and others, 2020). Major Late Cretaceous vent complexes



formed within and adjacent to the thrust belt in western Montana, including the Bannack, Elkhorn Mountains, Adel Mountain, and Sliderock Mountain volcanic fields (fig. 1) (see Scarberry and others, 2020).

Thrust belt shortening had slowed or ceased in southwest Montana by the early Paleogene as the Cordilleran orogenic belt began to extend by either gravitational collapse or a change in the regional stress field related to plate dynamics. Extension of the Sevier–Laramide belt in Montana started at approximately 53–52 Ma (Constenius, 1996; Foster and others, 2001, 2007, 2010), shortly after or possibly overlapping shortening deformation in the foreland ca. 55–52 Ma (Harlan and others, 1988; Sears and Hendrix, 2004). Crustal extension was accompanied by voluminous magmatism and the formation of several major Eocene vents in western Montana.

Thrust Belt Geology near Bannack

The Bannack area is approximately 10 km (6 mi) east of the Grasshopper thrust plate, a major thrust belt structure that carries Mesoproterozoic Belt Supergroup strata over Phanerozoic units (Ruppel and Lopez, 1984). The Ermont thrust is the most prominent structure near Bannack, with a sinuous trace that is mappable for nearly 30.5 km (19 mi) to the north, near the southern margin of the Pioneer batholith (fig. 2; Brandon, 1985; Pearson, 1996). The Ermont thrust generally places Paleozoic carbonate rocks and minor clastic intervals over Late Cretaceous volcano-sedimentary sequences of the Lower Beaverhead and Bannack Groups. Near Badger Pass, the Ermont thrust gently dips to the west, forming a hanging wall flat above truncated footwall strata (Lowell, 1965). The thrust nappe contains several northerly trending detachment or fault propagation folds crossed by Highway 287. West of Badger Pass, the dip of the Ermont thrust is interpreted to steepen to 20–40° in the subsurface, forming a footwall ramp beneath a series of west-dipping imbricate thrust faults (Kalakay, 2001).

Southward, between Badger Pass and Bannack, the trace of the Ermont thrust is increasingly obscured by Late Cretaceous plutons (Bannack–Blue Wing Granodiorite) that appear to have been emplaced as local shortening slowed or ceased (fig. 5). At Bannack, the Ermont thrust dips to the west, placing Mississippian marine strata over the Beaverhead and Bannack Groups. In the footwall of the thrust, volcano-sedimentary sequences contain growth strata that onlap the east-vergent Madigan Gulch anticline exposed in Grasshopper Creek gorge (fig. 5; Azevedo, 1993). Radiometric dating of growth strata and undeformed plutons intruding the Ermont thrust constrain local shortening to approximately 80–72 Ma. A cartoon block model portraying the chronology of Bannack’s shortening, sedimentation, and magmatism is provided in figure 6.

South of Grasshopper Creek, the Ermont thrust is not easily traced but likely merges with the Armstead thrust

system and basement-cored Armstead anticline (fig. 2). Coryell and Spang (1988) interpreted the Armstead anticline as a duplex structure with an intensely imbricated fault-bend fold. Thermochronology data record rapid cooling of the basement rocks coring the Armstead anticline ca. 100–80 Ma, likely due to shortening exhumation (Mosolf, 2021). Kalakay and others (2001) interpreted the Ermont and Armstead thrust systems as structurally linked, suggesting the Ermont thrust system steps down to deeper structural levels southward.

Extensional Structures

Cenozoic extensional faults overprint the thrust-belt structures in the greater Bannack area. The Muddy–Grasshopper fault is the most significant extensional fault mapped near Bannack, and is part of a regional fault system stretching over 90 km (56 mi), connecting the Grasshopper and Muddy Creek basins (Jänecke and others, 1999). Near Bannack, the fault places poorly consolidated Tertiary volcano-sedimentary deposits against deformed Paleozoic rocks (figs. 2, 5), with the trace of the Muddy–Grasshopper fault projecting beneath the Bannack townsite. The Muddy–Grasshopper detachment fault likely inverts an imbricate thrust or fault ramp within the Ermont thrust system. The timing of fault movement is poorly constrained but was likely synchronous with hanging-wall sedimentation recorded by the Eocene–Oligocene Medicine Lodge beds (Janecke and others, 1999). Lag deposits derived from the Hangmans Gulch gravel appear to drape the fault trace locally; therefore, it’s unclear if the fault system was active during the Miocene (see Sears, this volume).

MINERALIZATION AND MINING

The principal lode deposits near Bannack consist of replacement bodies in skarn derived from the metamorphism of limestone near Late Cretaceous granodioritic intrusions (Pearson and Childs, 1989). The ore bodies are tabular, pipe-like, irregular, and sometimes vein-like. The primary ore consists of gold-bearing quartz and pyrite with various skarn minerals including garnet, diopside, vesuvianite, specularite, and magnetite (Shenon, 1931). Minor amounts of chalcopyrite, galena, and sphalerite occur locally.

The main periods of mining in the Bannack area were early placer, hydraulic, and lode activity ca. 1862–1875, followed by the dredging of Grasshopper Creek from 1895–1902 (MTDEQ, 2007). The chief lode mines in the Bannack district were the Priscilla, Golden Leaf, Excelsior, Hendricks, and Gold Bug mines. The Golden Leaf group was the largest mine operation and produced about 75,000–125,000 oz of gold between the 1860s and 1955 (Pearson and Childs, 1989). Placer gold and silver mined along Grasshopper Creek were likely sourced from the local lode deposits. The placer deposits are located upstream from their sources, indicating that the flow of the proto-



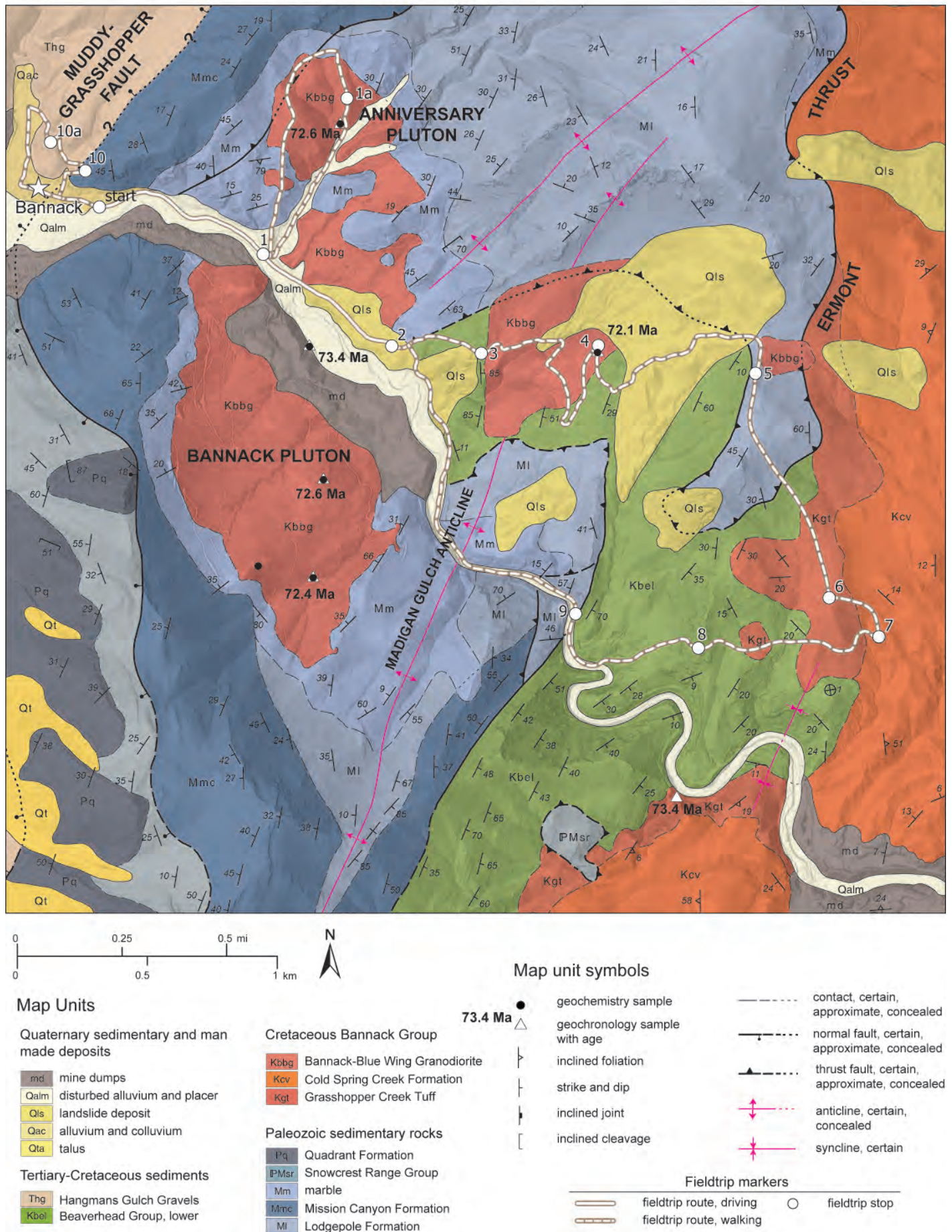
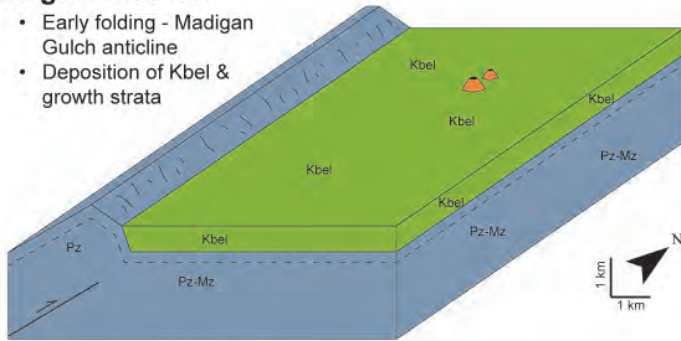


Figure 5. Geologic map of Bannack State Park area and field trip route. Modified from Mosolf (2017) and Mosolf and McDonald (2022).



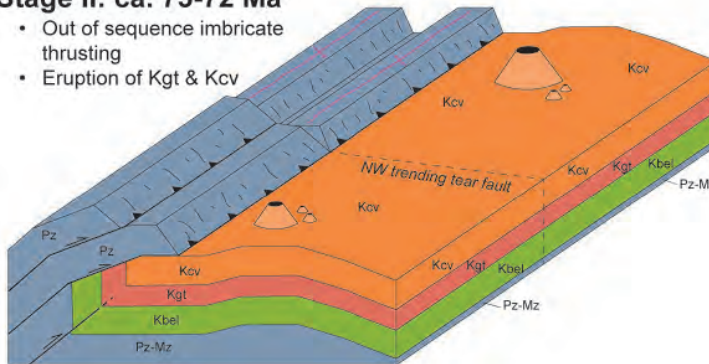
Stage I: ≥ 80 Ma

- Early folding - Madigan Gulch anticline
- Deposition of Kbel & growth strata



Stage II: ca. 75-72 Ma

- Out of sequence imbricate thrusting
- Eruption of Kgt & Kcv



Stage III: ca. 72 Ma

- Shortening stops
- Lava domes and granitoid intrusion emplacement
- Kbeu deposition

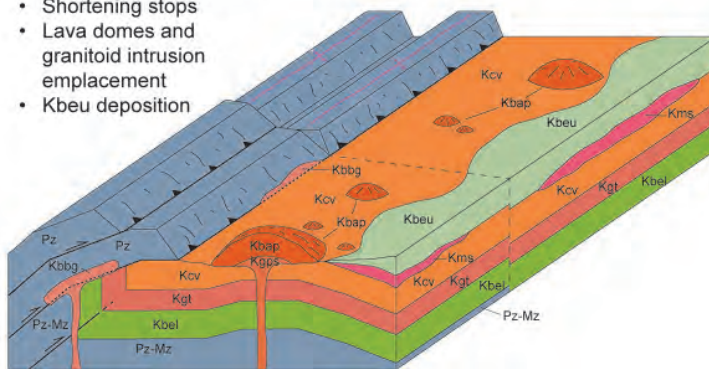


Figure 6. Three-stage cartoon block model illustrating the Bannack area's chronology of shortening, sedimentation, and magmatism. Units are defined in figure 5.

Grasshopper Creek reversed after placer deposition. See Sears (this volume) for a landscape model explaining how Bannack's placer gold is upstream of its sources.

The principal mines of the Blue Wing district, about 1 mi north of the town of Bannack, included the New Departure mine and the Blue Wing, Whopper, and Bannack Chief mines, later known as the Kent Group (Sassman, 1941). In contrast to Bannack, the Blue Wing district was mainly a silver district with lesser production totals (Geach, 1972).

See MTDEQ (2007) for a complete summary of mine activity in the Bannack area.

GEOLOGIC MAPPING AND RESEARCH IN THE BANNACK AREA

Previous geologic mapping in the Bannack area was conducted by Lowell (1965, 1:31,680 scale); Thomas (1981; 1:24,000 scale); Johnson (1986; 1:24,000 scale); Ruppel and others (1993, 1:250,000 scale); and Kalakay (2001). The Bannack 7.5' quadrangle (Mosolf, 2017, 1:24,000 scale), Eli Spring 7.5' quadrangle (Mosolf, 2021, 1:24,000 scale), Burns Mountain 7.5' quadrangle (Mosolf, 2024; 1:24,000 scale), the west half of the Dillon 30' x 60' quadrangle (Mosolf and McDonald, 2022; 1:100,000 scale), and the Salmon 30' x 60' quadrangle (Lonn and others, 2019, 1:100,000 scale) were mapped by the MBMG for the STATEMAP component of the United States Geological Survey (USGS) National Cooperative Geologic Mapping Program. The Beaverhead Group near Bannack was first mapped by Lowell (1965), and the sedimentology was described by Azevedo (1993). The Late Cretaceous volcanic stratigraphy and related plutons were defined and mapped by Ivy (1988) and Pearson and Childs (1989) and were further characterized by petrography, geochemistry, and geochronology studies conducted by the MBMG. Matoush (2002) mapped and described the Tertiary sedimentary units in detail.

ROAD LOG

The road log begins in the large parking lot near the east side of the Bannack town center (45.1606°N, 112.9923°W). Figure 5 shows the geology of the Bannack area and field trip stops. Mileage is cumulative and includes both driving and walking distances; coordinates are in the WGS84 datum. Drive east from the parking lot, pass the gate, and continue on the dirt road for 0.5 mi (0.8 km). Park in the large pullout near the fence and limestone and marble outcrops.

STOP 1—Hanging Wall Geology of the Ermont Thrust Fault and Crosscutting Granodioritic Plutons

0.4 mi (45.1590°N, 112.9843°W)

The clifty rock outcrops along the north side of the road are Paleozoic limestone beds that are variably metamorphosed to marble and calc-silicate skarn near granodiorite contacts (figs. 5, 7A). Pervasive alteration and tectonic deformation make rock identification at the formational level difficult in this area. Here, the protolith was likely the Mississippian Lodgepole Formation of the Madison Group. Unmetamorphosed occurrences of the Lodgepole Formation generally consist of gray to dark gray, thin- to medium-bedded, fossiliferous limestone with subordinate chert lenses and nodules, mudstone, and red calcareous siltstone. Limestone beds are generally planar to undulose and rhythmically bedded. Thinner beds and shaly partings are more common near the unit's base, while thicker bedding is more common near the top. The limestone beds commonly contain fossils of corals, crinoids, and bryozoans. Metamorphism commonly obscures primary sedimentary textures, but bedding planes are sometimes preserved.

Follow the overgrown two-track road northward for approximately 560 m (0.35 mi) and then climb the small knoll with exposures of the Anniversary pluton near the head of the drainage catchment.

STOP 1a—Anniversary Pluton

0.9 mi (45.1644°N, 112.9802°W)

The Anniversary pluton (fig. 7B) is one of several Late Cretaceous plutons in the Bannack area collectively mapped as the Bannack–Blue Wing granodiorite with U-Pb ages spanning 73.4–71.1 Ma (Mosolf, 2017; Mosolf and McDonald, 2022). Granodioritic plutons are typically recessive and susceptible to mass wasting, forming slopes and topographic depressions. Where exposed, the rock is commonly gray to greenish-gray diorite and granodiorite that is composed of fine-grained, equigranular to slightly porphyritic crystals of euhedral plagioclase, biotite, hornblende, and minor amounts of quartz and augite. The rock contains xenoliths of gneiss and ultramafic rock, locally. Granodioritic stocks are commonly associated with the alteration and mineralization of the surrounding country rock.

Looking up the slope, the Anniversary pluton appears to intrude a bedding parallel detachment fault in the Madison Group (fig. 8A). The granodiorite does not exhibit post-magmatic solid-state deformation, suggesting its emplacement occurred after movement on the fault had ceased. The rock outcrop at STOP 1a yielded a U-Pb age of 72.6 ± 0.4 Ma (fig. 7B). Foster and oth-

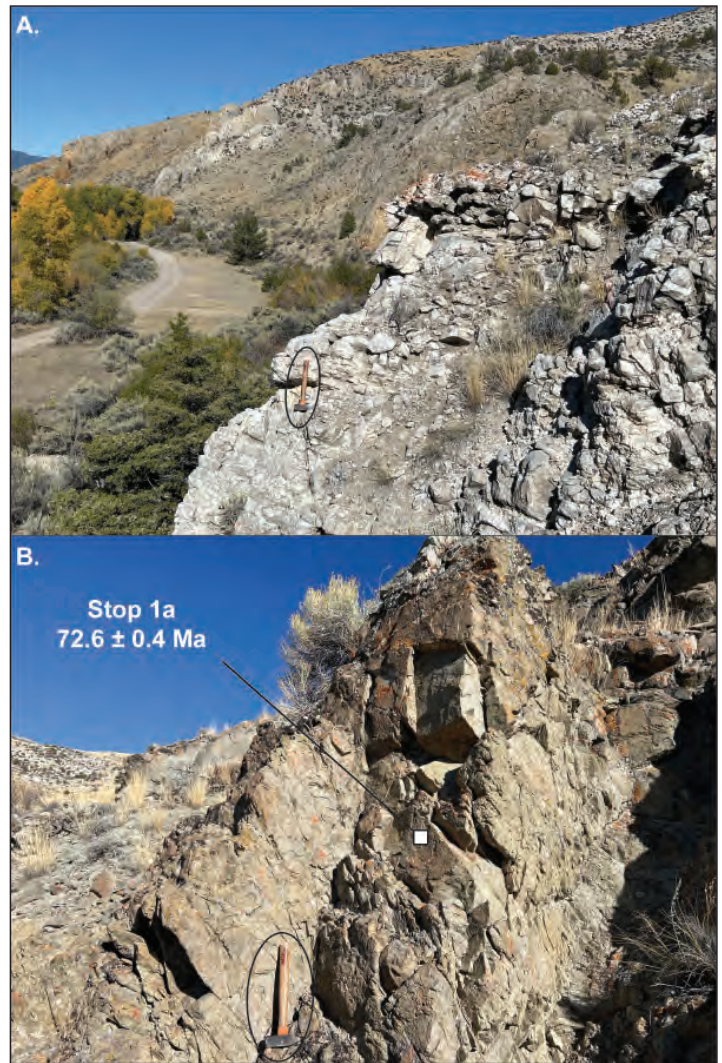


Figure 7. (A) View west of clifty Madison Group carbonates along the north side of Grasshopper Creek near STOP 1. The limestone beds have been variably metamorphosed to marble and skarn, giving the outcrops a variegated appearance. (B) Blocky, gray-green, and brown weathering granodiorite of the Anniversary pluton at STOP 1a. A sample from this stop yielded a U-Pb zircon age of 72.6 Ma.

ers (2012) reported a U-Pb age of 73.4 ± 0.8 Ma for the Anniversary pluton.

Looking south across Grasshopper Creek, an extensive network of mines, roads, and tailings piles trace the contact between the Bannack pluton and country rock (fig. 8B). The pluton is generally recessive and susceptible to mass wasting, forming hummocky slopes with little timber cover and rare rock outcroppings. Like the Anniversary pluton, the Bannack pluton was emplaced after local shortening slowed or ceased, intruding the Ermont thrust system. U-Pb ages from the Bannack pluton span 73.4–72.4 Ma (fig. 7B).

Continue to walk upslope approximately 180 m (600 ft) to the contact between the Anniversary pluton, Madison Group units, and associated marble and calc-silicate skarn. Contour southwest an additional 300 m





Figure 8. (A) View northwest of the Anniversaries pluton and STOP 1a. The pluton appears to intrude a small bedding-parallel thrust fault in the Madison Group. (B) View southeast of the Bannack pluton, the Anniversaries pluton, and STOP 1a are visible in the foreground. See map legend in figure 5 for symbol definitions.

(1,000 ft), following the intrusive contact and intermittent skarn deposits. Turn south at the first prominent spur ridge and follow it back to the vehicles, stopping to observe granodiorite, limestone, and marble outcrops along the way. Continue driving downstream another 640 m (0.4 mi) and park in the large pullout on a flat terrace near the road junction before the stream crossing.

STOP 2—Ermont Thrust System and Related Geology 2.0 mi (45.1558°N, 112.9780°W)

STOP 2 is near the trace of the Ermont thrust (fig. 5), which is covered by Quaternary sedimentary deposits and mine tailings at this location. To the west, the historic mine infrastructure and extensive mine dumps on the opposite side of Grasshopper Creek are part of the Priscilla and Golden Leaf mills (fig. 9). Hike up the steep gravel road to the east for 320 m (0.2 mi) to STOP 3, crossing the fault and traversing the footwall rocks.



Figure 9. Historic photo of the Golden Leaf Mill, viewed from the north. Remnants of the historic mill can be seen from STOP 2. Photo from the Beaverhead County Museum.

STOP 3—Lower Beaverhead Group 2.2 mi (45.1556°N, 112.9736°W)

Steeply tilted beds of limestone-clast conglomerate and interbedded sandstone of the Lower Beaverhead Group are well exposed near the bend in the road at STOP 3. The Lower Beaverhead Group is generally a reddish-brown to gray weathering, clast-supported conglomerate consisting of 1- to 10-m-thick (3–33 ft) beds of subangular to rounded pebbles and cobbles (fig. 10A). Clasts are mostly Paleozoic limestone with some quartzite clasts (<10 percent). Subordinate interbeds of maroon-red weathering sandstone are composed of carbonate-cemented, medium- to coarse-grained quartz, and chert sand (figs. 10A, 10B). Azevedo (1993) interpreted the lithofacies and stratigraphic architecture of the Lower Beaverhead Group in the Grasshopper Creek gorge to be an alluvial fan containing a progressive unconformity recording growth of the Madigan Gulch anticline further discussed at STOP 8. The Beaverhead Group is underlain by a discontinuous andesite tuff that yielded a U-Pb zircon age of 79.3 ± 1.3 Ma near Grasshopper Creek (Murphy, 2000; Kalakay, 2001); Laskowski and others (2013) reported a detrital zircon max depositional age of 77.9 ± 1.3 Ma for the Lower Beaverhead Group in the Bannack area.

Continue walking up the road 1.1 km (0.7 mi), crossing the contact between the Lower Beaverhead Group and an unnamed pluton. Green to brown weathering granodioritic rocks are exposed in several roadcuts between STOPS 3 and 4.



STOP 4—Bannack–Blue Wing Granodiorite and Mass Wasting Deposits**2.9 mi (45.1558 °N, 112.9679 °W)**

The unnamed granodioritic pluton crossed between STOPS 3 and 4 is age-correlative to the Anniversary and Bannack plutons discussed at STOP 1. Most of the outcrops observed in this area consist of brown, green, and purple, phaneritic to porphyritic granodiorite, with variable amounts of medium- to fine-grained plagioclase and hornblende (fig. 10C). The subcrop exposed in the road at STOP 4 yielded a U-Pb zircon age of 72.1 ± 0.4 Ma. Looking across the small drainage to the north, the pluton is covered by significant colluvial and mass wasting deposits, but it likely intrudes the Ermont thrust.

The hummocky hills to the north and east are large landslides with areal extents exceeding 3 km^2 (1.1 mi^2). Mass wasting is prevalent in the Grasshopper Creek gorge, obscuring the Ermont thrust's trace across the drainage. Continue hiking up the road to the pass, traversing the hummocks of a large landslide deposit. Turn south at the pass and hike off-trail to the rocky outcroppings of Mississippian carbonate rocks.

STOP 5—Overview of the Ermont Thrust Fault
3.5 mi (45.1549°N, -112.9602°W)

Stop 5 provides exceptional views of the Grasshopper Creek gorge, which forms an erosional window in the Ermont thrust sheet, revealing the tectono-magmatic anatomy of the thrust system. We are standing on a thin, partial klippe of Mississippian limestone in the hanging wall of the Ermont thrust (figs. 11, 12). Looking north, the Ermont thrust places Mississippian carbonate rocks over synorogenic volcano-sedimentary deposits of the Beaverhead and Bannack groups. The fault trace “Vs” towards the canyon bottom, where it is obscured by Late Cretaceous intrusions and mass wasting deposits (figs. 5, 11B).

Looking south across Grasshopper Creek, the east-vergent Madigan Gulch anticline, cored by the Lodge Pole Formation, is exposed in the opposite gorge wall. Steeply dipping beds of the Lower Beaverhead Group can be seen onlapping the eastern limb of the anticline and are interpreted to be growth strata (e.g., Azevedo, 1993). The Madigan Gulch anticline and Beaverhead Group will be further discussed at STOP 8. The trace of the Ermont thrust south of Grasshopper Creek is not easily identifiable, but it is likely obscured by the Bannack pluton, merges with the Madigan Gulch anticline, or is congruent with faults observed near the Beaverhead Group unconformity at STOP 9.

To the east, the Cold Spring Creek drainage is underlain by extensive gray-purple weathering Late Cretaceous volcanic deposits (fig. 11A). The prominent



Figure 10. (A) Steeply tilted conglomerate beds of the lower Beaverhead Group (Kbel) at STOP 3. The conglomerate beds are massive to crudely bedded, and dominated by Paleozoic limestone clasts. (B) Maroon-colored massive sandstone interbedded with limestone conglomerate characteristic of Kbel. (C) Light green-gray and weathering, fine-grained to slightly porphyritic granodiorite intruding Kbel between STOPS 3 and 4.



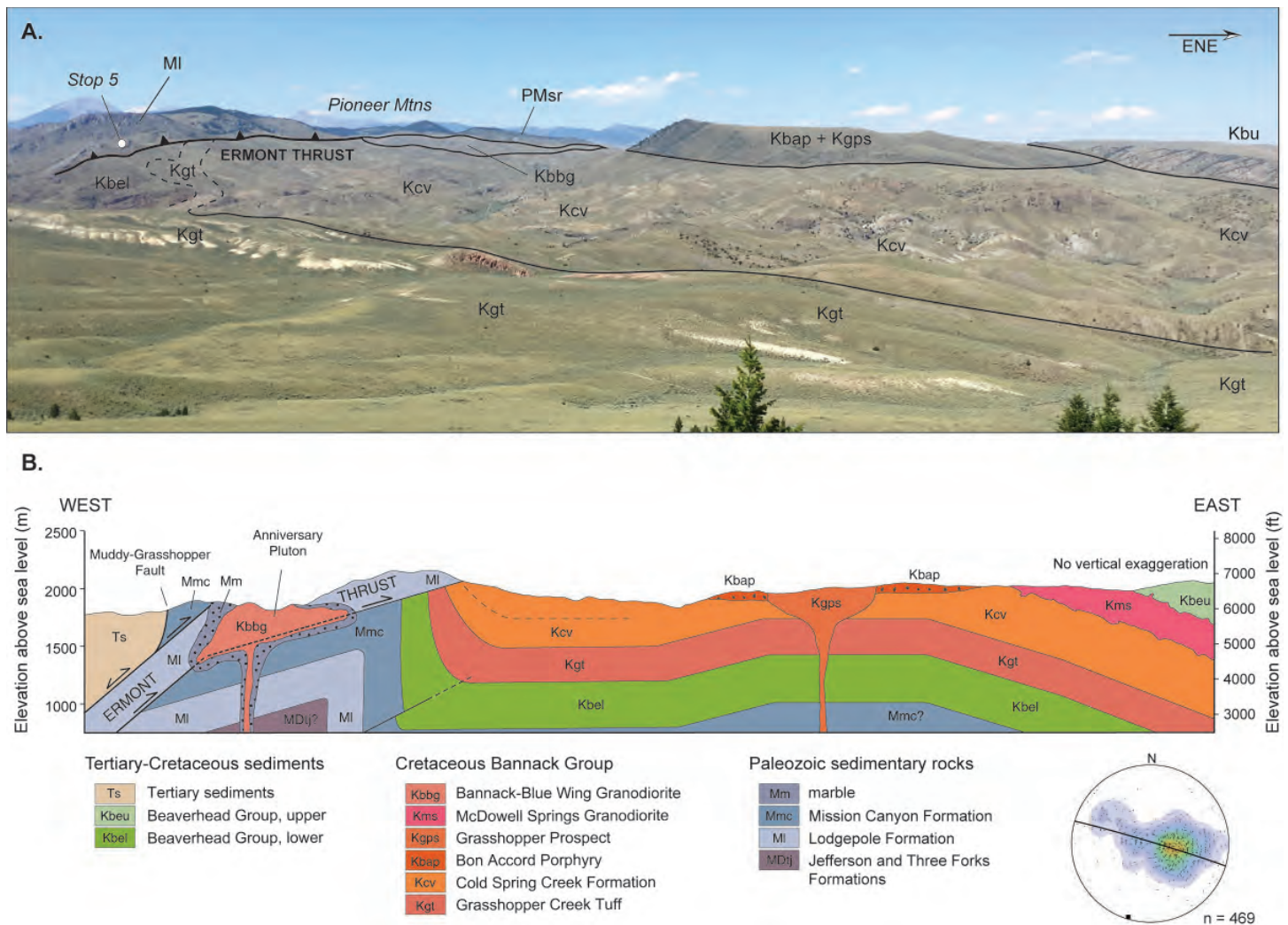


Figure 11. (A) View northwest of the Ermont thrust's trace north of Grasshopper Creek gorge. The thrust places cliffy outcrops of Madison Group carbonates over gray slopes underlain by recessive volcanic deposits of the Cold Spring Creek Formation. (B) Schematic cross-section representing the area's structural geology approximately 400 m (0.25 mi) north of Bannack and the Grasshopper Creek gorge. The approximate section line is shown in figure 1. The inset stereonet is a lower-hemisphere projection of poles to Paleozoic-Mesozoic bedding measured in the Bannack 7.5' quadrangle (Mosolf, 2017). The data are plotted on an equal-area stereonet and fit with a Kamb contour. The cylindrical best fit of poles to bedding is a near-vertical profile plane with a gently south-southwest-plunging pole (black box) representing the general axis of folding in the Ermont thrust system.

mountain east of Cold Spring Creek is buttressed by a remnant rhyo-dacitic lava dome complex (Bon Accord Porphyry and Grasshopper Prospect). The Late Cretaceous magmatic units are discussed further at STOPS 6 and 7.

Continue walking south along the ridge for another 200 m (600 ft) to the saddle. Turn southeast and carefully descend the steep southeast-facing slope, crossing the trace of the Ermont thrust. Follow the short spur ridge to the small rounded hill.

STOP 6—Grasshopper Creek Tuff 4.1 mi (45.1471°N, 112.9565°W)

STOP 6 is underlain by float and subcrop of the Late Cretaceous Grasshopper Creek tuff (Kgt). The tuff is the oldest informal Bannack Group unit mapped in the area, with U-Pb zircon ages spanning 75.3–73.3 Ma. Where exposed elsewhere in the Bannack area, the tuff

unit consists of white, light gray, yellowish gray, to pale red deposits of thin-bedded vitric tuff and massively bedded crystal-lithic rhyolitic tuff containing varying proportions of sanidine, quartz, and plagioclase crystals and accidental fragments of volcanic and sedimentary rock in a matrix of glass shards that are mostly zeolitized and silicified. A compaction foliation formed by the collapse of pumice fragments is commonly observed; welded ignimbrite intervals occur locally. Tuff beds are locally interfingered with intervals of breccia, conglomerate, and immature sandstone primarily composed of volcanic debris and ash, with some detritus derived from Proterozoic–Paleozoic units. The stratal thickness of the Grasshopper Creek Tuff exceeds 300 m (985 ft). At this location, the tuff beds were tilted eastward by movement on the Ermont thrust.

Turn east-southeast and walk downslope 280 m (920 ft) to the rocky knob of dacite porphyry at STOP 7.





Figure 12. (A) View east-northeast of the Ermont thrust and the Grasshopper Creek gorge. North of the creek, the Ermont thrust places carbonate beds of the Mississippian Lodgepole Formation (MI) over eastward-tilted red beds of the lower Beaverhead Group (Kbel). Cretaceous volcanic rocks of the Bannack Group can be seen in the background. (B) East-vergent folds in rhythmically bedded limestone of the Lodgepole Formation in the hanging wall of the Ermont thrust, northwest of STOP 5.

STOP 7—Cold Spring Creek Formation, Bon Accord Porphyry, and Grasshopper Prospect 4.3 mi (45.1457°N, 112.9541°W)

At STOP 7, dark weathering dacite porphyry forms blocky outcrops resistant to weathering. The porphyry is massive, highly fractured, and possibly brecciated. Phenocrysts of plagioclase, hornblende, biotite, and rare quartz are readily identifiable. The dacite observed here is part of a complex, intertonguing sequence of variegated porphyritic lava flows, volcanic agglomerate, and ash-flow tuff mapped as the Late Cretaceous Cold Spring Creek Formation (Ivy, 1988; Mosolf and McDonald, 2022). Porphyritic lavas are generally dark colored and contain crowded phenocrysts of plagioclase, orthopyroxene, clinopyroxene, hornblende, biotite, and quartz in a plagioclase-phyric groundmass with abundant Fe-Ti oxides. Intercalated volcanic agglomerates and breccias occur as massive, greenish-gray intervals 3–5 m (10–16 ft) thick composed of poorly sorted, subrounded-to-subangular clasts of

andesite–dacite porphyry. Discontinuous tuff interbeds 2–10 m (6–33 ft) thick contain broken phenocrysts of plagioclase, hornblende, rare quartz, and accidental volcanic clasts in a eutaxitic groundmass of devitrified and flattened glass shards. Tuff intervals commonly contain varying amounts of pumice fragments and accidental clasts of porphyry lava. Ivy (1988) reported $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende ages of approximately 80–76 Ma for this unit. U-Pb zircon ages span 73.6–71.7 Ma. Thickness as much as 490 m (1,600 ft).

Looking eastward across the drainage catchment, the Cold Spring Creek Formation is overlain and possibly intruded by the rhyo-dacitic Bon Accord Porphyry and Grasshopper Prospect (fig. 13A). The Bon Accord Porphyry forms steep columns rising above the more subdued topography of the other Late Cretaceous units, creating a doughnut-shaped body of rock 3 km (1.9 mi) wide in plan view. Blocky dacite outcrops contain phenocrysts of plagioclase, hornblende, biotite, sanidine, and embayed quartz in a devitrified glass matrix with some plagioclase microlites. Abundant xenoliths of hornblende gneiss and chlorite actinolite schist are up to 1 m (3.3 ft) in diameter. The rock exhibits a pronounced foliation that consistently dips inward towards the Grasshopper Prospect (Kgps). The Bon Accord Porphyry yielded U-Pb zircon ages of 72.6–71.4 Ma.

The Grasshopper Prospect is a white weathering, friable, hypabyssal rhyolite intrusion that forms a circular

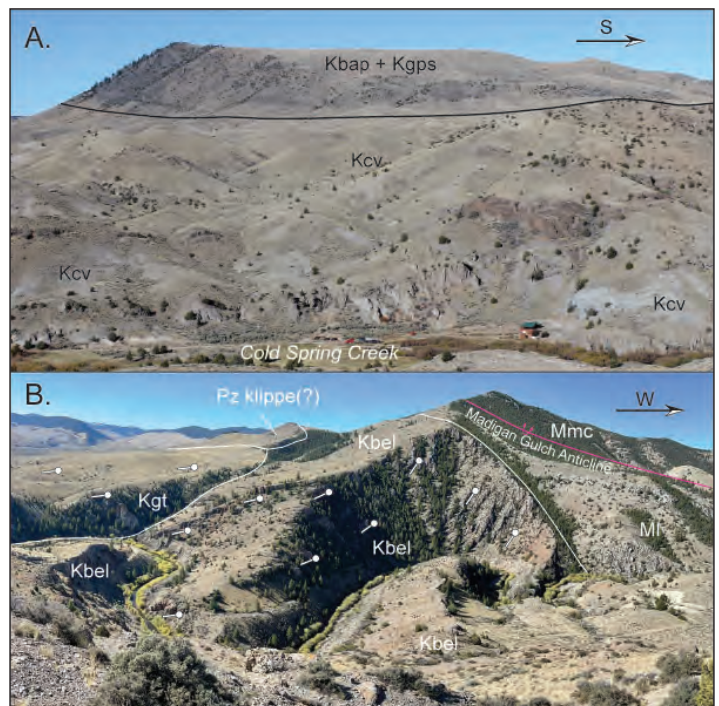


Figure 13. (A) View east of the Cold Spring Creek volcanic sequence (Kcv) and Bon Accord Porphyry–Grasshopper Prospect lava dome complex. (B) View south across the Grasshopper Creek gorge of the Madigan Gulch anticline and growth strata in the Lower Beaverhead Group and Grasshopper Creek Tuff. Apparent dip of bedding is outlined by ball and bars.



depression surrounded by dacite columns of the Bon Accord Porphyry. The original texture of the rock is commonly obscured by pervasive quartz-sericite-pyrite alteration (Meyer, 1980) and abundant veins of fine-grained quartz. Where preserved, the rhyolite has a seriate to porphyritic texture, and contains phenocrysts of quartz, altered plagioclase, orthoclase, and mafic minerals altered to Fe-Ti oxides. U-Pb zircon ages from the Grasshopper Prospect span 73.0–72.2 Ma, nearly indistinguishable from the Bon Accord Porphyry ages. It's unclear if the Grasshopper Prospect and Bon Accord Porphyry are distinct units or if the Grasshopper Prospect is hydrothermally altered dacite porphyry (Uribe-Mogollon and Maher, 2019).

Eastward, the Cold Spring Creek and Bon Accord Porphyry units are overlain by poorly consolidated deposits of the Upper Beaverhead Group, and intruded by the McDowell Springs granodiorite. The Upper Beaverhead Group consists of poorly lithified, clast-supported conglomerate beds dominated by well-rounded quartzite cobbles, likely derived from the Belt Supergroup. Detrital zircon max depositional ages span 72.7–65.9 Ma (Laskowski and others, 2013; Garber and others, 2020; Mosolf and McDonald, 2022).

The Bannack and Beaverhead Groups are intruded by the McDowell Springs granodiorite throughout the Armstead Hills area, mainly east of the Ermont thrust trace. The granodiorite is a gray-green to brown weathering porphyry containing phenocrysts of plagioclase, pyroxene, biotite, hornblende, and rare quartz in a groundmass of plagioclase and altered mafic microlites with abundant magnetite. A U-Pb zircon age of 71.8 ± 0.5 Ma was obtained for this unit. Given its appearance, composition, and age, the McDowell Springs granodiorite is likely genetically related to the Bannack–Blue Wing Granodiorite.

Walk 0.8 km (0.5 mi) west to STOP 8, carefully contouring the steep, cliffy slopes of the Lower Beaverhead Group.

STOP 8—Lower Beaverhead Group and Madigan Gulch Anticline

4.7 mi (45.1453°N, 112.9629°W)

The Lower Beaverhead Group is well exposed in the cliffy outcrops at STOP 8, and consists of channelized limestone-clast conglomerate and interbedded maroon-colored sandstone, similar to the beds viewed at STOP 3. Looking south across Grasshopper Creek, the Lower Beaverhead Group can be seen in contact with the asymmetric, east-vergent Madigan Gulch Anticline (fig. 13). The Mississippian Lodgepole Formation is the deepest stratigraphic level exposed in the core of the fold. Down-plunge projections of surface data show the Madigan Gulch anticline is a fault propagation fold

involving crystalline basement (Kalakay, 2001), similar to the basement-cored Armstead anticline to the south.

The Lower Beaverhead Group rests in angular unconformity on the deformed Mississippian limestone and marble units exposed in the eastern limb of the Madigan Gulch anticline. Faulting has occurred along and parallel to the unconformity at this location. The apparent dip of bedding and intraformational unconformities visible within the Lower Beaverhead Group decreases from west to east (fig. 13B). At least two discrete dip domains are separated by angular unconformities that record the progressive tilting of a single alluvial fan sequence during growth of the Madigan Gulch anticline (Azevedo, 1993; Kalakay, 2001). Radiometric dating of an underlying andesite tuff and detrital zircon from the Beaverhead Group constrain the onset of folding deformation to approximately 80 Ma.

Large slabs (up to 0.5 km²; 0.2 mi²) of Paleozoic rock can be seen resting on the Late Cretaceous units east of the Ermont fault's main surface trace (fig. 13B). Assuming the slabs are not massive landslide blocks, the displaced Paleozoic units are klippen isolated from the main Ermont allochthon, constraining minimum fault heave to approximately 2.5 km (1.6 mi). The thrust dips eastward beneath the klippen, and therefore it is either folded or has been rotated by an unrecognized normal fault located to the east of the klippen (Kalakay, 2001).

Carefully contour west around the nose of the ridge and then descend the steep slope westward towards Grasshopper Creek, crossing the jeep road *en route*. Carefully climb down the steep cut bank and mine dumps to the creekside jeep road. Follow the road north 280 m (900 ft) to STOP 9.

STOP 9—Fault or Angular Unconformity?

5.2 mi (45.1466°N, 112.9690°W)

STOP 9 provides an opportunity to inspect the contact between the Lower Beaverhead Group and complexly deformed Paleozoic limestone and marble units. Here, conglomerate and sandstone beds of the Beaverhead Group are near-vertical (fig. 14) and are in contact with deformed Paleozoic strata previously interpreted to contain at least two generations of faults and folds (Johnson, 2007). Lowell (1965) first mapped the contact separating the Paleozoic and Cretaceous units as a fault, possibly the southern continuation of the Ermont thrust. Johnson (1986, 2007) reinterpreted the contact as depositional angular unconformity formed by two deformational events comprising early Laramide foreland deformation later overprinted by the Sevier-style Ermont thrust system (Johnson and Sears, 1988). Mosolf (2017) mapped the contact as a folded part of the Ermont thrust system formed by a single progressive deformation.



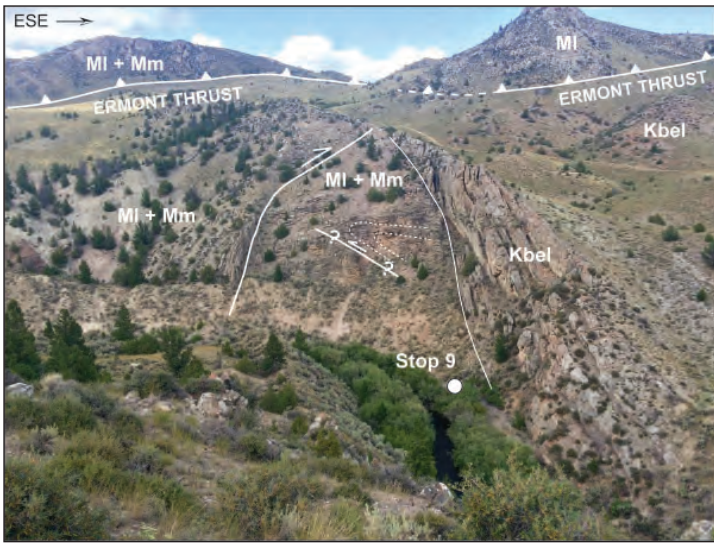


Figure 14. Overview of STOP 9 showing the contact between deformed Paleozoic strata and Lower Beaverhead Group. Note the west-vergent or rotated fold in the carbonate units. The trace of the Ermont thrust can be seen below the skyline, where Mississippian carbonates are placed over the Lower Beaverhead Group.

Follow the road back to the vehicles, hiking approximately 1.5 km (0.9 mi) along Grasshopper Creek. Drive back to the field trip's starting point for optional STOP 10.

OPTIONAL STOP 10—Muddy–Grasshopper Fault and Tertiary Sediments

7.3 mi (45.1619°N, 112.9930°W)

Walk west through the Bannack town center for approximately 170 m (560 ft), then turn north following the signage to the horse trail. Continue on the trail for approximately 30 m (100 ft) and then turn east, hiking off-trail to the small adit and tailings pile visible on the hillside, approximately 75 m (250 ft) from the trail.

The Muddy Creek–Grasshopper fault zone consists of a series of westerly dipping, listric normal faults that can be traced for nearly 90 km (56 mi), extending from the Grasshopper Creek basin north of Bannack to the Muddy Creek basin bordering the Tendoy Mountains in far southwest Montana (Jänecke and others, 1999). In the Bannack area, the fault system bounds the eastern margin of the Grasshopper Basin, a structurally complex, east-tilted half-graben filled with Tertiary sediments with minor volcanic intercalations (Matoush, 2002; Lonn and others, 2019). To the north near Badger Pass, the Muddy–Grasshopper fault dips approximately 30°, placing poorly consolidated Tertiary sedimentary deposits in the hanging wall against complexly deformed Paleozoic rocks in the footwall. The fault progressively steepens to approximately 60° southward near Bannack, possibly inverting an imbricate fault or ramp related to the Ermont thrust system (fig. 11). Looking south across Grasshopper Creek, a complex network of faults deforms the Paleozoic units, some

of which are likely extensionally reactivated imbricate thrusts (fig. 5).

At STOP 10, we are near the contact separating complexly deformed Paleozoic strata from poorly consolidated sedimentary deposits of the late Miocene(?) Hangmans Gulch gravel (Mosolf and McDonald, 2022). The surface trace of the Muddy–Grasshopper fault projects under STOP 10; however, it's unclear whether the contact observed here is a fault or depositional contact. Brecciated limestone and a light-colored fault gouge are visible in the small adit and associated tailings pile.

Return to the horse trail and continue hiking northward towards the cemetery. Turn left (southwest) at the trail junction and walk 30 m (100 ft) to the town overlook.

STOP 10a—Tertiary Basin-Fill Deposits: Hangmans Gulch Gravel

7.8 mi (45.1629°N, 112.9947°W)

The Hangmans Gulch gravel consists of unconsolidated, poorly organized deposits of angular to round cobbles and boulders. Clasts are mostly Proterozoic–Paleozoic limestone and quartzite sourced from the adjacent footwall of the Muddy–Grasshopper fault, likely as alluvial fan and debris flow deposits. Diagnostic coffin-shaped boulders of quartzite approximately 1 m (3 ft) in length are common to the north towards Badger Pass (fig. 15), with some boulders exceeding 3 m (10 ft).

Follow the trail northwest into Hangmans Gulch to inspect gravel deposits, and then follow the trail south back to the Bannack town center.

END OF ROAD LOG



Figure 15. Hangman's Gulch gravel unit near Bannack. Recessive slopes are commonly draped with poorly sorted Paleozoic boulders with diameters exceeding 2 m (6.5 ft).



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CHALK BLUFF: FROM CORDILLERAN SYNTECTONIC PLUTONISM THROUGH CENOZOIC FORMATION OF THE UPPER BIG HOLE VALLEY BASIN

Colleen Elliott

Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

INTRODUCTION

Chalk Bluff, which is a prominent landmark about 17 mi (27 km) north of Wisdom, MT, along State Highway 43 (figs. 1, 2), has a key exposure of contact relationships within the Cordilleran Foolhen Plutonic–Metamorphic Complex. The Foolhen Complex (FC) dominates the West Pioneer Mountains and has these components (Elliott and others, 2025):

1. Cretaceous sillimanite-bearing metamorphic to migmatitic rocks with Proterozoic and Carboniferous through Cretaceous sedimentary protoliths (fig. 1, KYm; fig. 3);
2. Gneissic to nonfoliated biotite $\pm$ hornblende-bearing granodiorite (Kgd, $\sim$ 72–74 Ma; fig. 4);
3. Gneiss of Foolhen Ridge (Kgn, $\sim$ 73 Ma); and
4. Paleogene biotite $\pm$ muscovite-bearing granitoids ($\sim$ 68–63 Ma).

Elliott and Lonn (2021) and Elliott (2022) concluded that Cretaceous plutonism and peak metamorphism were syntectonic. Evidence for this includes the following observations:

1. Gneiss of Foolhen Ridge (Kgn) has the same zircon U–Pb age and locally has gradational contacts with Kgd;
2. Migmatite in the FC has the same U–Pb zircon age as Kgd; and
3. Intrusions in the FC are elongated NW–SE, parallel to dominant NW-trending foliation in Kgd.

The evidence also includes the nature of the Kgd–KYm contact at Chalk Bluff, where Kgd grades from weakly foliated near the highway to gneissic over a short vertical distance. The gneissic Kgd appears to interfinger with KYm towards the top of the ridge (fig. 2).

Another key exposure found at Chalk Bluff is the eastern edge of the paleo-Big Hole Basin. The valley wall is represented by crystalline rocks—Kgd and KYm. Across a steep contact, Oligocene Renova Formation-equivalent sediments dip moderately west into the Big Hole Valley. The exact contact is not exposed, and has been interpreted as depositional or tectonic by different mappers. Hanneman (1987) and Ruppel and others (1993) mapped the contact as

depositional, presumably because the Oligocene sediments dip towards the valley, opposite of what would be expected in the case of faulting. Roe (2010) and Elliott (2022) interpreted the sediment/basement contact as a fault based on its abrupt nature and the presence of a possible gauge zone. Chalk Bluff is capped by quartzite–cobble gravel (fig. 5) that has been mapped across the basement–Renova contact without offset.

The upper Big Hole Valley basin contains as much as 16,000 ft (4,900 m) of Lower Eocene through Quaternary volcanic and sedimentary strata, and valley-fill immediately west of the field trip area is as much as 6,600 ft (2,000 m) thick (Roe, 2010). This remarkable thickening over a short distance must be explained if we are to understand the history of the valley.

Doughty and Sheriff (1992) suggested that the Big Hole Basin formed where crustal extension in the Bitterroot valley stepped *en echelon*-style to the southeast. Janecke (1994) proposed that the current Big Hole Basin lies within a regional, north–south-trending rift zone about 100 km wide in which Paleogene gravitational collapse created linear basins filled with alluvium and mass wasting deposits derived from high-relief rift shoulders. Beginning with the first description of the Anaconda Metamorphic Core Complex by O'Neill and others (2004), most authors implicitly or explicitly assume that the Big Hole Basin was formed by extension on the Anaconda Detachment Zone (ADZ), forming a syntectonic, asymmetric, hanging wall basin.

Elliott (2022) concluded that basin depth was increased by movement on east–southeast-, north–northeast-, and north-trending faults that cut through the basin.

Recently, Schwartz and others (in review) compiled detrital zircon, provenance, paleocurrent, and sediment facies data for a large part of SW Montana, and concluded that the Big Hole Basin was initially formed by incision during the waning stages of Cordilleran orogenesis. Evidence for this includes the existence of outcrops in which >50 Ma sediments rest unconformably on exhumed Late Cretaceous and Paleocene intrusive rocks. This implies that the crystalline rocks surrounding the Big Hole Basin were exposed before regional extension began—around 53 Ma before present (Foster and others, 2010). Synorogenic incision is typical of modern orogenic belts where contraction-driven uplift is accompanied by drainage along linear river valleys perpendicular to the regional shortening direction.



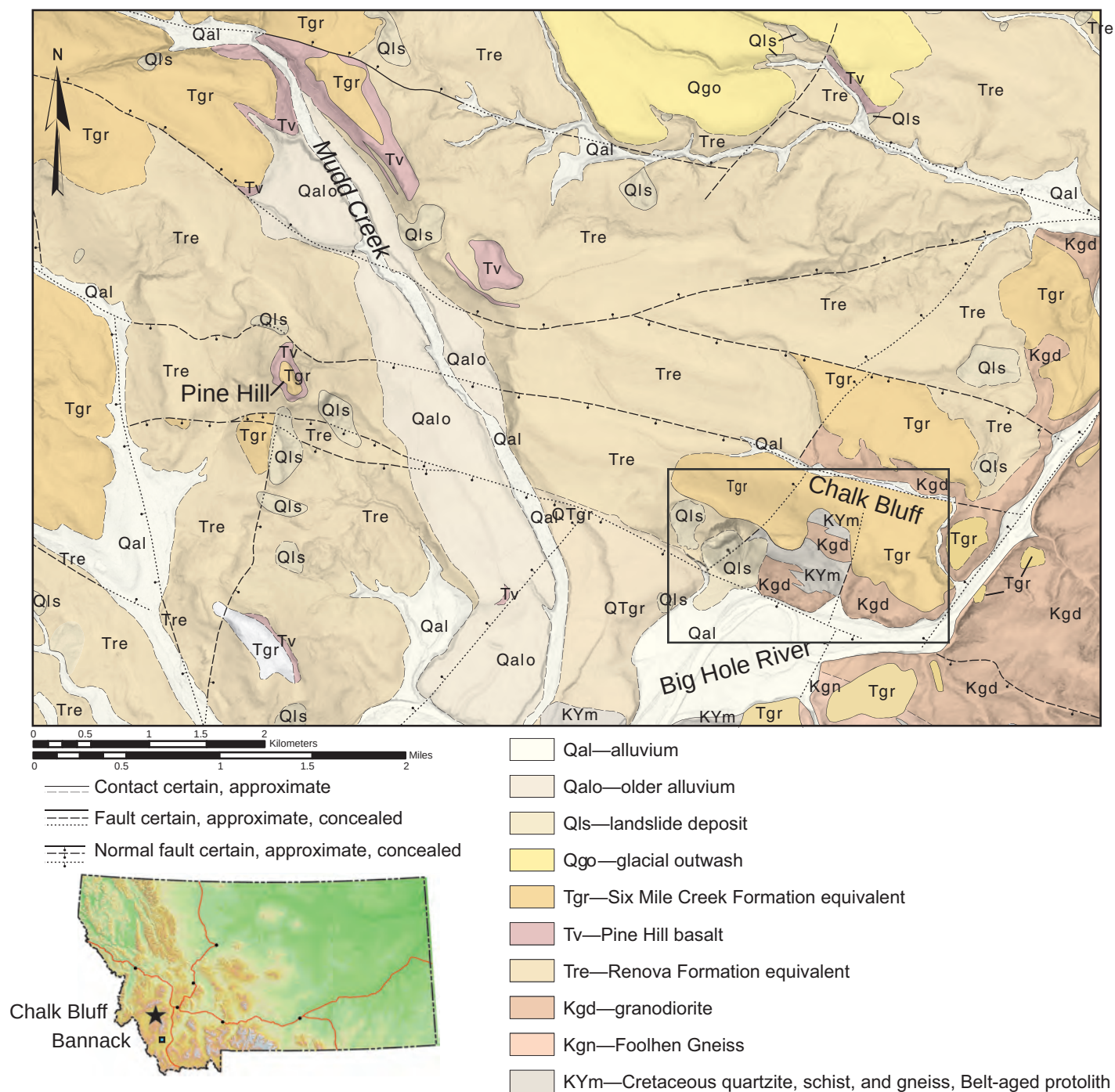


Figure 1. Geologic setting and location of Chalk Bluff (after Elliott and Lonn, 2021). The box shows the location of figure 2.

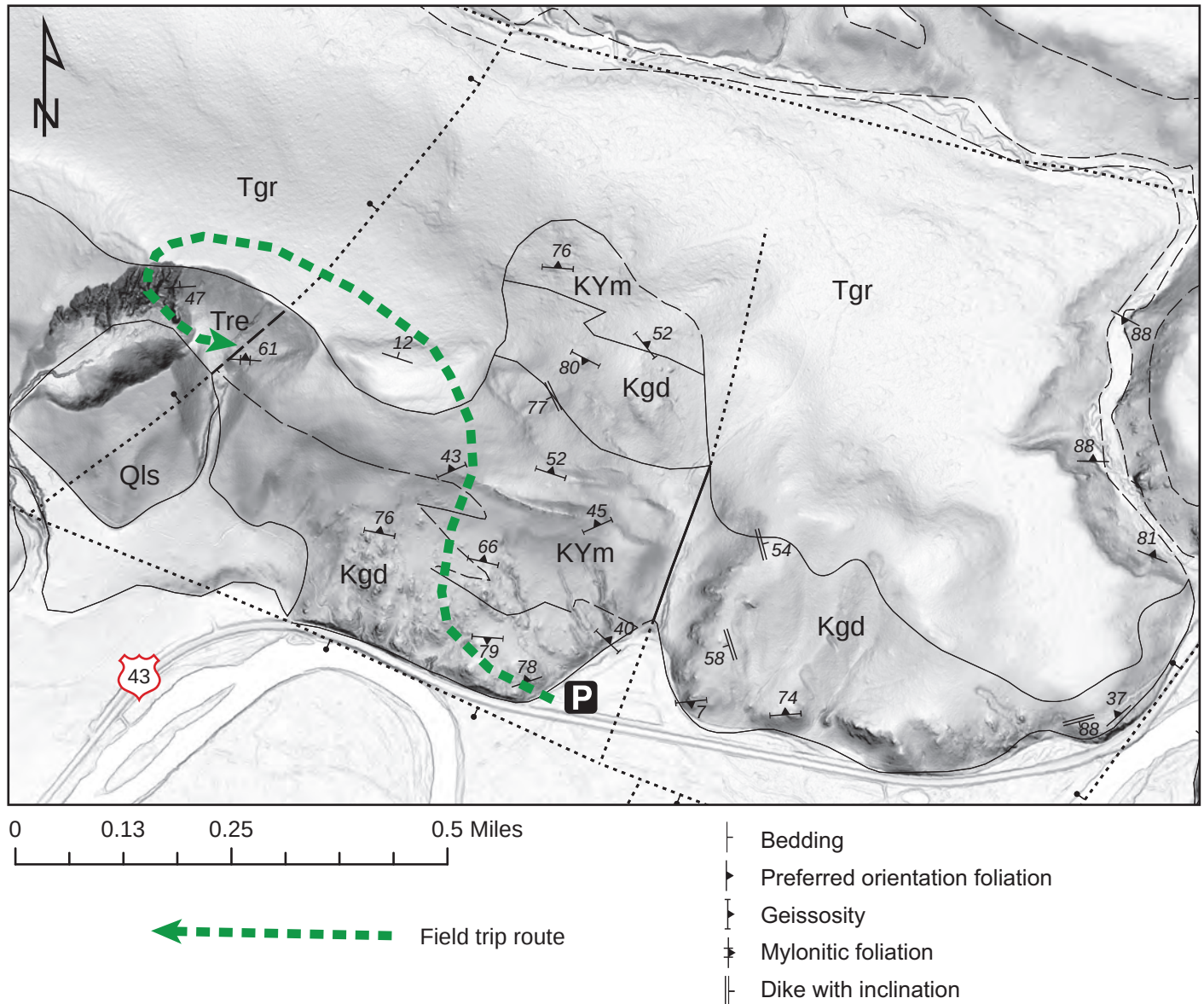


Figure 2. Lidar image of Chalk Bluff with geology overlain.

FIELD TRIP GOALS

The questions we will try to answer on the Chalk Bluff field trip include;

1. What was the role of syntectonic magmatism in the Foolhen Plutonic–Metamorphic Complex (FC) of the West Pioneer Mountains?
2. What are the relationships between Eocene Renova Formation sediments, the adjacent crystalline rocks, and the overlying cobble gravel?
3. Why does the Big Hole Basin deepen from zero at Chalk Bluff to as much as 2,000 m (6,600 ft) within 4 km (2.5 mi) of Chalk Bluff?

GEOLOGIC SETTING

Rock types we will see at Chalk Bluff (fig. 2):

Tgr Gravel (Miocene)—Cobble and boulder gravel with matrix of sand and silt. Clasts are sub- to well-rounded, with low sphericity, and commonly have tan weathering rinds (fig. 5). Clasts are mostly metamorphic rock types and granite, and dominated by quartzite. Tgr is not well-exposed, but is characterized by smooth, rounded topographic surfaces. Roe (2010) describes lateral equivalents of Tgr as laterally and vertically variable, with channels of sand and gravel interlayered with thinly bedded, finer sediments. Correlates with Sixmile Creek Formation. Miocene age comes from tentative fossil identifications and detrital zircon spectra that indicate a minimum age of 11.6 Ma (Roe, 2010). Thought to be up to 60 m (200 ft) thick.





Figure 3. KYm at Chalk Bluff is thought to have a Mesoproterozoic protolith and a Cretaceous metamorphic age. (A) Thin layering disrupted by asymmetric foliation boudinage and quartz-filled symmetric boudinage (center of photo). View is down the long axis of the boudin necks. (B) Quartzitic layers with what appear to be partial melt segregations—the rusty blebs on the right half of the photo—formed along foliation boudinage planes. Rock hammer for scale.



Figure 4. Well-foliated but not gneissic Kgd granodiorite.

Tre Renova Formation equivalent (Oligocene)—White and gray silt and clay (fig. 5) with rare sand and gravel lenses. Regionally these sediments are pale when dry, but olive green, brown, and gold when wet, and may have ochre to maroon Liesegang banding. Gray paleosols occur near the top in some places. Clay intervals weather with a popcorn texture that is typical of the Renova Formation. At Chalk Bluff, a single tooth fragment was tentatively dated as Oligocene (Hanneman and Nichols, 1981; Roe, 2010). U-Pb zircon ages for tuffs within Renova-equivalent sediments in the Big Hole Basin range between 26 and 31 Ma (Roe, 2010; Elliott and others, 2025).

Kgd Granodiorite (Cretaceous)—Gray, medium- to coarse-grained granular to porphyritic granodiorite and tonalite named the Nichols granodiorite at Chalk Bluff (Roe, 2010), though it is part of the Foolhen Mountain Pluton (fig. 4; Elliott and Lonn, 2021). Contains 5–35 percent biotite, hornblende, or both biotite and hornblende and 5–56 percent potassium feldspar (Truckle, 1988). Hornblende appears to increase in abundance towards the south and west. Mafic inclusions are abundant. The granodiorite varies from massive with no foliation, to weakly foliated (PDO), to strongly gneissic, particularly near contacts with KYm. Contains light gray, fine-grained aplite dikes with up to 15 percent plagioclase and 5 percent biotite. Also contains metasedimentary inclusions of all sizes, from hand sample to map scale. Roe (2010) obtained a U-Pb zircon age of 71.8 ± 2.0 Ma for a sample from Chalk Bluff. Zircons from a biotite-rich, unfoliated sample of Kgd from the adjacent Foolhen Mountain quadrangle yield a U-Pb zircon age of 73.89 ± 0.39 Ma (Elliott and Lonn, 2021). Kgd is unit Kgdf in Elliott and others (2025).

KYm Quartzite, schist, and migmatite gneiss (Mesoproterozoic: Cretaceous metamorphic age)—

The metasedimentary rocks at Chalk Bluff (fig. 3) are narrowly banded and so fine-grained that individual minerals can be hard to discern. The greater FC is composed of fine- to medium-grained amphibolite-facies quartz–potassium feldspar–plagioclase–muscovite–biotite–sillimanite metasedimentary gneiss and schist that is commonly migmatitic. Migmatite neosome is fine- to medium-grained biotite–muscovite granite locally injected or segregated in foliation-parallel sheets (lit-par-lit), as blobs in fold hinges, or parallel to the axial plane of folds. KYm is typically complexly deformed, with several fold generations. Elliott and Lonn (2021) obtained detrital zircon spectra that suggest the protolith was Mesoproterozoic Belt Supergroup. KYm is Ysqg in Elliott and others (2025).

ROAD LOG

To the trailhead:

From Bannack, drive north to the junction of Highway 278. Turn west. Connect to Highway 43 at Wisdom (47.4 mi) and continue north another 15.6 mi. At 63 miles the highway crosses the Big Hole River, and Chalk Bluff is visible straight ahead. Continue driving another 1.7 mi (64.7 mi) and find rough parking on either side of the road where the shoulders are wide enough, and by the field access gate on the north side of the road. There is more parking another $\frac{1}{2}$ mi further along, beside the river.

This field trip involves a roughly 1-mi hike with roughly 200 ft of elevation gain. There is very little shade or shelter and no bathroom facilities.





Figure 5. Tertiary sediments on the west side of Chalk Bluff and the Big Hole Valley. View is to the west. Red-stained gravel (Tgr) overlies clay-rich white sediments (Tre), and is raveling downhill from the flat surface at the top of the bluff.

Start at the gate in the fence at the base of the bluff. The biotite–hornblende granodiorite here (Kgd) is about 72 Ma based on U-Pb ages reported by Foster and others (2010) and Roe (2010). Near the base of the hill the granodiorite has a foliation defined by a preferred dimensional orientation (PDO) of biotite and hornblende grains. As we walk up the hill, the PDO becomes a gneissosity, foliation-parallel elongate xenoliths of metasedimentary rock become more common, and we transition into thinly layered quartzite-dominated KYm.

The contact between Kgd and the metasediments trends NW and is not quite parallel to the WNW-trending gneissosity in Kgd. One interpretation is that the gneissosity is the axial plane foliation for very tight folds in the Kgd–KYm contact. Another possibility is that Kgd was injected lit-par-lit into the strongly layered country rock during intrusion.

Things to look for:

- Are Kgd and KYm really interfolded? Can we trace the contact on an outcrop scale?
- Kinematic indicators. If this is a shear zone, we should be able to see lineations and asymmetric structures generated by rotational strain.

Continue climbing above the metasedimentary rocks into the gravel that caps the hill (Tgr). Follow the base of Tgr west to where it overlies the white, fine-grained Oligocene sediments (Tre)

Things to look for:

- Can we find bedding in the gravel?
- Does Tgr really drape the unconformity?

Climb carefully down a landslide scarp into the underlying sediments.

Things to look for:

- What was the depositional environment for Tre? Are they basin-margin or axial sediments?
- Are the sediments *in situ*, or are they part of a larger landslide complex?

Make your way carefully east towards the contact between Tre and the KYm and Kgd basement. What is the nature of the contact? Roe (2010) found a seam of red clay gouge along the contact. If it is a fault, why don't the Tre beds dip towards it?

When all the questions have been answered, the field trip is over and we can make our way back to the cars.

ACKNOWLEDGMENTS

I am grateful to Hans Humbert of the Dell Bacon Ranch Company for allowing land access for this field trip. Thanks to Jeff Lonn for his review.

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GEOHAZARDS AND GROUNDWATER IN VIRGINIA CITY, MONTANA

Mandy Willingham, Andy Bobst, and Yann Gavillot

Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana

This field trip provides an opportunity to see landslides, landslide-controlled springs, Quaternary faults, and points of interest around the historic mining town of Virginia City. We will stop at vista points highlighting local geohazards, visit the town's main spring to discuss local hydrology and how it relates to mass wasting, and go for a hike in the landslide complex that forms the Axolotl Lakes Wilderness Study Area. Figure 1 shows the field trip route.

INTRODUCTION

Virginia City is best known for its history as one of Montana's oldest and richest placer mining districts (Ruppel and Liu, 2004), eventually becoming the territorial capital of Montana. It was founded soon after placer gravels were discovered in Alder Gulch around 1863, and more than 2.5 million oz of gold and 350,000 oz of silver were produced there (Shawe and Weir, 1989). Evidence of this mining activity is present and well preserved in Virginia City today, from the original boomtown buildings to the placer workings along Alder Gulch.

The region is not only interesting from a mining perspective, but also exhibits multiple prehistoric landslides that control modern hydrology of the town's main water source. Landsliding in this area is commonly associated with a package of volcanic units referred to as the Virginia City Volcanic Group and appears to correlate with the failure of thick deposits of the tuff units such as the Alder Gulch Tuff (Tagt), Virginia City Tuff (Tvct), and Nevada City Tuff (Tnct), overlain by mafic and intermediate lavas of the Virginia City Formation (Tvcl, Tvcm, Tvcu; fig. 2). Quaternary faults are also visible from the area, including the Ruby Range northern border fault, the Ruby Range western border fault, and the southern extent of the Tobacco Root fault.

Virginia City's main drinking water source is a spring within a massive landslide deposit that contains large displaced volcanic blocks. The spring water discharges at the contact between the tuff and the overlying fractured lava flow units (fig. 3). Numerous other springs exist throughout the landslide deposit at similar geological contacts in the gravitationally slid blocks. In 2017, the Montana Bureau of Mines and Geology (MBMG) conducted a groundwater investigation in conjunction with a STATEMAP bedrock mapping project to better understand the local geology controlling the groundwater system (Bobst and others, 2022; Mosolf, 2021).

ROAD LOG

From Bannack, drive about 54 mi on MT-278 E and MT-41 N to Jensen Park in Twin Bridges. Reset your odometer here.

STOP 1—Jensen Park 0.0 mi

45.54500, -112.33520

From this park, you can see the Tobacco Root Mountains to the east, Pioneer Mountains to the west, and Ruby Mountains to the south.

The Road Agent Trail passed near here, connecting Bannack to Virginia City. It was terrorized in 1863 by a gang of "road agents" targeting the valuables of travelers. The gang was eventually stopped by vigilantes from Bannack and Alder Gulch. Today, Montana Highways 41 and 287 closely parallel the old Road Agent Trail (MDT, 2025).

The former Montana state orphanage is also visible from the parking lot to the west (also known as Montana Children's Center, founded in 1894). It was shut down in 1975, as foster care became a more popular alternative to orphanages. Throughout its operating days, the Center housed over 5,000 children (Amestoy and Mott, 2023).

Driving towards Virginia City looking south at the Ruby Range, notice the range-front Quaternary fault, the Ruby Range northern border fault (RRNBF) on your right, with triangular facets and young alluvial fans (fig. 4). Notice how the RRNBF displays a more youthful landscape from a geomorphology perspective, with active deposition and young fans sourced from a watersheds with very steep gradients, which contrasts with the less active Ruby Range western border fault (RBWBF) west of the Ruby Range (fig. 5). Recent cosmogenic dating on alluvial fans indicates fault activity along the Ruby Range has occurred in the last ~40 kyr (Parker and Gavillot, 2024; Gavillot and others, 2024). The 2007 Sheridan earthquake with magnitude of 4.6 was consistent with a focal mechanism on a NNW-trending normal fault dipping 55–60° to the NE, and was likely caused by the Northern Ruby Range northern border fault (fig. 5; Stickney, 2022).

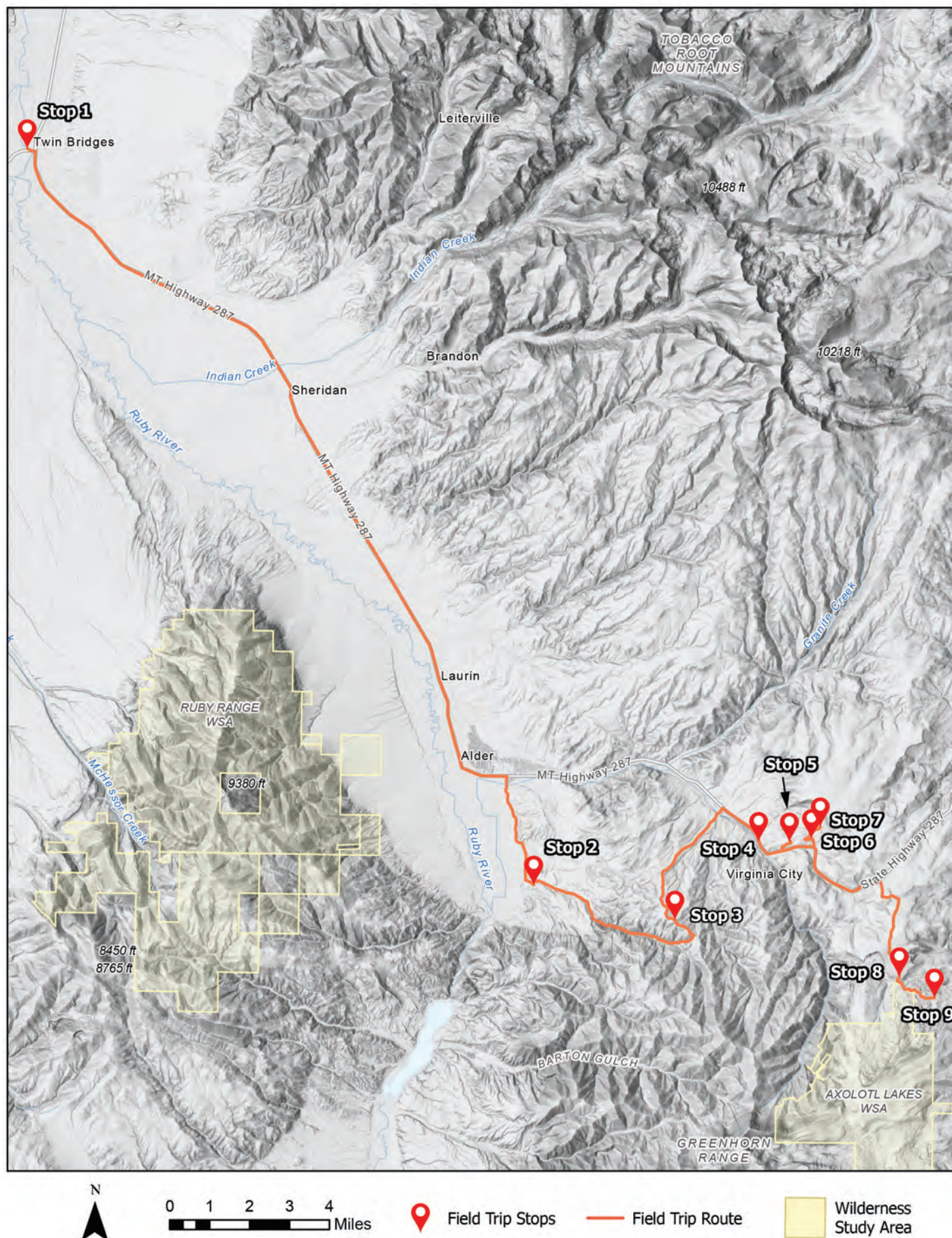


Figure 1. Regional map showing the field trip route and stops. The Road Log begins at Jensen Park in Twin Bridges, MT, and ends at the Twin Lakes pullout near Axolotl Lakes Wilderness Study Area.



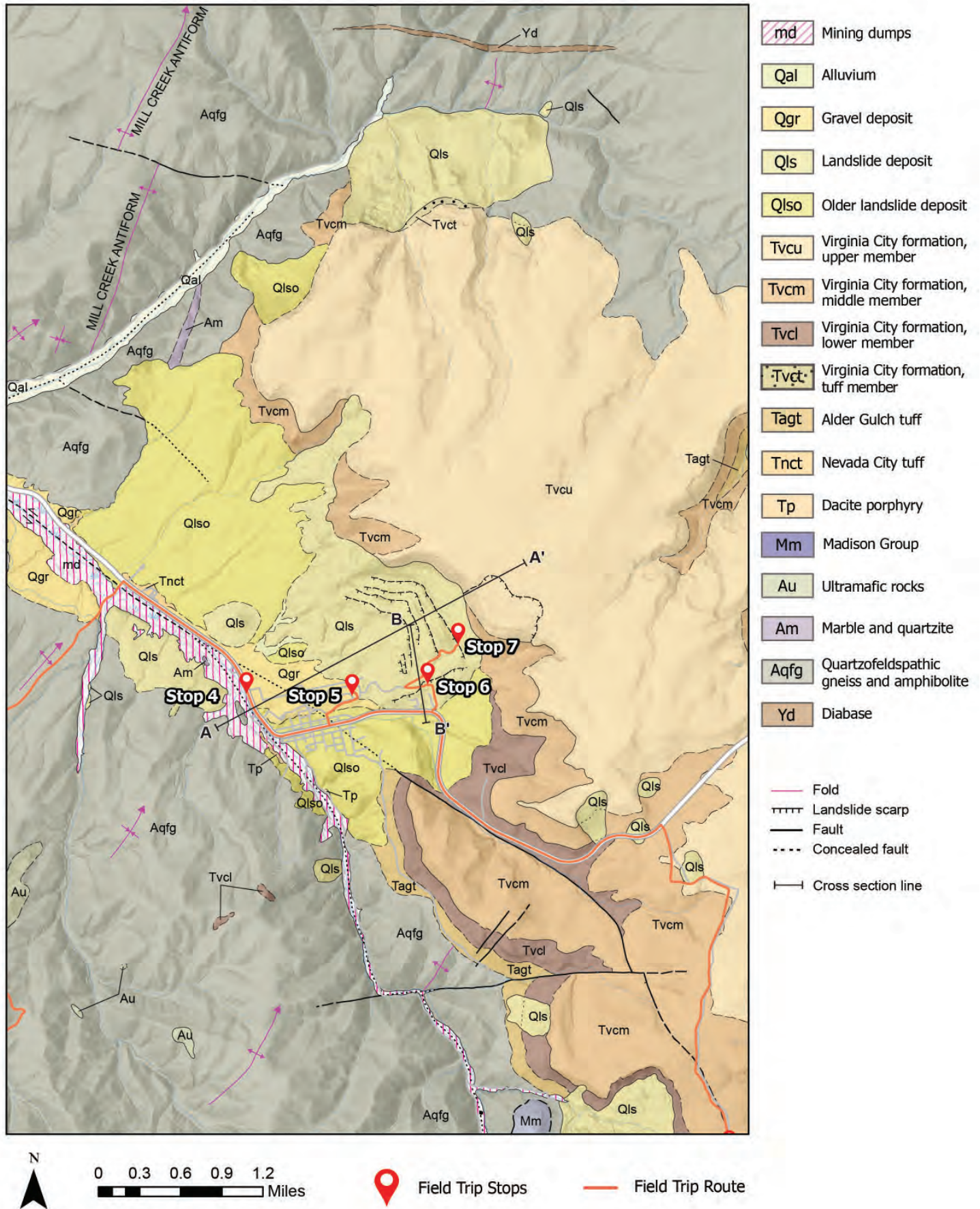


Figure 2. Geologic map of the Virginia City area, modified from Mosolf (2021). Cross-sections are shown in figure 3.



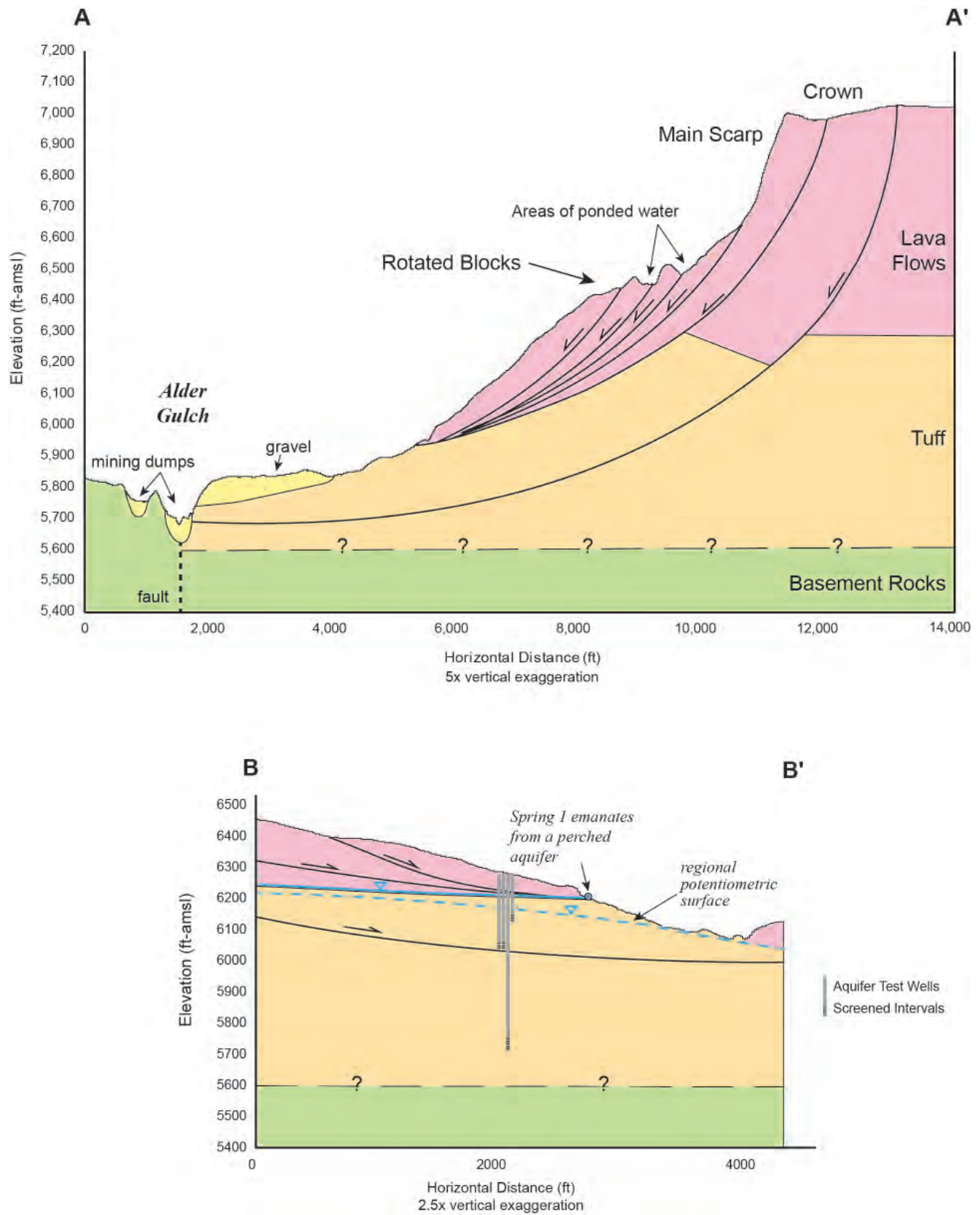


Figure 3. Interpretive cross-section through landslide deposits, from Bobst (2022). Cross-section locations indicated in figure 2.



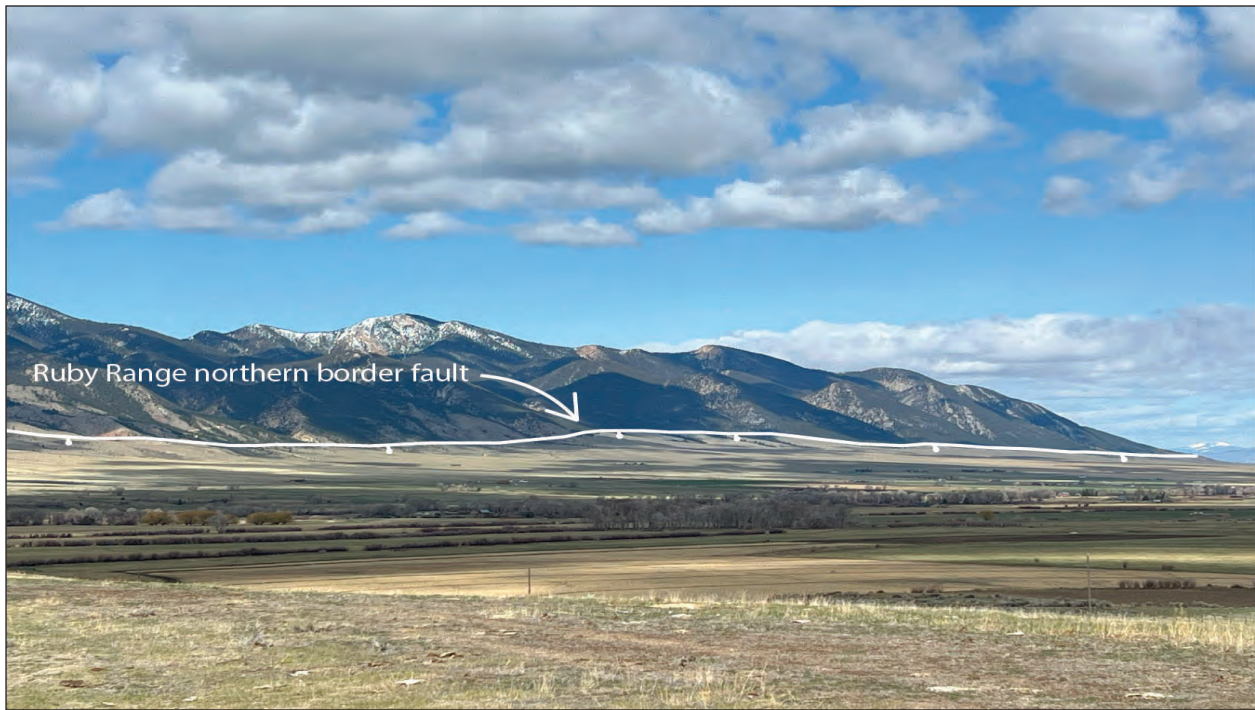


Figure 4. Looking northwest from STOP 2, highlighting the Ruby Range northern border fault.

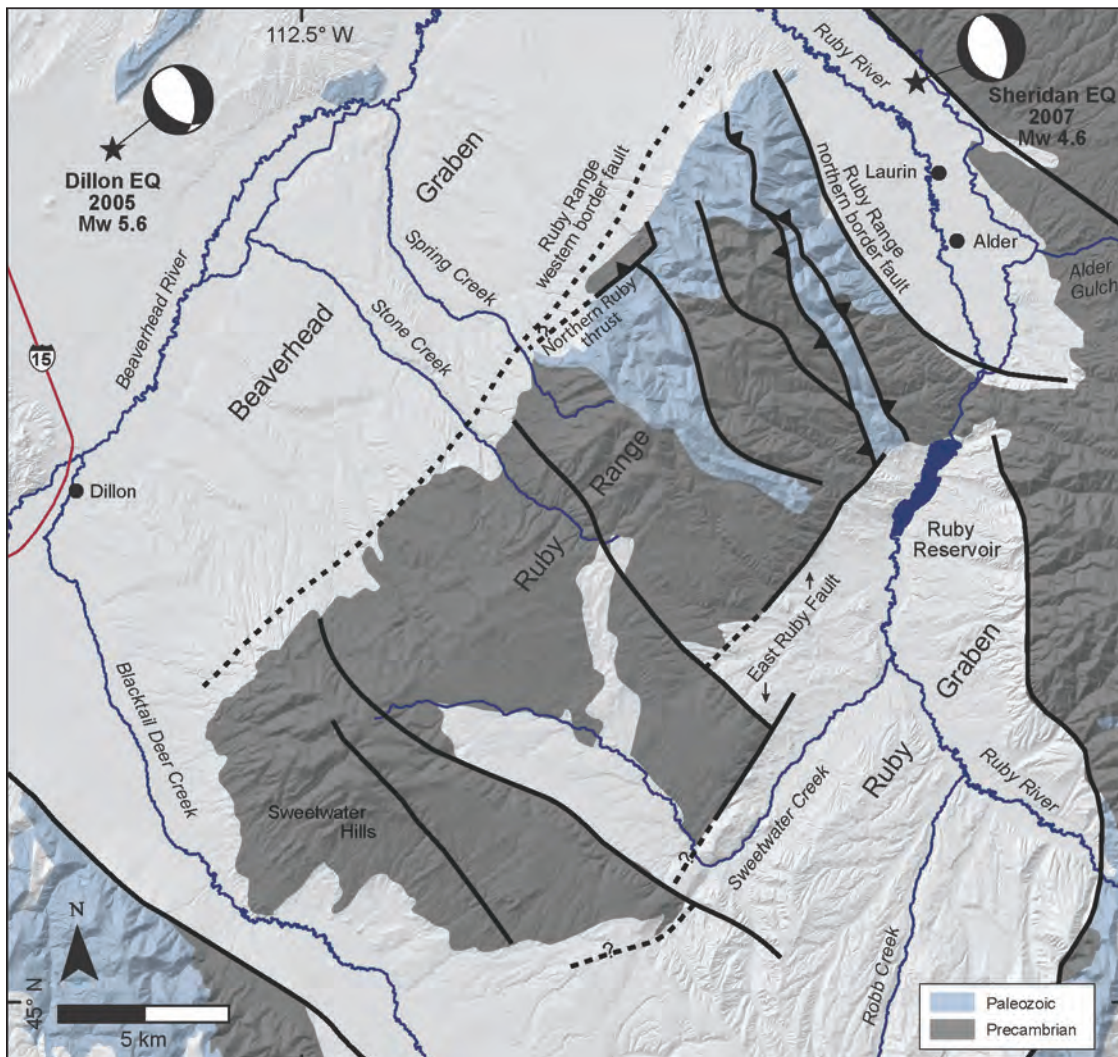


Figure 5. Regional overview of Quaternary faults and recent seismic activity, from Stickney (2022).



Drive south on MT-287 for 20.8 mi, passing through Sheridan and Alder.

20.8 mi Turn right onto Anderson Lane and drive for 2.8 mi.

23.6 mi Sharp left onto Williams Creek Road and drive for ~0.3 mi, past an intersection and uphill a bit on a two-track near the power line.

STOP 2—Williams Creek Road Overlook

45.28065, -112.06928

The southeastern section of the RRNBF range front looks quite steep with a sharp lineament between mostly Precambrian basement in the footwall and Tertiary units in the hanging wall. Fault displacement is likely not as youthful as the northern section of the RRNBF, given the deeply incised landscape and lack of preserved Quaternary deposits. Around Sheridan, movement along the Tobacco Root Fault transfers to the RRNBF.

Looking north, notice the steep triangular facets along the range front with steeply dipping Paleozoic units in the footwall and broad alluvial fan surfaces in the

hanging wall. Some of the most prominent beds are the massive cliff-forming limestone of the Mission Canyon Formation. The RRNBF is interpreted to be partly controlled at depth on reactivated older thrust faults (Parker and Gavillot, 2024).

Continue driving east on Williams Creek Road for 5.6 mi.

STOP 3—View of Virginia City Landslide 29.2 mi

45.26856, -111.99658

To the northeast is a great view of the main Virginia City landslide in the distance (fig. 6). Notice the scale of this feature and the location of Virginia City below the headscarp and built on landslide deposits. The headscarp is steep and densely forested. Can you see other internal landslide features? There is also another nice view of the RRNBF to the northwest.

Figure 7 shows the main Virginia City landslide in lidar imagery, enhanced to show relative local relief, with some of the internal features labeled.

Continue driving on Williams Creek Road for another 3.7 mi.

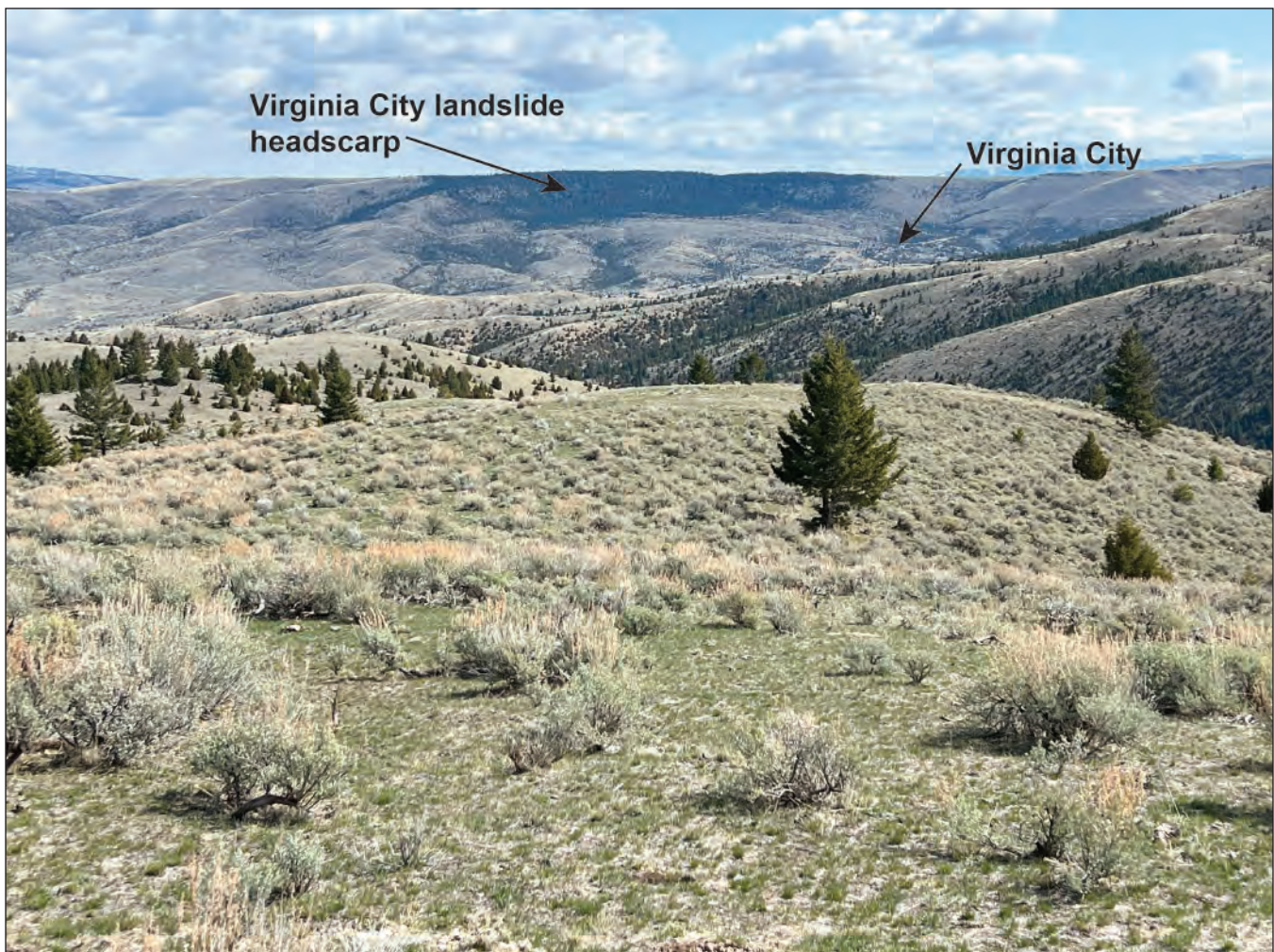


Figure 6. View northeast from STOP 3, showing the Virginia City landslide headscarp in the distance. Note Virginia City's location on top of the landslide deposit.



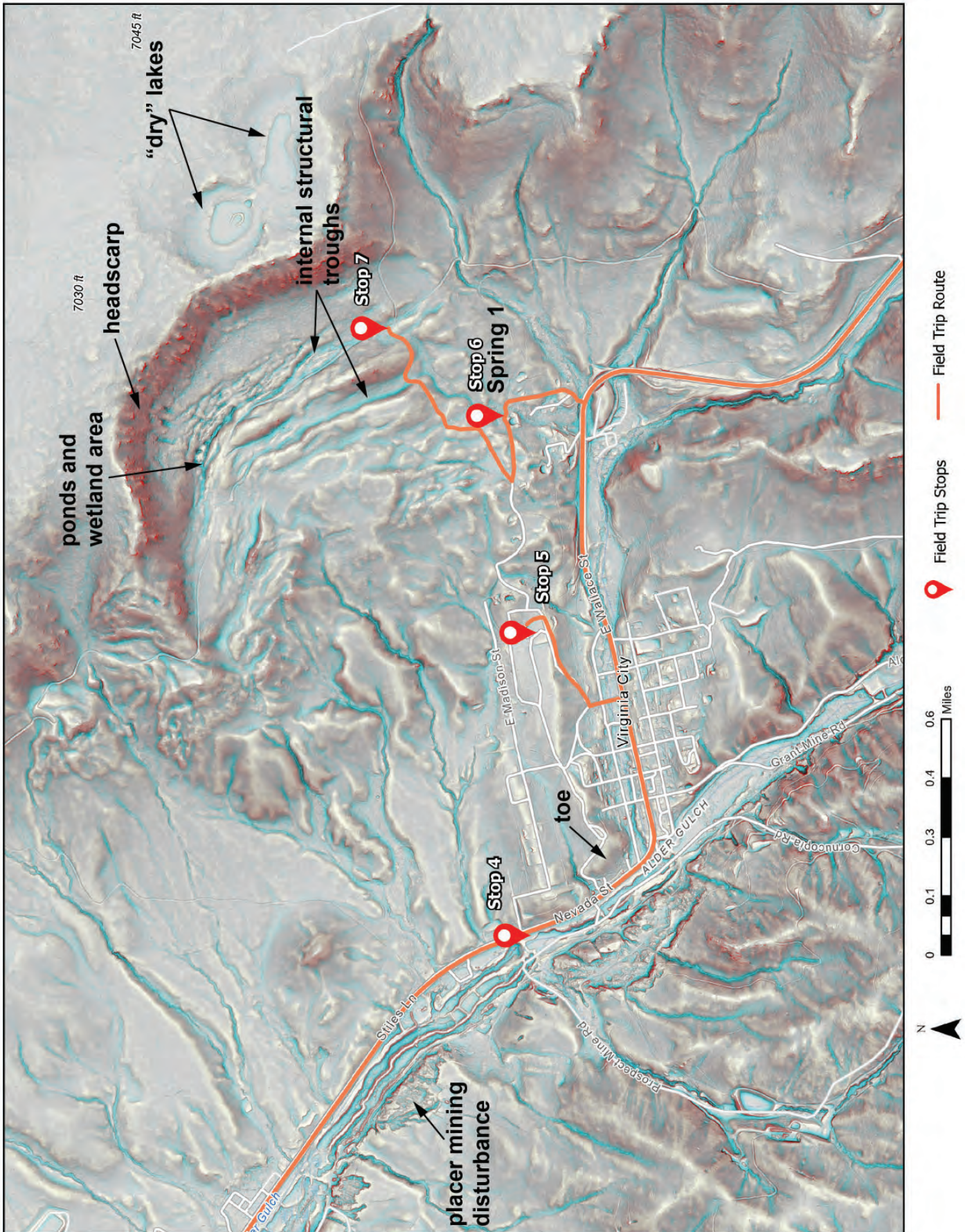


Figure 7. The main Virginia City landslide, as seen in lidar and enhanced with a simple local relief model and presented in a red image relief map (RRIM).



32.9 mi Turn right onto MT-287 and drive for 1.3 mi, then turn right into the parking lot.

When turning onto MT-287, Nevada City will be visible to the left. Nevada City was a mining ghost town that was restored by private owners between 1945 and 1978. Currently it functions as a living historical museum managed by the Montana Heritage Commission.

STOP 4—Alder Gulch 34.2 mi

45.29740, -111.95400

This is Alder Gulch, and the Alder Gulch Fault follows the creek (fig. 8). On the northeastern side of the creek, Virginia City volcanics have been modified by mass movement, and on the southwest side Archean gneiss is exposed. Take a moment to look around at some of the cobbles in the creek. You will find both lithologies represented. This is also a great spot to collect some Alder Gulch “rubies” present in the garnet-rich gneiss.

This is also in the toe of the main Virginia City landslide, though it is difficult to see from this vantage. During 2018, students from the Montana Tech Geo-

physical Engineering Field Camp conducted seismic surveys here to get a sense of the thickness of saturated unconsolidated sediments, and to evaluate the potential for an emergency backup water supply for the city from the Alder Gulch alluvium (Bobst, 2022; Bobst and others, 2022). It appears that the unconsolidated sediments are over 100 ft thick and are a viable target for a well. There are concerns, however, with water quality from the mining waste, and the treatment requirements for using groundwater that is in direct communication with surface waters.

Virginia City was heavily placer-mined. The ponds here are remnants of that activity. Placer workings are visible throughout the span of Alder Gulch Creek. Though the area was heavily placer-mined, the lode source was never identified. It was possibly eroded away (Ruppel and Liu, 2004). Remediation of the area has been partially completed, as represented by these ponds.

Turn right out of the parking lot and drive south on MT-287 for 0.8 mi.



Figure 8. View of Alder Gulch looking south from STOP 4. The pond in mid-frame is a remnant of placer mining activity, and has been reclaimed as a recreation area. In the left frame, you can see the toe of the Virginia City landslide. Alder Gulch fault, a strike-slip fault of unknown age, runs through this gulch, separating Virginia City volcanics to the east from Archean basement to the west.

35 mi Turn left onto Spencer Street and drive 0.1 mi. Turn right on East Douglas Road/Old Stage Road and drive 0.2 mi. Park around the back (northeast) side of the ballfield.

STOP 5—Ballfield 35.3 mi

45.29730, -111.93820

From here, there is a good view of the headscarp and landslide deposits to the northeast, and flat topography to the west (under the ballfield and cemetery) created by Quaternary gravels deposited when relative river level was higher.

Slightly uphill to the northeast is a quarry within a large displaced dacite porphyry block. This dacite and other similar blocks found throughout the landslide deposit have yielded U-Pb zircon ages 50.0 ± 0.4 Ma to 49.6 ± 0.7 Ma (Mosolf, 2021).

Return the way you came, by traveling west down East Douglas Road/Old Stage Road 0.2 mi.

35.5 mi Turn left onto Spencer Street and drive for 0.1 mi. Turn left onto East Wallace Street/MT-287 and drive 0.8 mi. Turn left onto Edgerton Street and drive 0.2 mi; Edgerton Street becomes East Douglas Street. Park near the water tank and pump house.

STOP 6— Virginia City Spring 136.6 mi

45.29860, -111.92690

This spring (Spring 1) is the main water source for Virginia City (fig. 9). It is a contact spring, emanating from the contact between overlying lava flows and underlying tuff. While this area is highly disturbed by landsliding, many large blocks remain intact, including this one. There are multiple springs throughout the landslide mass, at contacts between lava and tuff. It is thought that the groundwater here is sourced from

precipitation infiltration into the landslide deposits, which are highly fractured and permeable. In 2017, the MBMG conducted a groundwater investigation to help constrain extent of the source area and identify alternative sources of water for the community to augment water supply during times of high tourist visitation (Bobst and others, 2022). Behind the spring house are outcropping lava flows, and as the slope inflects and shallows to the south there is tuff. Students from the Montana Tech Geophysical Engineering Field Camp conducted resistivity profiles, very low-frequency electromagnetic surveys, and a seismic survey in this area to aid in understanding the distribution of the geologic units and fracture patterns (Bobst, 2022; Khalil and others, 2018). Further uphill to the north you will see the massive scale of the main Virginia City landslide headscarp, as well as the hummocky terrain of the landslide deposits.

Spring 1 has a flow of over 200 gpm, and is treated onsite and stored in a water tank for municipal use. An overflow channel runs when the tank is full. This spring has been in use since 1864. It provides sufficient water supply for the current population, but the secondary spring just uphill from here does not provide sufficient supply to act as a backup in case Spring 1 is compromised. Other springs have been identified in the landslide deposit, but none are as strong or consistent as Spring 1.

Continue west on Douglas Street for 0.15 mi.

Note: The road to the next stop requires a high-clearance vehicle. We will be coming back to the Spring 1 area, if a vehicle needs to be left behind.

36.8 mi Turn right onto a dirt road (Bozeman Trail) and head uphill to the northeast 0.6 mi.



Figure 9. Water infrastructure at STOP 5. (A) The main water tank for the town. Water is treated before entering this tank. When it is full, water is diverted into an overflow channel. (B) The pump house, located at the contact between lava flows (uphill) and tuff (downhill).



STOP 7—Main Landslide Scarp 37.4 mi

45.30290, -111.92240

This stop is essentially at the base of the main Virginia City landslide headscarp. Notice the scale of this feature and look down onto landslide deposits to the southwest. At the base of the headscarp, there are depressions and wetland areas that drain internally. These ponding features are common to landslide morphology. There are also two “dry lakes” on the plateau directly above the headscarp. These drain directly into the landslide and are likely landslide-controlled sag ponds related to tension above the scarp.

Turn around and go down the way you came up. Drive 0.6 mi.

38 mi Turn left on East Douglas Street and drive 0.2 mi. Turn right on Edgerton Street and drive 0.2 mi. Turn right onto East Wallace Street/MT-287 and drive for 1 mi. Turn left into the large public parking lot across from Virginia City Creamery, or park anywhere in the “downtown” area if this lot is full.

39.4 mi Lunch break to explore the main street area of Virginia City, visit the creamery, and/or eat lunch and socialize in the parking lot.

Once done in town, turn right out of the parking lot back onto East Wallace Street/MT-287, heading east. Drive for 3.4 mi and turn right onto a dirt road called Devils Lane. Continue driving south on Devils Lane for 2.9 mi.

STOP 8—Axolotl Lakes Overlook 45.7 mi

45.24882, -111.88108

This stop overlooks the Axolotl Lakes Wilderness Study Area. Visible to the southwest of here are many landslide features, including scarps, likely landslide-controlled sag ponds, and hummocky topography. A mapped thrust fault projects up the axis of the main landslide body (Kellogg and Williams, 2006).

The small lakes or ponds are formed as blocks slide and rotate, creating basins and damming pre-existing runoff channels. Why are there so many more lakes in this area? Why might the landslides look different here than nearer to Virginia City?

Figure 10 shows extensive landslide features in the Axolotl Lakes area as seen in lidar that is enhanced to show relative local relief.

Continue down some switchbacks for 0.2 mi, then turn left on Axolotl Lakes Road. Drive 1.1 mi, then turn right into the Twin Lakes pullout.

STOP 9—Twin Lakes Pullout 47 mi

45.24123, -111.86290

This area is home to the axolotl, a unique non-metamorphosing form of the barred tiger salamander that retains juvenile characteristics such as external gills and a caudal (tail) fin. The critically endangered true axolotl is native to central Mexico and is a different species. “Axolotl” roughly translates to “water dog” or “water monster” in the Aztec language. Montana’s axolotls typically live in deeper waters and average 6–10 in long. Some axolotls continue to develop and eventually become terrestrial, but some will stay in their “teenage” form for their whole lives. They have only been found in 2 of the 10 lakes in the Axolotl Basin. Trout transplanted in the lakes more than a century ago appear to have wiped them out of the rest (Driscoll, 2013).

This stop provides a different angle of the spectacular landslide topography of the area. The field trip will end here. Many trails lead off from this point to traverse the basin, so feel free to explore as long as you like! As you wander, notice the blocks of volcanics scattered throughout the deposit. Do they look the same or different from those observed near Virginia City?

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Figure 10. An overview of the Axolotl Lakes area, as seen in lidar enhanced with a simple local relief model and presented in a red image relief map. Note the extensive landslide features, landslide-controlled sag ponds, and lakes throughout this area.

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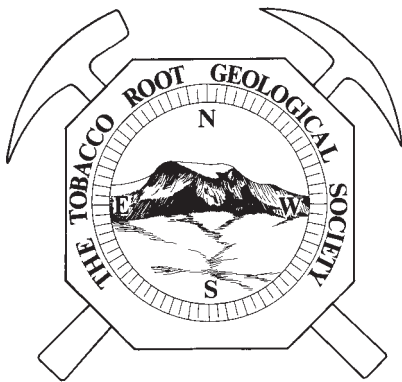
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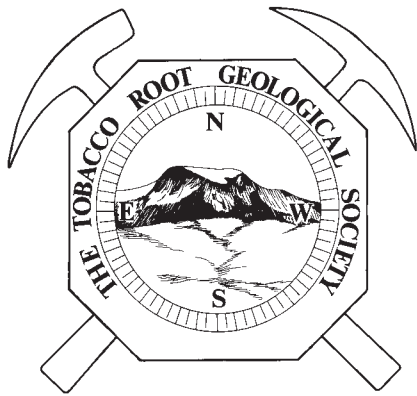
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