Guidebook of the 10th Annual Field Conference

Tobacco Root Geological Society

August 7-10, 1985 Bozeman, Montana

The Tobacco Root Geological Society, Inc. (TRGS) was organized in 1974 as a non-profit, scientific corporation to promote the study of the geology of the Northern Rocky Mountain Province, The organization sponsors annual conferences in the Northern Rocky Mountain Province and adjacent areas.

Cover Illustration: Upper Branham Lake in the Tobacco Root Mountains. Cover preparation courtesy of Wilkerson Photography, Dillon, Montana.

TOBACCO ROOT GEOLOGICAL SOCIETY

10th Annual Field Conference, 1985

GEOLOGY AND MINERAL RESOURCES
OF THE TOBACCO ROOT MOUNTAINS
AND ADJACENT REGION

"A Homecoming to the Tobacco Roots"

Bozeman, Montana
August 7-10, 1985

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SCHEDULE OF EVENTS

Wednesday, August 7, 1985

4:00-9:00 PM	Registration, poolside area, Gran Tree Inn (Inn of Bozeman). Preregistrants may pick up registration materials.
8:00 PM	Meeting of the Board of Directors and Executive Committee.

Thursday, August 8

8:00-12:00 AM	Registration continued.
8:50 AM	Technical Session begins. Welcome by Richard Berg, President TRGS.
9:00	**Christopher J. Schmidt** and **Beth Geiger**: Interaction of the Rocky Mountain foreland and Cordilleran Thrust Belt
9:30	**David R. Lageson**: Tectonic evolution of the Bridger Range and adjacent areas, southwest Montana
9:50	**Stephen S. Harlan** and **John W. Geissman**: Timing of deformation along the eastern margin of the Disturbed Belt, northern Crazy Mountains Basin, Montana
10:10	Coffee break
10:40	**Alan Stine** and **James G. Schmitt**: Sedimentology of the lower sandstone member, Lower Cretaceous Thermopolis Formation, southwestern Montana
11:00	**Donald L. Rasmussen** and **Robert W. Fields**: Cenozoic structural and depositional history, Jefferson and Madison intermontane basins, southwestern Montana
11:20	**Christopher W. Shaw**: Mass movements of the Gravelly Range, southwestern Montana
12:00	Luncheon (included in registration fee), Gran Tree Inn
1:30 PM	Technical Session resumes
1:30	**David W. Mogk**: Review of Archean basement geology of southwestern Montana

2:00	**M. Dean Kleinkopf** and **Viki L. Bankey**: Gravity and magnetic studies of the Stillwater Complex area, Montana
2:20	**Robert E. Derkey** and **M.J. Bartholomew**: Geologic history of the Deer Lodge-Elliston area with emphasis on the Elkhorn Mountains Volcanics and regional structures
2:40	Refreshment break
3:10	**Michael J. Trombetta** and **Robert A. Chadwick**: Geology of the Avon Rhyolite, Montana
3:30	**Lynne Dickman**: Assessment of mineral resource potential of the Helena National Forest
6:00	Social hour (no-host), Gran Tree Inn
7:00	Banquet, Gran Tree Inn (included in registration fee). Speaker: **Tom Straw**, Western Michigan U., "History and Activities of the Indiana University Field Station"

Friday, August 9

8:00 AM	Field Trips Nos. 1,2, and 3 depart from Gran Tree Inn parking lot
8:00 PM	General Business Meeting of Tobacco Root Geological Society - all members are urged to attend

Saturday, August 10

8:00 AM	Field Trips Nos. 4 and 5 depart from Gran Tree Inn parking lot

GEOLOGIC HISTORY OF THE DEER LODGE-ELLISTON AREA WITH EMPHASIS ON THE ELKHORN MOUNTAINS VOLCANICS AND REGIONAL STRUCTURES

by

DERKEY, Robert E., and BARTHOLOMEW, M.J., Montana Bureau of Mines and Geology, Butte, MT 59701

Volcanic rocks of the Deer Lodge-Elliston region include the 75-80 m.y. old Elkhorn Mountains Volcanics (EMV), the 48-50 m.y. old Lowland Creek Volcanics, and the 36-40 m.y. old Avon rhyolites. The EMV unconformably overlie sedimentary rocks of the Cretaceous Colorado Group and Kootenai Formation. Units of the Colorado Group and Kootenai Formation are also thrusted onto the EMV, northeast of Deer Lodge. Basalts and andesites involved in thrusting are considered part of the lower member of EMV, which has been dated at 79 m.y. on Cliff Mountain, east of Deer Lodge.

Two separate, but related groups of thrust faults, are recognized in the area. One is an imbricate, closely-spaced series of thrusts in EMV and the other is a major decollment involving EMV and pre-EMV sedimentary rocks. The closely spaced thrusts in volcanic rocks are a footwall imbricate fan, produced structurally below a major ramping thrust, which rode over the imbricate fan. Palinspastic reconstruction of this imbricate fan, utilizing an ash-flow tuff marker unit suggests a minimum eastward displacement of 6 kilometers (3.75 mi) for all faults of the fan. Based on this reconstruction, the major thrust ramp for the structurally higher thrust was in the vicinity of the present Flint Creek Range. The easternmost margin and the present position of this major decollement block extendssouthward from Elliston along the Little Blackfoot River and passes within 3 kilometers (1.9 mi) of the Boulder Batholith.

Northeast of Deer Lodge, younger N30W-trending normal faults form the northeast margin of Deer Lodge Valley. Similar normal faults form the margins of the Avon Valley Rhyolites, and minorbasaltic rocks occur around the southern margin of the Avon Valley and also occur along the southeast projected trend of the valley.

The main valley-filling sequence of the Deer Lodge Valley is a series of late Eocene through earliest Pliocene clastic sediments. The valley-filling sequence is unconformably overlain by coarse clastic detritus on late Pliocene through present, erosional surfaces. The oldest of the Pliocene through present erosional deposits caps the high pediment surfaces of the valley. At the north end of the valley, a series of three extensive Pleistocene deposits of coarse clastic detritus represent progressively lower drainages of the Baggs Creek-Cottonwood Creek area. Holocene deposits include terraces ranging from 1 to 2 meters (3.3 to 6.6 ft) to 8 meters (26.2 ft) above the present floodplains, as well as landslides, colluvium, and alluvial fans.

ASSESSMENT OF MINERAL RESOURCE POTENTIAL OF THE HELENA NATIONAL FOREST

by

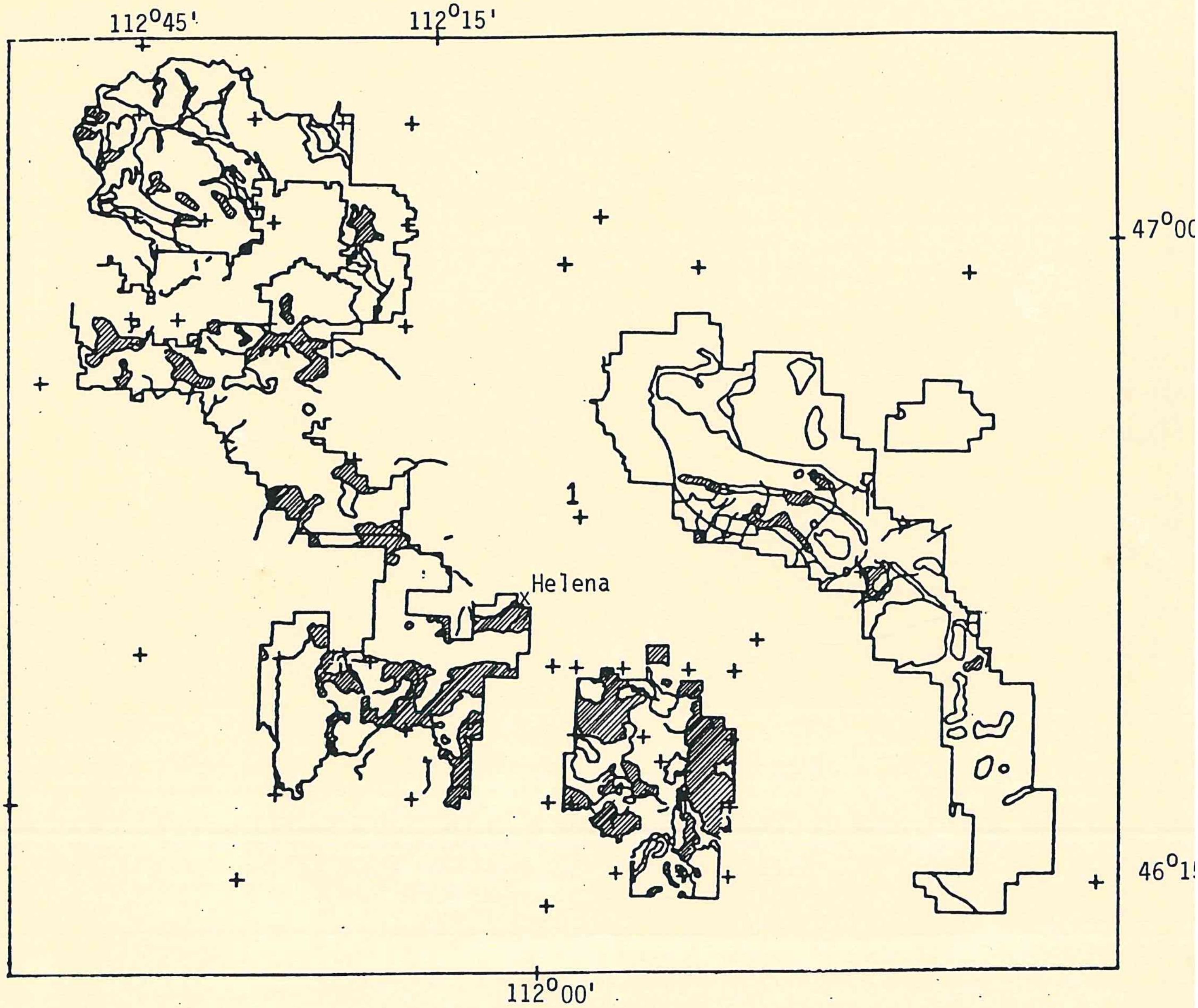
DICKMAN, Lynne, USFS Minerals and Geology, Missoula, MT 59802

The mineral resource potential of the Helena National Forest in West Central Montana was recently assessed for land management purposes. This potential is defined as "a measure of the likelihood of occurrence of valuable minerals or minerals that may become valuable within the foreseeable future" (Taylor and Steven, 1983).

Favorable geologic environments for the likelihood of occurrence of valuable minerals were identified by comparing available geologic, geochemical and geophysical data with that from known and developed deposits. U.S. Geological Survey compilations of ore deposit characteristics (Erickson, 1982 and Cox, 1983) provided the basis for a checklist approach in identifying these characteristics. Using this information, the land area was rated as having high, moderate, low or no mineral resource potential according to the "degree of fit" with occurrence model characteristics. This mineral potential rating was then qualified by a certainty rating which indicated the degree of reliability, i.e. sources, map scale, etc., used in the evaluation. The method essentially utilizes a pragmatic and subjective approach in which many diverse geologic factors have been integrated. It is similar to the methodology used by the U.S. Geological Survey for Wilderness and Conterminous U.S. Mineral Resource Assessment areas (Shawe, 1981). However, in this case, only secondary sources were used.

As documentation to accompany the principal mineral potential map, four technical reports were prepared that dealt with the forest in sections. The mineral commodities discussed in these reports are grouped as Locatables (lode and placer), Leasable-Nonenergy, Salable (common variety), and Leasable-Nonenergy. Each commodity was discussed with regard to the geologic environment in which it occurs, its distribution within the section, general characteristics, and the economic implications of its occurrence.

Favorable geologic environments suggesting significant mineral potential for precious and base metals occur throughout much of the forest. Most mineralization is associated with Cretaceous and Tertiary plutonic intrusion and volcanism, with the exception of copper, silver and lead found within stratabound deposits and cupriferous carbonate veins in faulted areas of copper- and silver-rich Precambrian formations. Recognizable deposit types include: epithermal gold and silver; silver, lead, zinc replacement in limestone; porphyry copper/molybdenum, high temperature silver, lead, and zinc replacements; contact metasomatic copper/gold/tungsten; breccia pipes; placer gold; and to a minor extent, bog manganese, titaniferous magnetite placers and uranium vein deposits.



Map depicting areas of high potential for precious and base metals in the Helena National Forest.

Areas of high mineral potential.

scale: 1" = approx. 14 mi.

The potential for oil and gas resources, while admittedly speculative, is assessed to be high for approximately half of the forest. This potential is defined as broad zones that are essentially the divisions of the Montana Disturbed Belt, as portrayed by Mudge (1982).

Numerous other mineral commodities exist within the Forest, which are present in minor amounts or are of little economic demand at this time. These include: phosphate, barite, limestone, talc, clay, pumice, building stone (including marble, granite, slate, and rhyolite), sand and gravel, iron, tin, manganese, fluorite, and sapphires.

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TIMING OF DEFORMATION ALONG THE EASTERN MARGIN OF THE DISTURBED BELT, NORTHERN CRAZY MOUNTAINS BASIN, MONTANA

by

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GEISSMAN, John W., Department of Geology, University of New Mexico, Albuquerque, NM 87131

The timing of thin-skinned fold and thrust belt deformation along the leading edge of the Disturbed Belt, in the Central Montana Salient, has not been well-defined. Although the easternmost folds of the Disturbed Belt in the northern Crazy Mountains Basin and Castle Mountains Uplift deform middle Paleocene and older strata, no upper limit for this deformation has been determined due to a lack of overlying, post-orogenic strata.

Alkaline igneous intrusives in the Robinson Anticline area, along the easternmost edge of the salient, have been variously described as being pre-, syn-, and post-tectonic with respect to fold and thrust belt deformation. Previous attempts to determine their relationship to folding through paleomagnetic and structural methods has been inconclusive. Our paleomagnetic analyses of sills of the Robinson Anticline Intrusive Complex indicate that in situ mean directions are well-grouped, and that a fold test is negative and significant at the 95% confidence level. This suggests that the Robinson Anticline intrusives were intruded subsequent to fold and thrust belt deformation. Radiometric (K/Ar) age dates of hornblende and biotite separates from these intrusives range from 52-48 mybp.

Paleomagnetic and radiometric age data, combined with pre-existing stratigraphic studies, thus indicate that Sevier-style fold and thrust belt deformation along the eastern margin of the Disturbed Belt in the Central Montana Salient is probably latest Paleocene to earliest Eocene in age. This age is in reasonably good agreement with that of similar structures along the eastern edge of the Northern Disturbed Belt.

GRAVITY AND MAGNETIC STUDIES OF THE STILLWATER COMPLEX AREA, MONTANA

by

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Positive gravity and magnetic anomalies indicate that the Stillwater Complex, mafic-layered intrusion of Archean age, may extend 20 km (12.5 mi) northeast of its outcrop at the Beartooth Mountains front. The regional geophysical patterns suggest that a sill-like northeastern extension of the complex underlies an area of about 4,400 km^2 beneath layered Phanerozoic sedimentary and scattered volcanic rocks. Additional evidence for an extension of the complex is xenoliths in the Tertiary Lodgepole intrusion about 8 km (5 mi) north of the surface exposures of the complex (Brozdowski and others, 1982); the xenoliths include foliated mafic amphibolite, Paleozoic sedimentary rocks, gneiss, and mafic cumulates.

A prominent high-gradient gravity zone of 20-mGal amplitude exists along outcrops of the Stillwater Complex and is related spatially to the faulted front of the Beartooth Mountains and to the Nye-Bowler structural zone, which continues east-southeast of the Beartooth Mountains (Foose and others, 1961). Gravity highs over the Stillwater Complex suggest the existence of thick pods of high-density rocks of the Basal and Ultramafic series of the Stillwater Complex. Positive magnetic anomalies and a steep magnetic-gradient zone to the northeast parallel Stillwater Complex outcrops and the more extensive, high-gravity gradient zone along the faulted front of the Beartooth Mountains. The steep magnetic-gradient zone forms the southwest flank of a conspicuous negative magnetic trough; the trough is inferred to reflect a combination of dipole effects of the Stillwater Complex magnetic mass and probably a deep structural zone along the northwest-trending Beartooth Mountains front, in which highly deformed and shattered rocks of the Stillwater Complex were depleted in magnetite during tectonic uplift of the Beartooth Mountains.

A broad gravity high parallels the Beartooth Mountains fault front and trends west-northwest for some 80 km (50 mi) about 10 km (6.25 mi) northeast of Stillwater Complex outcrops. This area of high-intensity gravity was interpreted from reconnaissance data by Bonini (1982) as a sheetlike eastward extension of the Stillwater Complex in crystalline basement beneath sedimentary rocks. The findings, based on interpretations of added gravity control, gravity modeling, and reconnaissance aeromagnetic data, tend to support Bonini's hypothesis but show the Stillwater Complex to be more limited in extent and to have a synformal configuration.

The magnetic patterns north of the Stillwater Complex are predominantly northeast-trending and are consistent with the magnetic grain and pronounced lineations in much of east-central Montana. Two prominent, northeast-trending, linear, high-intensity magnetic anomalies extend some 20 km (12.5 mi) beyond the Beartooth Mountains fault front. The relationship of the sources of these magnetic anomalies to the Stillwater Complex is uncertain, but they may be bodies of highly magnetic iron-formation similar to those of pre-Stillwater age near the base of the complex (Page and Nokleberg, 1974). The magnetic gradients suggest that the sources of the two high-intensity magnetic anomalies northeast of the Stillwater outcrop are bodies of iron-formation, some 3 km (1.9 mi) below the ground surface, with symmetrical shapes that can be inferred to indicate that the bodies are near-vertical units; they may be steeply dipping units of iron-formation in the basement complex. This conclusion is supported by results of an analysis of a detailed magnetic survey flown across the complex by Anaconda Minerals Corp., which show that the greatest intensity anomalies, exceeding 6,000 gammas, occur over the iron-formation within the metamorphic aureole of the complex (R.J. Blakely, written commun., 1984).

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TECTONIC EVOLUTION OF THE BRIDGER RANGE AND ADJACENT REGIONS, SOUTHWEST MONTANA

by

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The Bridger Range records over 1.5 billion years of superimposed tectonism in southwestern Montana and includes elements of all major styles of deformation that have shaped western Montana. The major tectonic events recorded in this range include: 1) Proterozoic normal faulting associated with development of the Belt Basin, 2) "thin-skinned" fold and thrust belt deformation, 3) Laramide crustal shortening, and 4) Neogene normal faulting associated with crustal extension north of the Snake River Plain. Superposition of these different tectonic events has resulted in overlapping, reactivated structural elements.

Proterozoic Belt strata of the northern Bridger Range were thrusted over Archean crystalline rocks of the southern Bridger Range along an array of northwest-trending, oblique-slip lateral ramps, known as the Ross Pass and Cross Range Fault Zones. Structural analysis and the relationship of younger-over-older suggests that these Sevier-style faults are reactivated Proterozoic normal faults, subsequently folded in the early Eocene by uplift of the Archean core of the Bridger Range. Gravity and reflection seismic show that the crystalline basement of the southern Bridger Range has been thrusted eastward as much as 5 km (3.1 miles) over Phanerozoic strata of the Crazy Mountains Basin. Neogene extension of the greater Three Forks Basin has exploited NW-trending segments of the older Sevier-style faults in the development of the listric normal fault along the west flank of the Bridger Range.

A REVIEW OF THE ARCHEAN GEOLOGY OF SOUTHWESTERN MONTANA

by

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The Archean basement of southwest Montana records at least one billion years of evolution of continental crust. Within this terrane there are fundamental differences in the nature of the crust with dominantly igneous rocks in the Beartooth Mountains and dominantly meta-sedimentary rocks in the ranges to the west. A working model for the evolution of this terrane calls for: 1) ancient, thick, continental crust in the Beartooth Mountains (ca. 3.6-3.0 Ga); 2) shedding of sediments into an adjacent basin(s); and 3) late-Archean orogeny that resulted in generation of andesites and voluminous calc-alkaline granitic rocks in the Beartooth Mountains, nappe formation, and upper-amphibolite to granulite grade metamorphism of the sedimentary suites.

The Beartooth Mountains are comprised of voluminous 2.75 Ga old granitic rocks with inclusions of 3.4 Ga old granulite-grade meta-supracrustal rocks, and 2.8 Ga old amphibolites of andesitic composition (composite work of Mueller, Wooden, Henry, Mogk, Richmond). The supracrustal assemblage was deposited in a stable platform environment and metamorphism occurred at T=700-800^{o}C and P=6-7 Kbar. The generation of large volumes of andesite at 2.8 Ga marks the beginning of a second orogenic cycle. Emplacement of large volumes of late to post-kinematic granites occurred throughout the Beartooth Mountains at about 2.75 Ga.

The North Snowy Block, northwestern Beartooth Mountains, is a late Archean mobile belt (Mogk). The NSB is comprised of 4 linear belts: 1) a 3.6 Ga old ductiley-sheared trondhjemitic gneiss-amphibolite complex; 2) phyllitic Davis Creek Schist; 3) an upper amphibolite (700^{o}C) paragneiss; and 4) the 2.75 Ga granitic Mount Cowen Augen Gneiss. These units have been tectonically emplaced (except for the Mount Cowen unit); sub-horizontal lineations suggest emplacement along transcurrent faults. Overlying the linear belts are 2 thrust sheets, the Pine Creek Nappe Complex (quarzite-marble amphibolite), and the migmatitic Heterogeneous Gneiss.

To the west of the NSB, the Archean basement is predominantly of supracrustal origin, with coarse clastics in the northern Gallatin and Madison Ranges (Spencer and Kozak; Salt) and passive margin-type sediments in the Tobacco Root, Gravelly and Ruby Ranges (Vitaliano et al., Garihan). These sequences are characteristically isoclinally folded with associated nappes, and metamorphism is in the upper amphibolite to granulite grade (Cordua, Dahl, Erslev). A recently discovered batholithic complex in the Spanish Peaks (Salt) represents the only significant late-Archean plutonism in this area. The Highland and Blacktail Ranges are comprised dominantly of quartzofeldspathic gneisses which have recently been interpreted as supracrustal in origin (Clark). The structural and metamorphic style of the western Archean basement requires large-scale compression and deep burial, presumably in response to tectonic thickening.

Late Archean generation of andesites and granitic rocks, and associated transcurrent and thrust faulting in the Beartooth Mountains is probably the result of oblique subduction. A somewhat later collisional event may be responsible for the collapse of the stable platform sediments and subsequent deformation and metamorphism. Possible mechanistic models include accretion of allochthonous terranes, or collision of another continent or island arc. Current geochronological, geochemical, petrological and structural studies are addressing these questions.

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CENOZOIC STRUCTURE AND DEPOSITIONAL HISTORY, JEFFERSON AND MADISON INTERMONTANE BASINS, SOUTHWESTERN MONTANA

by

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Recent seismic and gravity data from the Cenozoic Jefferson and Madison Basins provide new information concerning their structural and depositional histories. Both are elongated structural basins, trending north-south, formed as a result of horizontal extension, after Laramide horizontal thrusting. Each basin is bounded on the east side by a sinuous, faulted, steep mountain front, and large, west-sloping, alluvial fans extend almost completely across both basins.

Gravity data show that each basin in the subsurface is asymmetric with a large, steep, west-dipping fault on the east flank and one or more east-dipping faults of smaller magnitude, on the west flank. The deep axis of each basin runs parallel to the east mountain front, and lies east of the surface geographic axis. The Jefferson Basin has two deep closed structural lows, one east of Silver Star and one east of Twin Bridges, which are separated by a structural arch. Sediment depth on the arch exceeds 3000 meters (9850 ft.). The Madison Basin is shallow on its north end (approximately 2100 meters or 6900 ft.) where it is terminated by the prominent northwest-southeast Spanish Peaks structural trend, and progressively becomes deeper south of Ennis: 4500 meters (14750 ft.) or more.

Seismic data confirm or support the gravity interpretation. Seismic data also show the folded and thrusted rocks of the east mountain foot-wall block dipping steeply westward to where they gradually disappear beneath the thick Tertiary sediments. Tertiary strata lying directly against the large west-dipping basin fault show dip reversal caused by drag-folding during basin subsidence. Downthrown "roll-over" type anticlines are thus present on the east side of the basins. Numerous small faults, many antithetical, cut the deeper strata and diminish in throw upward.

Strata seen on the seismic sections can be subdivided into a lower set which forms the bulk of the basin fill (possibly equivalent to the Renova Formation, Late Eocene to Early Miocene); a thinner middle set unconformably overlying the lower set (equivalent to the Sixmile Creek Formation, Miocene and Pliocene); and an upper set composed of west-dipping Quaternary alluvial fan deposits. Each set thickens toward the east basin-bounding fault. In the lower "Renova" set, lacustrine intervals are indicated by their consistent lateral seismic character, whereas fluvial intervals appear to terminate abruptly.

INTERACTION OF THE ROCKY MOUNTAIN FORELAND AND CORDILLERAN THRUST BELT: SOUTHWESTERN MONTANA

by

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GEIGER, Beth, University of Montana

The geometry and movement patterns in the frontal fold and thrust belt in southwestern Montana were strongly influenced by the geometry of the basement surface within the adjacent and overlapping Rocky Mountain foreland. The pre-thrusting geometry of the basement surface was controlled principally by three sets of structures: an east-trending normal fault (Willow Creek Fault) of Middle Proterozoic age which marked the southern boundary of the Helena Embayment of the Belt Basin; northwest-trending left-reverse slip faults of late Cretaceous age, inherited from Middle Proterzoic trends; and north-to-northeast-trending thrusts associated with broad (50-80 km or 31-50 mi) arches in the basement surface.

The basement surface in the Willow Creek Normal Fault was down on the north in Middle Proterozoic time and was oriented roughly normal to the west-east shortening in the thrust belt. The resulting thrust belt structural trend (southwest Montana Transverse Zone) is chiefly a major lateral ramp in the thrust belt with significant right-lateral strike slip movement.

The northwest-trending faults were associated with basement-cored anticlines on their hanging walls. These structures controlled the geometry and kinematics of thrust belt structures in the northern Blacktail Salient, northern McCartney Salient, and the anastomosing pattern within the southwestern Montana Transverse Zone. The principal effect of these foreland anticlines was to deflect thrust sheets into oblique ramps. The ramps are characterized by local younger-over-older thrust relationships, refraction of shortening directions, and minor folds and solution cleavage which formed normal to shortening directions.

The two westernmost basement arches in the foreland (Tobacco Root-Ruby Arch and Blacktail-Snowcrest Arch) are interpreted as ramp anticlines above major northwest-dipping basement thrusts. The northwestward dip of the basement surface in these anticlines was partly responsible for the termination of thrusting in the frontal thrust belt in latest Cretaceous and Paleocene time. Major lateral or oblique ramps have been documented within thrust sheets in far southwestern Montana, where the foreland basement thrusts strike beneath the thrust belt.

MASS MOVEMENTS OF THE GRAVELLY RANGE, SOUTHWESTERN MONTANA

by

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Mass movements have significantly influenced the geomorphology of the Gravelly Range of southwestern Montana. Sixty km (37.5 mi.) south of the Tobacco Root Mountains, the Gravelly Range is the eastern limb of the north-south trending Ruby Syncline. Mississippian and Permian rocks make up the higher mountains (at el. above 3000 m. or 9843 ft.), with Archean gneiss and schist to the east. The Gravelly Range Road follows Triassic and Jurassic rocks of the range crest, and dipping Cretaceous rocks make up the western portion. Capped in places by Tertiary volcanic rocks, this area is quite susceptible to mass wasting.

Using aerial photography in a reconnaissance study, spectacular examples of mass movements of the slump and earth flow types were found. These were delineated as part of a preliminary surficial geology and geologic map of the Cliff Lake and Monument Ridge Quadrangles. Volcanic cap-rocks and other resistant rocks often exhibited near-equidimensional slumps, ranging from hectare to several square kilometer size, and failed along arcuate shear zones when under-cut. Elongate flows of shale and sandstone range from one-half to seven km from head to toe. Slumps were found to occupy 5% of the area on the Cliff Lake and Monument Ridge Quadrangles, and flows occupy 8% of the same area. Twelve percent of Cretaceous rocks in this area were found to have flowed. If the source of these flows is taken into account, up to one quarter of the area of Cretaceous rocks were affected by mass movement of possible Pleistocene or younger age. With the presence of resistant Tertiary volcanic rocks prone to slumping, along with the dip slope of Cretaceous rocks likely to flow, the Gravelly Range has spectacular and well-exposed mass movements.

SEDIMENTOLOGY OF THE LOWER SANDSTONE MEMBER, LOWER CRETACEOUS THERMOPOLIS FORMATION, SOUTHWESTERN MONTANA

by

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Detailed lithologic observation and sedimentary structure analysis of the lower sandstone member of the Lower Cretaceous (Albian) Thermopolis Shale in the Bozeman, Montana area document deposition in a storm-dominated, shallow marine, clastic, shelf setting. Four facies and two sub-facies are present in the sandstone member. These facies include: 1) crossbedded sandstone, 2) mixed sandstone/shale, which is subdivided into flaser bedded and rippled sheet sandstone sub-facies, 3) bioturbated sandstone/siltstone, and 4) shale.

The crossbedded sandstone facies is comprised of medium to very fine-grained, chert-bearing quartzarenite which contains abundant pebble-sized mudstone clasts. Sedimentary structures include, in decreasing order of abundance, large-scale tabular, wedge, and trough crossbedded sets, horizontal to subhorizontal, planar to wavy stratification, current and wave ripple cross-lamination, and medium-scale, trough crossbedded sets. Ophiomorpha and Thalassanoides are common.

The mixed sandstone/shale facies contains 10-15% light to medium gray shale and more than 50% very fine to fine-grained mudstone-clast and chert-bearing quartzarenite and siltstone. Small horizontal trails are common at sandstone-shale contacts. Ripple cross-laminated sandstone, with associated symmetrical and asymmetrical ripple form sets, as well as horizontally to wavy laminated, and structureless, thinly bedded sandstone characterize the flaser bedded sub-facies. Shale is present as clay drapes, parting, or in thin lenses. The rippled sheet sandstone sub-facies contains tabular to slightly lenticular sandstone beds 5-25 cm (2-10 in) thick separated by shale intervals up to 5 mm (0.2 in) thick. Sandstone beds contain horizontal to wavy stratification with symmetrical and asymmetrical ripple marks on upper bed surfaces.

The bioturbated sandstone/siltstone facies is comprised of very fine-grained quartzarenite and siltstone. Sedimentary structures are lacking due to extensive bioturbation.

Gray shale containing less than 25% interbedded very fine grained sandstone and siltstone characterize the shale facies. Wavy and subhorizontal stratification are common.

Cross-stratified sandstone portions of the above facies were deposited by a variety of bedforms including sandwaves, megaripples, and ripples. Depending upon hydraulic regime and sand availability, these bedforms existed both as discrete bedforms migrating upon a muddy substrate as well as complexes of superimposed bedforms within sand fields or ridges of lateral extent up to at least 5 km (3.1 mi). Erosion of sediment from these sand fields during storm events supplied sand for the mixed sandstone/shale facies. Mud was locally deposited in protected areas between sand ridges and fields. Where sand sedimentation rates were low, the deposits were reworked by deposit feeders resulting in bioturbated sandstone intervals.

GEOLOGY OF THE AVON RHYOLITE, MONTANA

by

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The Avon Rhyolite, exposed in the southern portion of the Avon Valley, Montana, is a remnant of a more extensive volcanic pile, perhaps 400 meters (1312 ft) or more thick. The volcanics were emplaced in Eocene time (40-36 Ma) as lava flows and ash flows. The rhyolite erupted onto Cretaceous sedimentary rocks, Elkhorn Mountains Volcanics, and Lowland Creek Volcanics.

The lava flow sequence is approximately 250 meters (820 ft) thick and is light tan, purple or dark red. Phenocrysts comprise an average of 10% of the rock and are composed of: sanidine, quartz, biotite, plagioclase, and Fe/Ti oxides. Sparse xenoliths (up to 3 cm or 1.2 in) of rhyolitic composition and lithophysae (up to 6 cm or 2.4 in) are present throughout the lava flow sequence. The middle and upper portions of individual flows have a dull, glassy matrix and may exhibit concentric flow folds. Dark crystal-rich perlitic zones and similar contorted folds may be found in the glassy matrix in the lower third of individual flows.

Flow banding under magnification appears as dark bands within a glassy matrix, and as aligned elongate phenocrysts. In outcrop, it is seen as cm-scale platy jointing, commonly folded. The steep dip and concentric folding of these bands is characteristic of a viscous lava flow. Paleo-flow directions have been estimated from examination of flow banding.

The ash flow unit, which lies above the lava flow sequence, is 150 meters (492 ft) thick, but was probably thicker prior to erosion. Its color ranges from white to orange. Clasts of rhyolite and pumice (up to 6 cm or 2.4 in) are common and phenocrysts similar to those in the lava flows are abundant in the matrix.

The Avon region is strongly influenced by extensional tectonics. The Avon Valley is a graben containing a thick accumulation of sediment. Faulting may predate or be coeval with the rhyolite emplacement, and has continued into Holocene time. Rhyolite has eroded from uplifted highland blocks and may be buried beneath the downdropped valley floor. The easily erodible ash flow unit was evidently once an extensive sheet, and now occurs only as isolated remnants.

ARCHEAN GEOLOGY OF THE SPANISH PEAKS AREA, SOUTHWESTERN MONTANA: FIELD TRIP

by

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Archean exposures of the Spanish Peaks area of the northern Madison Range, southwestern Montana, consist of at least four distinct terranes, including a batholithic complex unrecognized by previous study (Salt and Mogk, 1985). Differences in lithology, metamorphic grade, and structural style characterize these terranes. This area represents a structural and lithologic transition between the older (3.6 b.y., Mueller et al, 1982), predominantly granitic, Beartooth Block to the east, and the younger (2.7 b.y., James and Hedge, 1980) metasupracrustal sequences of the Dillon Block to the west (Fig. 1).

Spencer and Kozak (1975) describe the major structural trends of this area in the context of a single lithostratigraphic sequence. Recent detailed field and petrographic studies, however, show that the Spanish Peaks area can be divided into at east four distinct terranes, from the Beartrap Canyon of the Madison River to the Gallatin River Canyon (Fig. 2).

The Gallatin Peak terrane is characterized by a batholithic complex consisting of at least four intrusive phases, including: foliated quartz diorite, hornblende monzodiorite, granodiorite, and unfoliated granite. These phases are intrusive into a metasupracrustal sequence. Assemblages of this sequence are characterized by metapelites, in which kyanite is the stable aluminosilicate, and orthoamphibole-bearing gneisses and schists. Several stages of amphibolitized mafic dikes and sills occur throughout this terrane and are composed of hornblende and plagioclase, with lesser amounts of quartz and garnet.

In the Jerome Rock Lakes terrane, most of the batholithic phases are conspicuously absent. The lithologies are dominantly metasupracrustal, with abundant metapelites in which kyanite breaks down to form sillimanite. Mafic bodies exhibit transitional granulite facies assemblages consisting of hornblende, diopside, garnet, plagioclase and quartz. These assemblages exhibit partial re-equilibration to upper amphibolite facies.

The Beartrap Canyon and Gallatin River Canyon terranes are similar to one another and consist primarily of biotite quartzofeldspathic gneiss of arkosic composition, which is locally heavily migmatized, probably by anatectic melting. As in the Jerome Rock Lakes area, mafic assemblages consist of transitional granulite facies assemblages, partially re-equilibrated to upper amphibolite facies. Metapelitic assemblages, however, are rare in these two terranes.

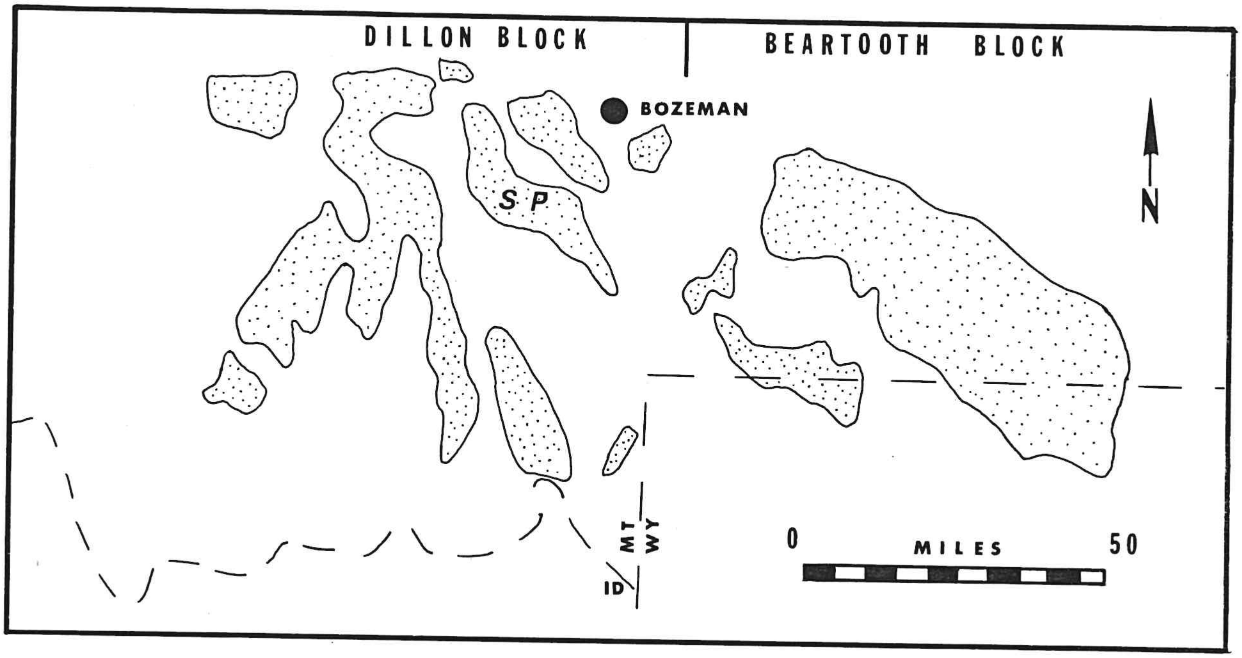

Figure 1. Archean exposures of southwestern Montana. ***SP***: Spanish Peaks.

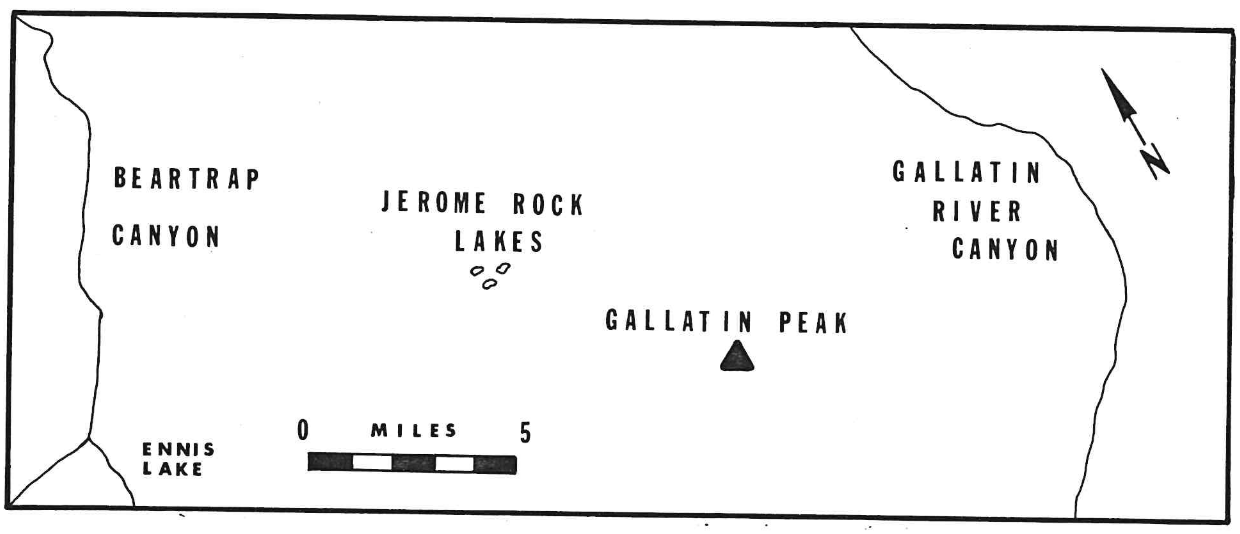

Figure 2. Location of Archean terranes in the Spanish Peaks area, as described in text.

A ductile shear zone separates the Gallatin Peak and Jerome Rock Lakes terranes. Transcurrent faulting is also strongly suspected between the Jerome Rock Lakes and Beartrap Canyon terranes and between the Gallatin Peak and Gallatin River Canyon terranes.

Similar relationships involving the juxtaposition of allochthonous units have been described in the southern Madison Range (Erslev, 1983). These relationships suggest that the Archean exposures of the Madison Range represent either the collapse and juxtaposition of a basin, marginal to the Beartooth Block or, alternatively, the Madison Range may be the zone of docking of an entirely allochthonous Dillon Block to the Beartooth Block.

FIELD TRIP

The highlight of this trip is a 2-day hike through the heart of the spectacular Spanish Peaks Wilderness Area. Many key relationships are well-exposed and provide an excellent insight into the Archean history of southwestern Montana. A schematic diagram of the scheduled route is shown in Figure 3. More detailed information concerning other trails in the area is found on the U.S. Forest Service topographic map of the wilderness area, which can be obtained upon request from the Gallatin National Forest.

Exposures will be examined from the previously described Gallatin River Canyon, Gallatin Peak, and Jerome Rock Lakes terranes. Several stops in the Gallatin River canyon will be made while driving to the trailhead. The first day's hike from Bear Basin to Thompson Lake will be in the batholithic complex. On the hike out to Spanish Creek, the shear zone will be crossed just north of Mirror Lake. The trail then proceeds through the metasupracrustal sequences of the Jerome Rock Lakes terrane.

Gallatin River Canyon, Hwy 191

Stop 1: Kitchen Rock, 28 miles (17.5 km) south of Bozeman. Characteristic exposure of migmatitic quartzofeldspathic gneiss of Gallatin River Canyon. Meter-scale nappe, kink folds in the gneiss.

Stop 2: Greek Creek Campground (USFS), 35 miles (21.9 km) south of Bozeman. Folded mafic boudins in gneiss. Transitional granulite facies assemblages in mafic bodies.

Stop 3: Big Sky, Montana, 46 miles (28.8 km) south of Bozeman. Spanish Peaks Fault, a Laramide structure, expressed by steeply dipping, overturned Lower Paleozoic strata. To the west off of Hwy 191, Bear Basin trailhead is about 6 miles (3.7 km) along the road to the ski resort.

Bear Basin to Thompson Lake Trail, 8.5 miles (5.3 km)

Mile 0(km 0): Trail begins in glacial deposits at the mouth of Bear Basin and continues through the heavily timbered creek bed. Moose and/or deer may be seen along the lower trail. Black bears

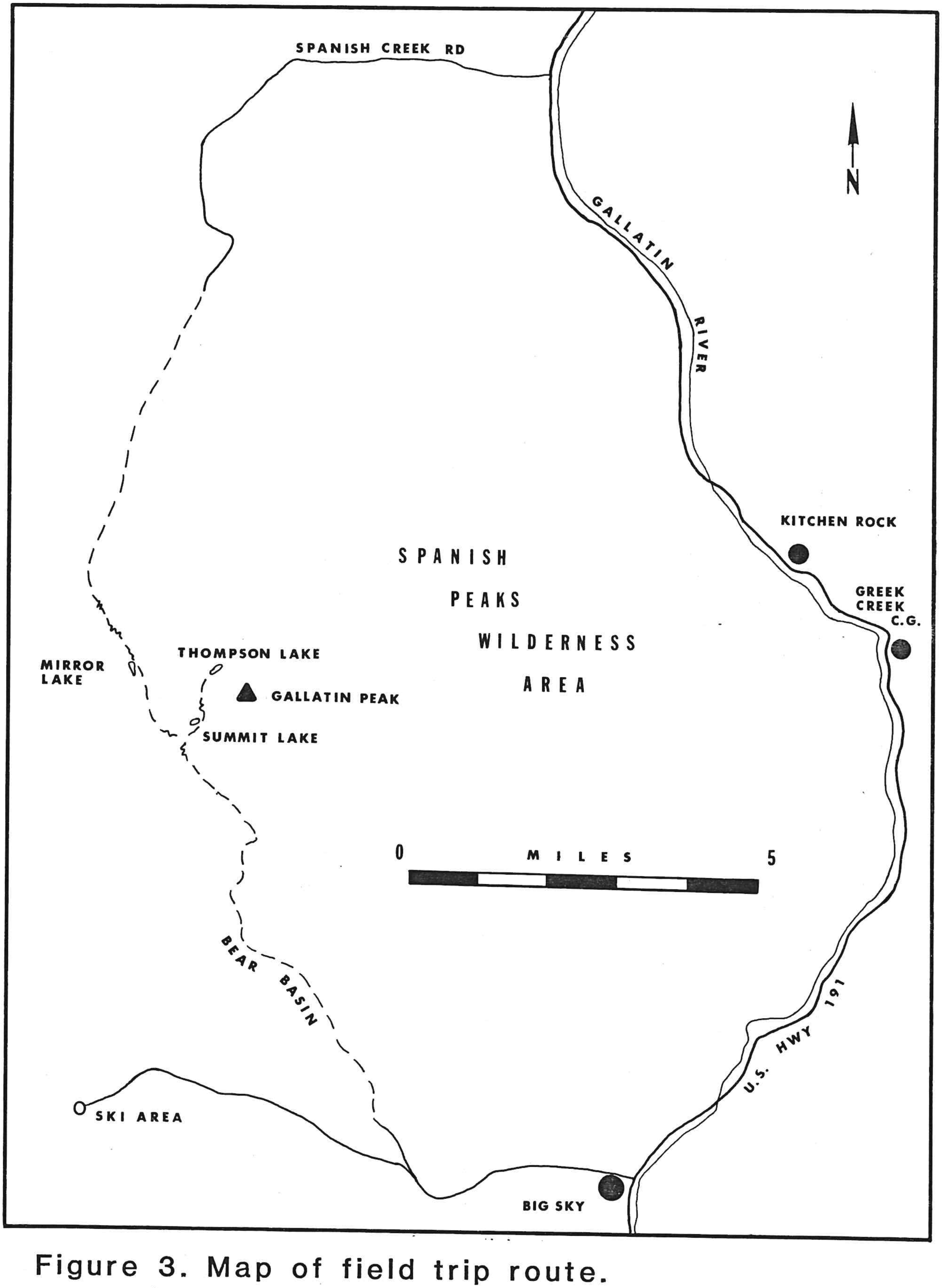

Figure 3. Map of field trip route.

are also local inhabitants; Old Ephraim (GRIZ) is known to migrate through this area.

Mile 2.5(km 1.6): Northwest trending Spanish Peaks Fault. Orthoamphibole-bearing gneiss is among the first Archean outcrops encountered on the trail north of the fault.

Mile 3.0(km 1.9): Foliated quartz diorite and granodiorite exposures. Cross-cutting relationships are well-exposed on ridgetop to east. North of these exposures is a sequence of metasupracrustal rocks. Some of the boulders along the trail consist of orthoamphibole, kyanite, plagioclase and quartz.

Mile 4.5(km 2.8): Cliff-scale agmatite exposure on the west wall. Contact between biotite schist and foliated hornblende monzodiorite has been forcibly intruded by unfoliated granite.

Mile 5.5(km 3.4): Summit Bear Basin Divide. The trail passes through a continuous, amphibolitized dike. Numerous similar dikes can be seen from the divide. The continuous nature of these dikes contrasts sharply with those found in the other terranes.

Mile 6.5(km 4.1): Summit Lake. A major body of granite intrudes foliated hornblende monzodiorite. Mafic bodies at outlet of the lake contain assemblages hornblende-cummingtonite-plagioclase and hornblende-garnet-plagioclase-quartz. Some garnets are as much as one inch in diameter. An unlineated amphibolitized dike cuts the unfoliated granite.

Mile 8.5(km 5.3): Thompson Lake, elevation 9200 ft.(2804 m); Gallatin Peak rises to just over 11,000 ft.(3353 m). Numerous amphibolitized dikes are exposed on the face of Gallatin Peak.

Thompson Lake to Spanish Creek Trailhead, 11.5 miles(7.2 km)

The trail returns to Summit Lake, then turns north to Mirror Lake. The lithologies along this portion of the trail are similar to the Thompson Lake area, dominated by hornblende monzodiorite injected by unfoliated granite, with scattered amphibolites. At the north end of Mirror Lake, three miles from Thompson Lake, a northeast-trending ductile shear zone is encountered. North of this zone, the batholithic phases are conspicuously absent. Sillimanite-bearing metapelites are encountered, as well as mafic bodies with transitional granulite facies assemblages. The area north of Mirror Lake is heavily timbered and it will be necessary to climb out of the creek bed to see these exposures. The trail ends at the main fork of Spanish Creek. From here, a light duty road leads back to Highway 191, about 18 miles (11.2 km) south of Bozeman.

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STRUCTURAL AND STRATIGRAPHIC GEOLOGY OF THE CENTRAL BRIDGER RANGE, MONTANA
Field Trip Road Log

by,

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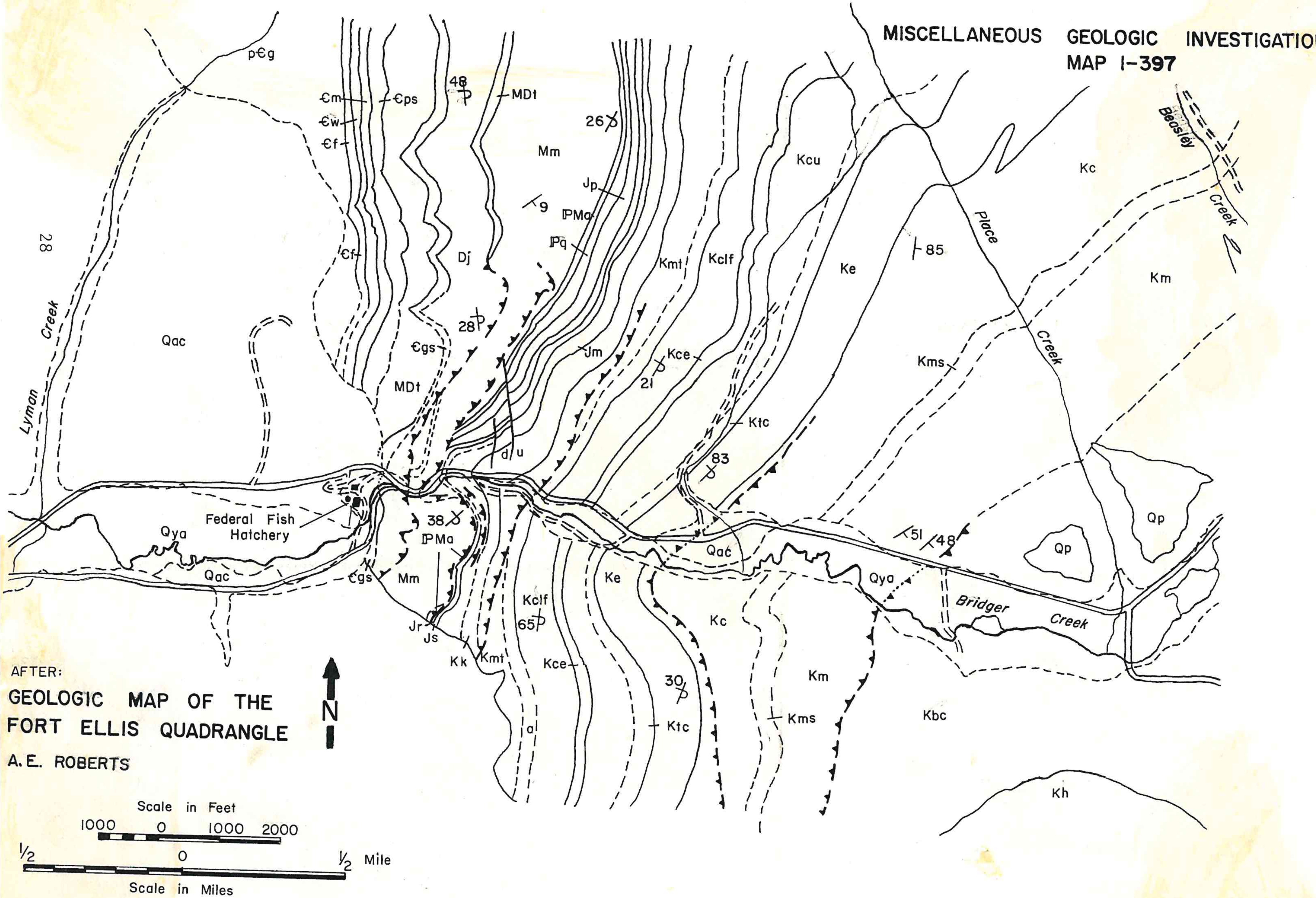
INTRODUCTION

The purpose of this field trip is to provide an overview of the geology and tectonic evolution of the Bridger Range and surrounding area. In addition, we will examine Mesozoic and upper Paleozoic rocks on the eastern flank and crest of the range. The field trip will be divided into two segments, the first from Bozeman to the Fairy Lake Campground by car, and the second involving a hike from Fairy Lake to the summit of Sacajawea Peak (elevation 9,665 feet; 2,946 meters). The road log is modified from Lageson and others (1983).

The Bridger Range trends roughly north-south for 25 miles (40 km) from Bridger Creek on the south to Blacktail Mountain on the north, and topographically separates the Gallatin Valley (eastern Three Forks Basin) to the west from the Crazy Mountains Basin to the east. The crest of the Bridger Range is composed of resistant, vertical to east-dipping limestones of the Mississippian Madison Group, which form spectacular cliffs in east-facing glacial cirque headwalls.

Structurally, the Bridger Range is an east-dipping homocline, with Precambrian rocks on the west side and Paleozoic and Meozoic rocks on the east side of the range. However, this apparent structural simplicity veils the complex tectonic history of the range. The Bridger Range records over 1.5 billion years of superimposed tectonism, and includes elements of all the major styles of deformation that have shaped southwestern Montana. The Bridger Range is located at the juncture of four major tectonic provinces (Woodward, 1982): 1) the south margin of the Proterozoic Belt Basin, 2) the south margin of the Helena salient of the Sevier fold and thrust belt, 3) the northwest margin of Laramide foreland deformation, and 4) the east margin of Neogene crustal extension north of the Snake River Plain. Superposition of these different tectonic regimes has resulted in overlapping and reactivated structural elements, as discussed below.

The first major deformational event recorded in the Bridger Range was Proterozoic rifting and normal faulting associated with subsidence of the Belt Basin (the Archean history will not be addressed by this field trip, as this will be the main topic of the Madison Range field trip). The Bridger Range overlaps the south margin of the Belt Basin, as evidenced by the Proterozoic Lahood Formation in the northern part of the range, while Archean gneiss (the source terrane of the Lahood) crops out in the southern part of the range and is nonconformably


MISCELLANEOUS GEOLOGIC INVESTIGATIO
MAP I-397
AFTER:
GEOLOGIC MAP OF THE
FORT ELLIS QUADRANGLE
A.E. ROBERTS
Scale in Feet
1000 0 1000 2000
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Scale in Miles
N
Federal Fish
Hatchery
Lyman Creek
Place Creek
Bridger Creek
Beasley Creek
pЄg
Qac
Qya
Qp
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MDt
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Jp
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Kms
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Km
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Kh
d u
a
48
26
9
28
21
83
85
51
48
38
65
30

FIGURE 1

overlain by the Cambrian Flathead Sandstone. The Pass fault, which trends northwest through Ross Pass, was one of several normal faults that formed the south margin of the Belt Basin (McMannis, 1963).

The second major deformational event was thrusting and folding during the late Mesozoic and early Cenozoic, largely correlative with the Sevier orogeny and the foreland fold-thrust belt of Wyoming, Utah and Idaho. In the Bridger Range, the Pass and Cross Range faults were reactivated as oblique-slip lateral ramps (right-lateral and north side relatively up) along the south margin of the Helena salient (Lageson and others, 1984; Schmidt and O'Neill, 1982).

The dominant east-dipping character of the Bridger Range was produced by the third deformational event. During the Laramide orogeny, Precambrian crystalline rocks were uplifted and thrusted up to 3 miles (5 Km) eastward over the west margin of the Crazy Mountains Basin. The present Bridger Range formed the east limb of this larger, Precambrian-cored anticlinorium.

The last major tectonic event is on-going today. Neogene crustal extension north of the Snake River Plain is expressed by north-trending valleys and ranges across southwestern Montana. The Gallatin Valley and greater Three Forks Basin have been extended and down-dropped to their present level since mid-Tertiary time (Robinson, 1963; Zim and Lageson, 1985). A large normal fault has formed along the western base of the Bridger Range, thus uplifting the present range as the footwall block. It is likely that segments of older faults have been reactivated as contemporary normal faults.

This field trip will provide an opportunity to observe the stratigraphy and various structural components of the Bridger Range, as discussed above. The emphasis will be on the tectonic history of the range, and how the Bridger Range fits into the broader tectonic evolution of western Montana

ROAD LOG

Mileage (in miles)

0.0 Holiday Inn, north 7th Avenue. Proceed north across I-90 interchange.

0.4 (incremental miles)

0.4 Turn right on Griffin Drive at corner of Conoco station and Bozeman Veterinary Hospital.

0.3

0.7 Bozeman Exxon terminal on right.

0.1

0.8 Cross railroad tracks.

1.2 Stop sign at intersection of Griffin Drive and Montana Highway 86; turn left and cross East Fork of Gallatin River.

1.3

2.5 Highway curves left. Tertiary Bozeman Group (upper part) on right, composed of poorly stratified, variously consolidated, tuffaceous siltstone, claystone, sandstone, and conglomerate. Hummocky landslide development in low hills to right. Bridger Range trending north-south from 10:00 to 12:00. High peak directly ahead is Baldy Mountain (elevation 8,767 ft.).

0.4

2.9 Cross Bridger Creek. Lyman Creek at 9:00, cutting down through alluvium and colluvium deposited on west flank of Bridger Range. Hills from 3:00 to 4:00 across Bridger Creek are composed of Tertiary Bozeman Group in a slumped topography.

0.9

3.8 Entrance on right to U.S. Fish and Wildlife Cultural Development Center.

0.2

4.0 Picnic area on left. Crossing approximate trace of Bridger range-front normal fault, representing the easternmost Basin and Range normal fault in this part of Montana. The normal fault is superimposed on an older Laramide structural uplift, with the hanging wall downthrown to the west forming the Gallatin Valley.

0.1

4.1 Narrows of Bridger Canyon. Overturned and thrusted Mississippian Lodgepole Limestone to left and right. The Lodgepole is well-bedded, cliff-forming, gray limestone. Landslide at 3:00 occurs in Big Snowy - Amsden strata; slide occurred in 1959 after construction widened the roadcut.

0.1

4.2 Imbricate thrusting at 9:00 with Mississippian strata in fault contact with the Jurassic Morrison Formation. The Laramide structure in this part of the Bridger Range may be interpreted as crowding and thrusting on the overturned, east-verging limb of the Bridger anticlinorium. Roberts (1964) mapped a total of six thrust faults from the narrows of the canyon to the strike valley of the

Upper Cretacous Billman Creek Formation (Livingston Group), a distance of approximately 1.6 miles (Figure 1).

0.3

4.5 Thrust fault to left has placed Lower Cretaceous Mowry and Thermopolis Shales (dark gray shales with basal sandstone) over the lower shale member of the Cody Formation and the Frontier Formation.

0.4

4.9 Upper Cretacous Eagle Sandstone in roadcut on left.

0.2

5.1 Crossing trace of inferred thrust fault in Upper Cretaceous Cokedale Formation (Livingston Group). The strata change dip from west-dipping, overturned beds to east-dipping, upright beds as the east limb of the Bridger uplift "rolls out" into the Crazy Mountains Basin.

0.8

5.9 Crossing trace of inferred thrust between Upper Cretaceous Miner Creek and Billman Creek Formations (both in Livingston Group). Quaternary alluvium of Bridger Creek floodplain at 3:00. Green Mountain (elevation 6,867 ft.) at 11:30.

0.6

6.5 Highway curves left. Intersection with Kelly Canyon Road on right; continue on Montana Highway 86. Place Creek at 9:00 and Bridger Canyon Community Center (old Lower Bridger School) at 3:00. As the road turns northeast, it follows the approximate contact of the Billman Creek Formation on the left and younger alluvium on the right. For the next several miles, the highway follows a scenic strike valley in the Billman Creek Formation.

1.5

8.0 Beasley Creek Valley at 9:00, and a view of the southern Bridger Range crest from 10:00 to 11:00 (elevation ranges from 8,000 to 9,000 ft.), with Mississippian Madison Group cropping out along the crest. Bridger Creek floodplain at 3:00, with hills between composed of east-dipping, Upper Cretaceous Hoppers Formation (Livingston Group).

1.1

9.1 Road curves right. Miner Creek Formation in roadcut.

0.3

9.4 Jackson Creek Road on right; continue north on Highway 86. Tree covered hills to left composed of Miner Creek Formation.

2.0

11.4 Road crosses Bridger Creek.

0.3

11.7 Stone Creek Valley at 3:00 cuts through the Billman Creek and Hoppers Formations. Beautiful panoramic view of the east flank of the Bridger Range from 9:00 to 11:00.

0.6

12.3 Cretaceous Billman Creek Formation in roadcut on left, dipping 50-60 degrees east. Alluvium in valley to left, with low, timbered hills in background (west) composed of Jurassic and Cretaceous strata rolling out with an east-dipping attitude at the base of the Bridger Range.

2.1

14.4 Olsen Creek Road on right.

1.0

15.4 Flaming Arrow Cross Country Ski Lodge on left.

0.5

15.9 Road to Bridger Bowl ski area on left.

0.2

16.1 Billman Creek Formation in road cuts.

0.5

16.6 Pavement ends.

2.4

19.0 Intersection of Bridger Canyon and Brackett Creek roads. Confluence of South and Middle forks of Brackett Creek. Bear left across the bridge and continue north on Highway 86.

1.9

20.9 Battle Ridge Pass (elevation 6,372 ft.). Battle Ridge Campground to right. From here, the road cuts down-section through northeast-striking, northwest-dipping (75 degrees, overturned) Livingston Group on the northwest side of Battle Ridge monocline. Battle Ridge monocline marks the southern boundary of the Helena salient of the fold-thrust belt, and may be underlain by a ramped thrust and offset Archean basement (down-to-the-north); Battle Ridge also separates the central Montana alkalic igneous province from the calc-alkaline province. Purple mudstones and light-colored sandstones seen in road cuts are in upper Billman Creek Formation.

1.0

21.9 TURN LEFT on Fairy Lake Road. Faulted and folded beds of upper Miner Creek Formation on both sides of road. The road to Fairy Lake is usually rough, so drive with caution. Exposures of Cretaceous and Jurassic strata between the junction with Highway 86 and Fairy Lake are poor due to forest cover, incompetence of the bedrock, thick soil and glacial deposits.

END OF ROAD LOG

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GOLDEN SUNLIGHT MINE - BUTTE FIELD TRIP

by

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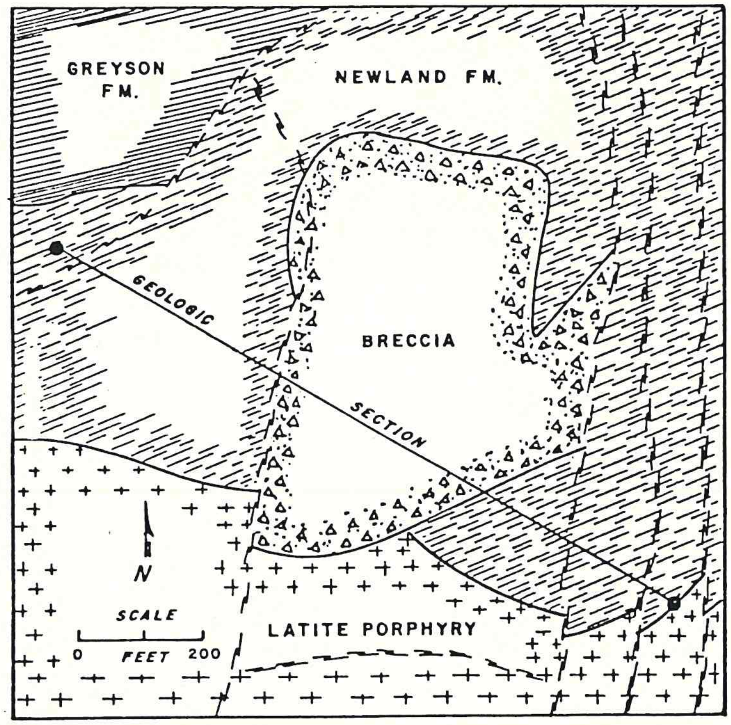
Leave from the parking lot of the Gran Tree Inn and drive west on Interstate 90 to the Cardwell exit 55 miles (88.5 km) west of Bozeman. Turn north at this exit and go 2.8 miles (4.5 km) west on frontage road on the north side of Interstate 90 to the graded road to the Golden Sunlight Mine.

Stop 1 - Guided tour of the Golden Sunlight Mine - The following brief description of this mine is from information provided by Golden Sunlight Mines, Inc. which is wholly owned by Placer U.S., Inc.

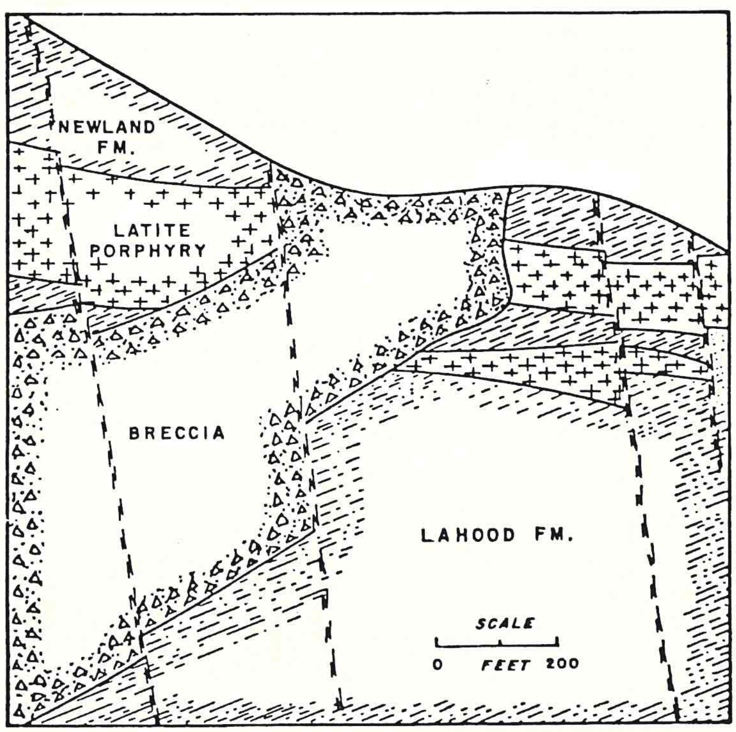
The first claims were staked on this gold deposit in 1890. Golden Sunlight Mines began open-pit mining of this deposit in December 1982 and poured their first gold in February 1983. Grade of the ore is 0.05 oz (1.4 gr) of gold per ton (907 kg) and indicated reserves are 25.8 million tons (23.4 billion kg). The stripping ratio is 2 to 1. The cyanide carbon-in-pulp method is used to recover gold in this 5000 ton(4.5 million kg)-per-day mill. As shown in the simplified geologic map and cross section (Figure 1) the Greyson and Newland Formations of the Precambrian Belt Supergroup have been intruded by latite porphyry of Eocene age which is cut by a breccia pipe. Gold in micron-size particles occurs in the breccia pipe and surrounding units. The Golden Sunlight Mine is the largest gold producer in Montana at this time.

Return to Interstate 90 and proceed west to the Montana Street exit in Butte. Go north on Montana Street to Park Street and then west on Park Street to Excelsior Avenue (see Figure 2). Go north on Excelsior Avenue to Missoula Avenue. The spectacular head frame of the Anselmo Mine is just east of Excelsior Avenue above Caledonia Street. The hoist house and idler towers for the cable also remain standing. The Anselmo Mine produced zinc, copper and silver. Uphill from the Anselmo Mine Excelsior Avenue crosses the tracks of the Butte, Anaconda and Pacific Railroad which although never reaching the Pacific Ocean, hauled a large quantity of ore and concentrate from the Butte mines to the smelter in Anaconda.

Some idea of the immensity of the Butte district can be gained from the following example: If all of the copper recovered from the Butte ores was used to pave Interstate 90 with a 6-inch (12.7 cm) thick pavement 156 feet (47.5 m) wide, this copper pavement would extend from Butte to Bozeman. The lead, zinc and silver recovered from the Butte ores would extend this pavement another 53 miles (85.3 km). Butte has produced enough gold to make a cube a little over 3 feet (0.9 m) on an edge (2,922,500 troy ozs or 90,900 kg).



SURFACE GEOLOGIC MAP OF PIT AREA



GEOLOGIC SECTION LOOKING NE THROUGH PIT AREA

Figure 1. Generalized geologic map and cross section, Golden Sunlight mine (Lusty and Roper, 1983, p. 11)

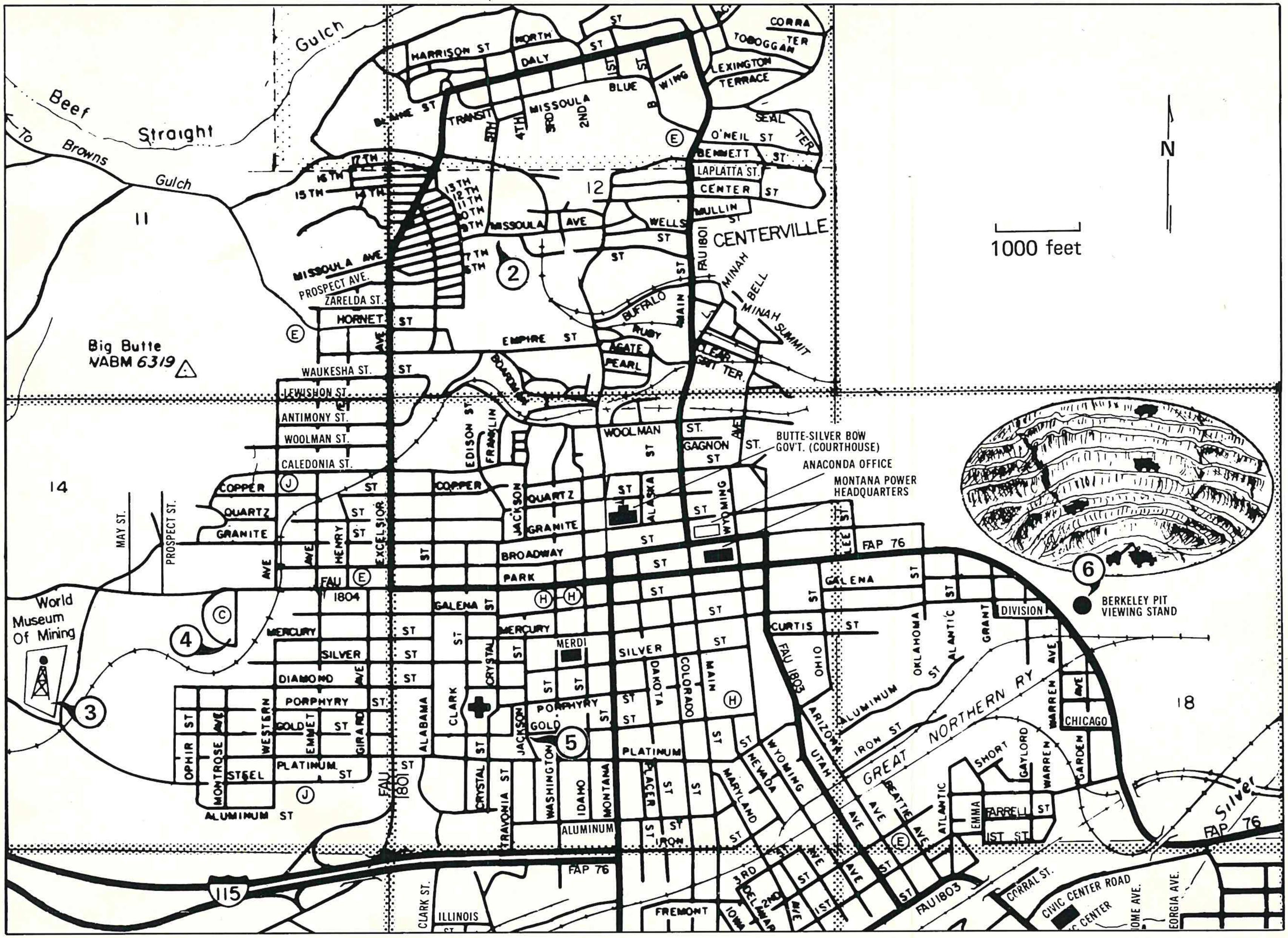


Figure 2. Map of the northwest part of Butte showing field trip stops.

Stop 2 - Butte overview - Turn right (east) on Missoula Avenue and go for 0.25 miles (0.4 km) to a baseball diamond where we will eat lunch and examine the blocks of waste rock east of the diamond. These blocks of waste may have come from a small open pit a few blocks to the south. They contain veinlets of pyrite and sphalerite. Choice small sphalerite crystals can be collected here. From this vantage point, from which much of the Butte District can be seen, Lester Zeihen will give a brief overview of the geology of this district.

Stop 3 - World Museum of Mining - Retrace the route back down Excelsior Avenue to Park Street (the second stop light). Turn right (west) on Park Street and go uphill across the BA&P tracks, past the statue of Marcus Daly and past the campus of the Montana College of Mineral Science and Technology to the World Museum of Mining at the Orphan Girl Mine (closed in 1958). Exhibits feature mining equipment, methods and a look at life around the turn of the century in the Butte District. Admission is free and summer hours (June through Labor Day) are from 9 a.m. to 9 p.m. daily. In spring and fall the Museum is open from 10 a.m. to 5 p.m. except on Mondays. The World Museum of Mining is closed during the winter.

Just to the west and north of the World Museum of Mining, Aires Resources is drilling and trenching silver veins that were mined underground at the Orphan Girl Mine. A drift connected the workings of the Orphan Girl Mine with those of the Anselmo Mine.

Stop 4 - Mineral Museum - Return to the campus of Montana College of Mineral Science and Technology and turn right at the sign for Mineral Museum parking. Although minerals from worldwide localities are displayed in the Mineral Museum, the displays of Butte minerals and of Montana minerals are of special interest. More than 120 different minerals have been identified from the Butte District. Admission to the Mineral Museum is free and hours are from 8 a.m. to 5 p.m. weekdays and from 1 to 5 p.m. on Sundays. From June 1 through Labor Day the Museum is open from 8 a.m. to 5 p.m. every day.

Stop 5 - Emma Vein - Return to Park Street and proceed east to Excelsior Avenue. Turn right (south) on Excelsior Avenue and go south to the first stop light at Platinum Street. Go east on Platinum Street up over the hill to Jackson Street and turn north on Jackson Street going one block north to Gold Street. Rusty brown exposures of the Emma Vein can be seen on the west side of Jackson Street. The Emma Vein was mined for manganese from the Emma Mine about 0.3 miles (0.5 km) northeast of this exposure at the southeast corner of Silver and Dakota Streets (Figure 3). The head frame has been torn down and the shaft has been bulkheaded. The finest rhodochrosite specimens from the Butte District are from the Emma Mine.

As exposed here, the Emma Vein consists of fine-grained quartz, jasper and black manganese minerals. Poorly developed boxworks and small irregular cavities indicate that some minerals, perhaps rhodochrosite, have been leached from this vein. The vein is locally brecciated. Irregular masses of white sericite of hydrothermal origin

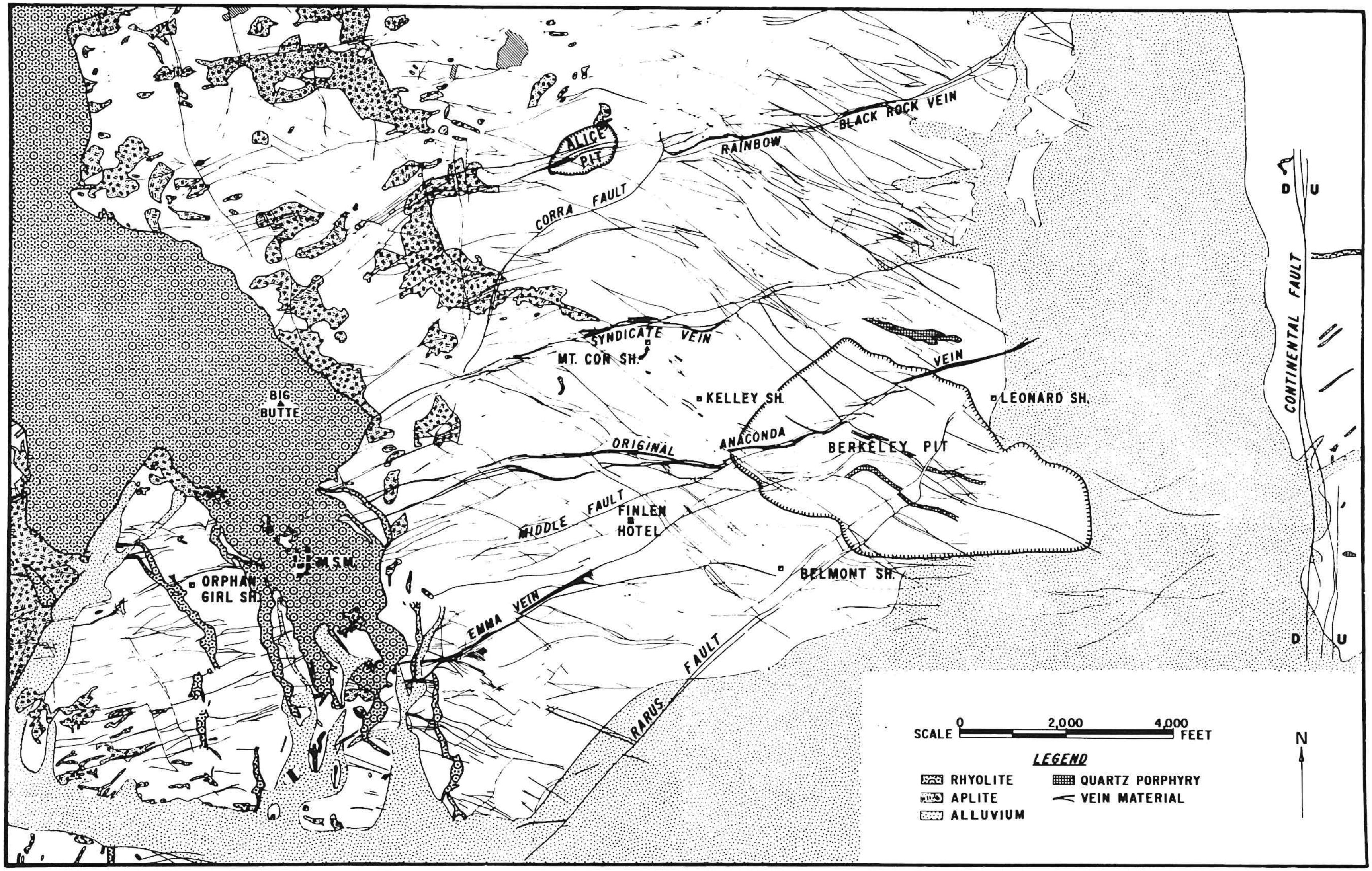


Figure 3. Geologic map of the Butte district (Meyer and others, 1968, p. 1380).

are exposed on the south side of the vein. Fractures in the less altered granite of the Butte Quartz Monzonite, one of the plutons of the Boulder Batholith, are coated with black manganese minerals.

Stop 6 - Berkeley Pit - Return south on Jackson Street to Platinum Street and then turn east (left) and continue on Platinum Street to Arizona Avenue. Turn north (left) on Arizona Avenue and continue to Park Street. Turn right (east) on Park Street and go east to the Berkeley Pit viewing stand. The Clyde E. Weed concentrator is to the southeast of the viewing stand. At the viewing stand, Ted Duaime, hydrologist with the Montana Bureau of Mines and Geology, will discuss the hydrological aspects of the flooding of the Butte Mines.

From the Berkeley Pit Overlook, drive east on Continental Drive over Interstate 90 to the Continental Drive interchange. Go east on Interstate 90 to Bozeman.

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Lusty, Q.C. and Roper, M.W., 1983, Golden Sunlight Mines, Inc., Montana, Placer U.S. Inc. in An in-depth study of 5 new silver and gold mines, Northwest Mining Association Short Course, 46 p.

Meyer, C., Shea, E.P., Goddard, C.C., Jr. and staff, 1968, Ore deposits at Butte, Montana, in John D. Ridge, ed., Ore deposits of the United States, 1933-1967, V. II, A.I.M.E., New York, p. 1373-1416.

NATURE OF DEFORMATION IN FORELAND ANTICLINES AND IMPINGING THRUST BELT: TOBACCO ROOT AND SOUTHERN HIGHLAND MOUNTAINS, MONTANA

by

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INTRODUCTION

This paper and field trip guide has two purposes: 1) to review some of the deformational features associated with basement-cored anticlines of the Laramide Rocky Mountain foreland in southwestern Montana, and 2) to examine the effect of these structures on the impinging frontal portions of the Cordilleran thrust belt. Much of the information contained in the discussion which follows and many of the interpretations, have been previously published or presented at meetings (see for example, Schmidt and Hendrix, 1981; Schmidt and O'Neill, 1982; Schmidt and Garihan, 1983; Schmidt, 1983; Brandon, 1984; Schmidt and others, 1984; and Geiger, 1985). However, most of the data and interpretations for small-scale structures (cleavage, fractures, and folds) related to thrust impingement against foreland anticlines is new. In addition, with the exception of the Pole Canyon Anticline (Stops 1 and 2) the specific features we will visit and discuss have been observed by comparatively few geologists. Construction of a U.S. Forest Service road across heretofore remote portions of the Pole Canyon foreland and thrust belt structures, recent M.S. thesis research (W. Brandon and B. Geiger) in the McCartney Salient, and mapping and structural analysis in the southern Highland Mountains (J. M. O'Neill and C. Schmidt) provided the impetus for showing these features to others.

FORELAND STRUCTURES

One of the two principal structural elements of the Laramide Rocky Mountain foreland is a set of approximately 30 major northwest-trending faults which dip steeply northeast (Schmidt and Garihan, 1983)(Fig. 1). The faults transect both Archean metamorphic rocks and the Phanerozoic section, are spaced 6 to 10 km (3.7 to 6.2 mi)apart, and have separations measured on the Archean-Cambrian unconformity which range from tens of meters to several kilometers. A Middle Proterozoic inheritance of these faults is demonstrated by: 1) the presence of diabase dikes dated at 1455 m.y. and 1130 m.y. in the fault zones, 3) separations of Archean rock units along the faults which differ from separations of the Archean-Cambrian unconformity, and 4) local control of depositional patterns of the Lahood Formation along the southern margin of the Belt Basin (Reid, 1957; Wooden and others, 1978; Schmidt and Garihan, 1983; 1985; O'Neill and others, in Early Tertiary time was oblique with roughly equal components of review). Slip on the northwest-trending fault set in Late Cretaceous and left-lateral and reverse movement (Schmidt and Garihan, 1983).

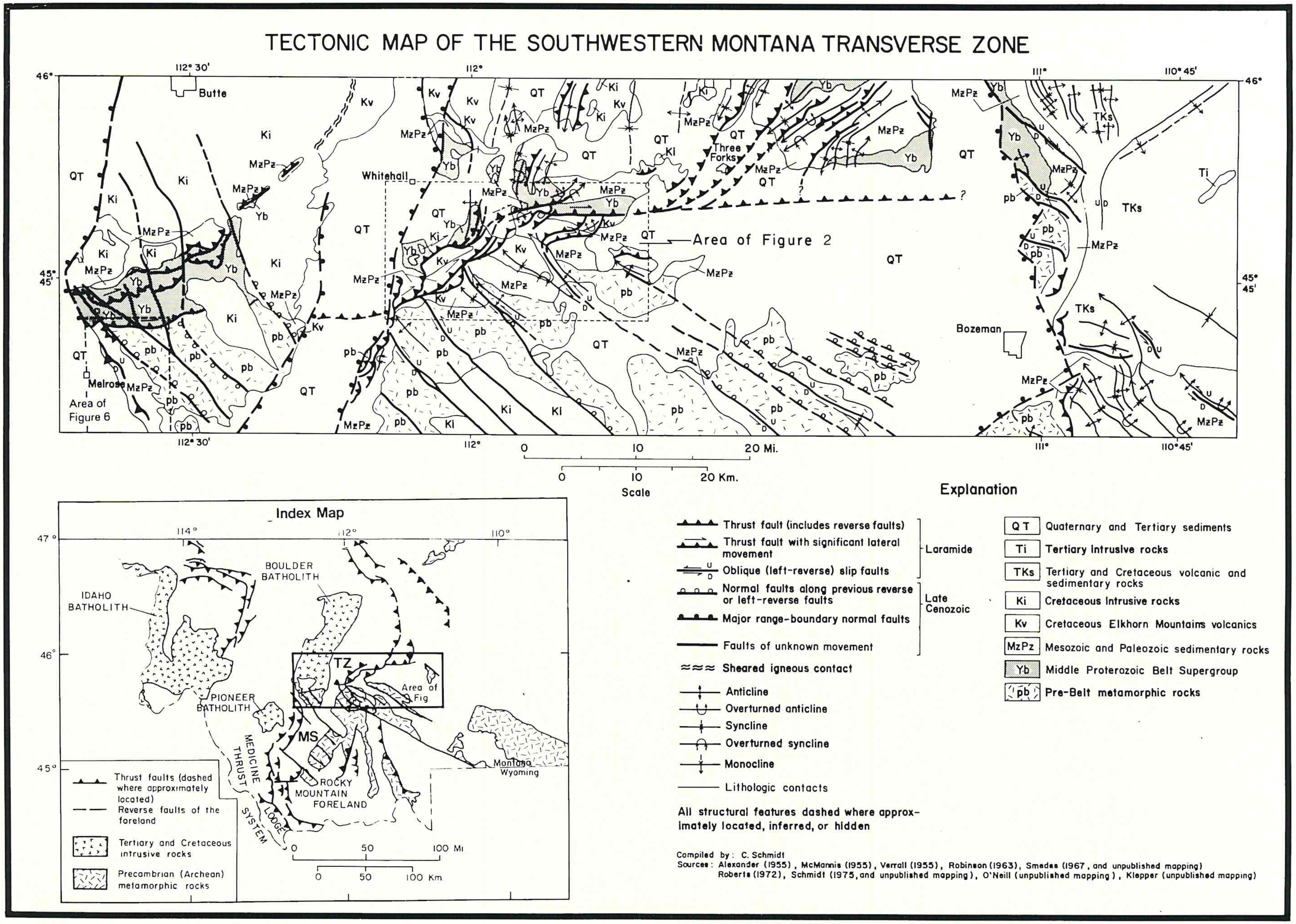

Figure 1. Tectonic map of the Rocky Mountain foreland and thrust belt overlap in the transverse thrust zone (TZ) and McCartney Salient (MS) areas of southwestern Montana.

The field trip route was planned to enable participants to examine two of these foreland structures (Pole Canyon Anticline and London Hills Anticline). Another foreland anticline (Rochester Anticline) is located in the southern Highland Mountains and had an important influence on the thrust belt structures we will examine at Stops 5, 6, and 7.

In most cases the foreland anticlines adjacent to the frontal portions of the fold and thrust belt were developed somewhat before thrusting (Coniacian-middle Campanian time).

THRUST BELT STRUCTURES

The Cordilleran thrust belt structures which we will see are located at the leading edge of the frontal thrust zone in southwestern Montana (Ruppel and Lopez, 1984). The two principal geometric elements of that portion of the frontal thrust zone included in the field trip itinerary are: 1) the southwestern Montana Transverse Zone and 2) the McCartney Salient (Fig. 1).

The transverse zone is characterized by an anastomosing system of thrusts which change trend frequently from east-west to north-south (Schmidt and Hendrix, 1981; Schmidt and O'Neill, 1982). The thrusts dip northerly and westerly with comparatively steep dips at the surface (35-75°). Slip on north-trending segments is generally reverse dip slip, whereas the more east-trending segments show a strong component of dextral slip (Schmidt, 1975). The aggregate eastward transport of the principal thrust sheet and several minor ones was at least 15 km (9.3 mi)(Schmidt and O'Neill, 1982, p. 214). We will examine the interaction of the southernmost thrusts of the transverse zone in the Tobacco Root Mountains with the Pole Canyon Anticline at Stops 2 and 3.

The McCartney Salient is defined by well-developed folds and thrusts in a convex arc immediately to the southwest of the transverse zone (Brumbaugh, 1973)(Fig. 1). The structural position and geological relationships of the salient are analogous to those in the Disturbed Belt of northwestern Montana and the Utah-Wyoming Thrust Belt east of the Paris-Willard Thrust (Brandon, 1984; Lopez and Schmidt, 1985). Brumbaugh (1973) has shown that thrust movements were normal to the curved eastern outline of the salient. This radial transport direction has been ascribed to the influence of the Pioneer Batholith as a driving force (Brumbaugh, 1973) and to the "buttressing" effect of an irregular foreland margin (Beutner, 1977; Brandon, 1984; and Schmidt and others, 1984). At Stops 5, 6 and 7 in the northern part of the salient, we will examine several of the mesoscopic effects of this buttressing and attendant thrust-sheet rotation.

ROAD LOG AND STOP DESCRIPTION

Detailed road logs are available which describe features along much of the field trip route (Chadwick, 1981; Johns and others, 1981; Lageson and Montagne, 1981; Garihan and others, 1982; and Lageson and others, 1983). We have made no attempt to reproduce these descriptions here, nor have we attempted to provide complete detailed descriptions between those stops not covered in earlier road logs.

We begin mileages and brief descriptions from the intersection of Interstate 90 and U.S. Highway 10 north of Cardwell (Fig. 2). Mileages are approximate.

Cumulative Mileage mi (km)	Increment Mileage mi (km)	
0.0 (0.0)	0.0 (0.0)	Proceed south on County road 359.
0.7 (1.1)	0.7 (1.1)	Passing through Cardwell
1.1 (1.8)	0.4 (0.6)	Crossing Jefferson River. Road turns left (east). LaHood Park can be seen across the Jefferson River. Outcrops are Proterozoic LaHood Formation on the Tobacco Root-Jefferson Canyon Thrust Sheet.
1.6 (2.6)	0.5 (0.8)	Road bends right (south through the community of Jefferson Island).
2.9 (4.6)	1.3 (2.1)	Hill on immediate left is underlain by Renova Formation (Oligocene) dipping east toward a late Cenozoic fault (Starets Ditch Fault) which bounds the western side of London Mountain. London Mountain (higher ridges across the South Boulder River) on the left (east) is the northern extent of the London Hills Anticline (Fig. 2).
5.2 (8.3)	2.3 (3.7)	Turn right on South Boulder Road (sharp curves)
6.3 (10.1)	1.1 (1.8)	Sharp bend to left (south). Outcrops of Elkhorn Mountains Volcanics (Cretaceous) occur on right (east) side of road.
8.0 (12.8)	1.7 (2.7)	Pavement ends at T intersection with Armstrong Lane. Continue south. Entering South Boulder River Canyon. Northeast-dipping upper Paleozoic and Mesozoic section crops out on both sides of the road. These rocks and the remainder of the section down to the Archean metamorphic complex make up the gentle limb of the Pole Canyon Anticline which we will see at Stop 1.

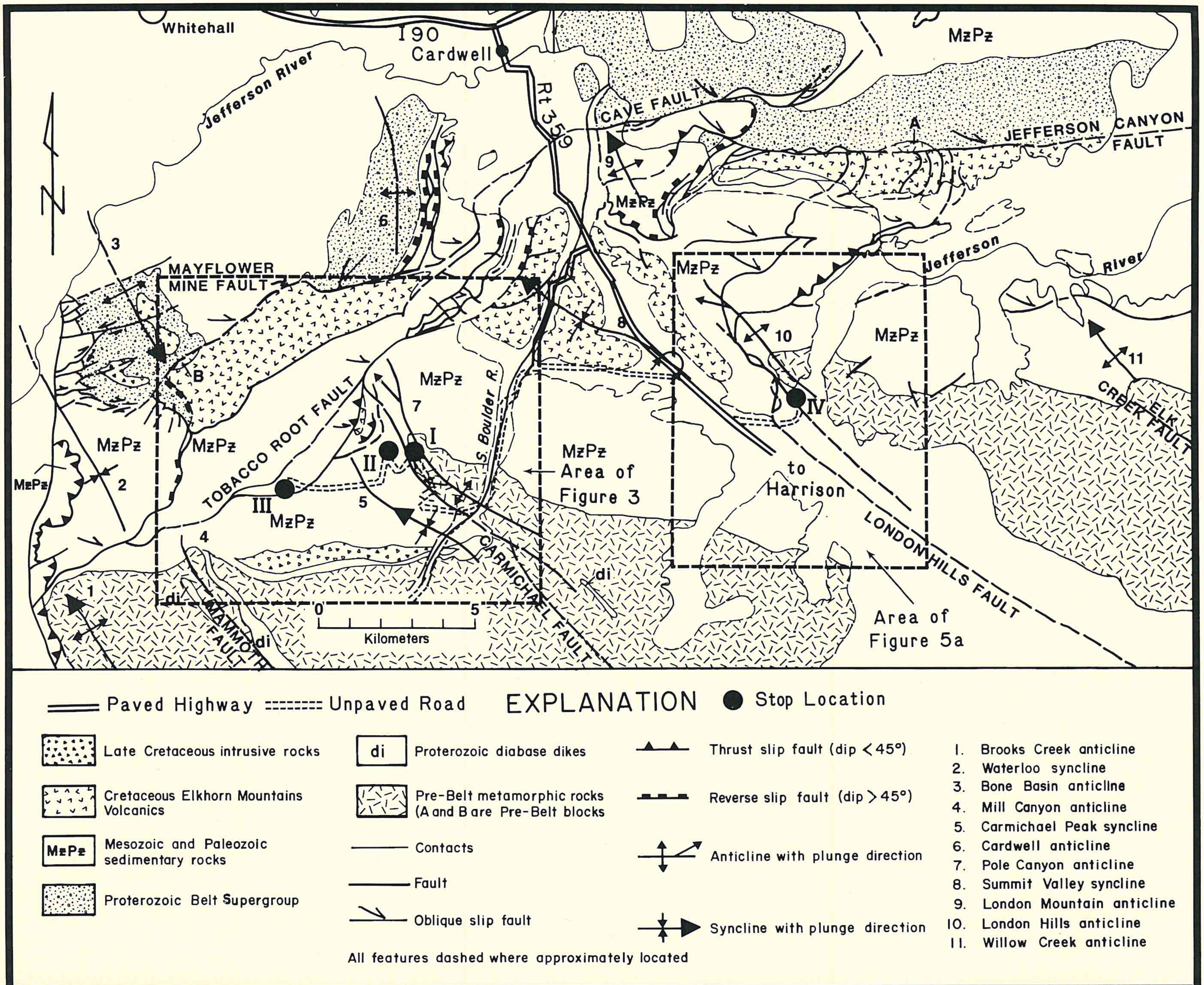

Figure 2. Tectonic map of the north-central Tobacco Root Mountains showing field trip route and Stops 1-4.

9.1 (14.6) 1.1 (1.8)

Cross South Boulder River. Outcrops of Mission Canyon Limestone (Mississippian) occur in massive cliffs on both sides of the road. Proceed south and continue going down section.

10.3 (16.5) 1.2 (1.9)

Approximate contact between Cambrian Flathead Sandstone and Archean metamorphic rocks.

11.4 (18.2) 1.1 (1.8)

Road to Indiana University Geologic Field Station on left.

11.6 (18.6) 0.2 (0.3)

Trace of the Carmichael Fault can be seen on either side of the road. On the right (west) Archean metamorphic rocks on the hanging wall are in contact with a syncline in Devonian and Mississippian rocks. A small horse block of Cambrian limestone is found between the Archean rocks and the Jefferson Dolomite (Devonian) immediately west of the road.

11.8 (18.9) 0.2 (0.3)

Turn right (west) on Brownback Road and cross cattle guard. The road follows a sill in the Jefferson Dolomite.

12.5 (20.0) 0.7 (1.1)

Switchback in road. A locked Forest Service gate is located about 0.1 mi (0.2 km) ahead. Proceed along road with caution after passing through gate. Carmichael Peak on left (northwest) is composed of Mission Canyon Limestone, sharply folded in a disharmonic anticline, on the footwall of the Carmichael Fault (Fig. 3). Smaller scale disharmonic folds in the core of this anticline occur in the Lodgepole Limestone along the road.

13.3 (21.2) 0.8 (1.3)

Road crosses Carmichael Fault near here and proceeds parallel to it for about 0.5 mi (0.8 km).

14.0 (22.4) 0.7 (1.1)

Stop 1. Park just before cattle guard. We are very close to the Carmichael Fault here. Blocks of Flathead Sandstone occur on the north side of the road and Mission Canyon Limestone crops out upslope to the south. We will make a short hike through the woods here to outcrops of diabase parallel to the Carmichael Fault. In addition to examination of the diabase and the Archean metamorphic lithologies, this stop will provide an excellent down-plunge view of the Pole Canyon Anticline on the hanging wall of the Carmichael Fault.

Discussion of Stop 1

The Pole Canyon Anticline is one of the best exposed foreland structures in southwestern Montana. Its mean axial orientation is 35° N30°W (Fig. 3), but it is not perfectly cylindrical and therefore has

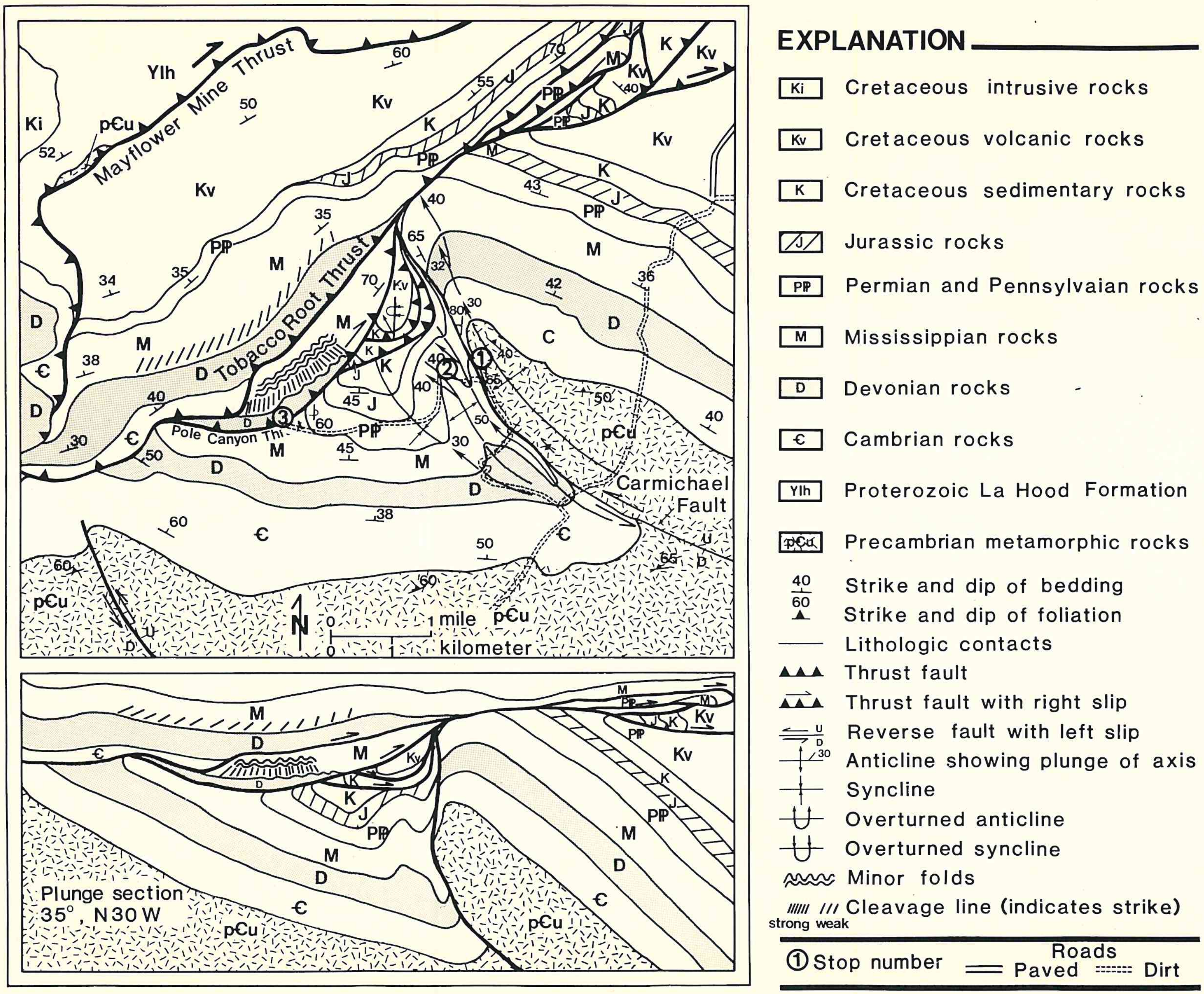

Figure 3. Geologic sketch map and down plunge section of the Pole Canyon Anticline and impinging Tobacco Root and Pole Canyon Thrust Sheets.

slightly different orientations of trend and plunge in different units. The contact between the Archean metamorphic rocks and the overlying Flathead Sandstone can be seen across Pole Canyon in the foreground. Although highly fractured, this contact defines a relatively smoothly curved arc around the fold. The Flathead dips about 50°W on the steep limb.

The Archean rocks in the core of the fold are not well-exposed and the basement-cover contact is not observed. However, foliation within the hinge region is bent nearly concordantly with the overlying cover rocks (Fig. 4a), suggesting that "basement" has been folded along foliation to accommodate folding in the cover rocks.

The Paleozoic section through the Mission Canyon Limestone shows varying mechanical behavior. All units are thinned along the steep limb. The Cambrian shales have been most noticeably thinned by cataclastic flow. The Park Shale, for example, has been totally removed and the limestone units above and below the Park (Pilgrim Limestone and Meagher Limestone) are in contact. Thick competent carbonates like the Mission Canyon Limestone are also moderately thinned and have deformed by locally intense fracturing and intracrystalline gliding. Sandstones and dolomites are intensely fractured and brecciated. Large, isolated boudin-like blocks of Devonian Jefferson Dolomite, Mission Canyon Limestone, and Pennsylvanian Quadrant Sandstone occur in the fault zone.

The principal footwall syncline is disharmonically folded into three large northwest-plunging folds on the footwall of the Carmichael Fault (Fig. 3). These folds have been interpreted as having been related to crowding upward in the major footwall syncline accentuated by compressional shortening of the beds against the hanging wall of the fault (Schmidt and Garihan, 1983).

The Carmichael Fault is not well exposed here, but it forms the obvious northwest-striking valley. The fault dips 50-70°NE. The slip direction was inferred to have been left-reverse oblique with the reverse dip slip component slightly greater than the strike slip component (Schmidt and Garihan, 1983). Net slip was 2-2.5 km (1.2-1.6 mi).

The diabase dike exposed here is one of nearly 100 such dikes which parallel the regional set of northwest-trending faults. This particular dike, as mapped by Jack Garihan, Furman University, can be traced parallel to the Carmichael Fault for slightly over one mile (about 2 km; 1.2 mi). The dikes in this northwest-trending set have been dated elsewhere as middle Proterozoic (Wooden and others, 1979). They are inferred to have been associated with Proterozoic movement on the northwest-trending faults (Schmidt and Garihan, 1979; 1985). There has been some debate about the nature of this early movement, and it is probably most accurate to say that neither its magnitude nor its direction is known.

Return to vehicles and proceed along Forest Service road.

14.6 (23.4)   0.6 (1.0)

Stop 2. Stop at first sharp counter-clockwise bend in the road. This stop will provide an overview of the

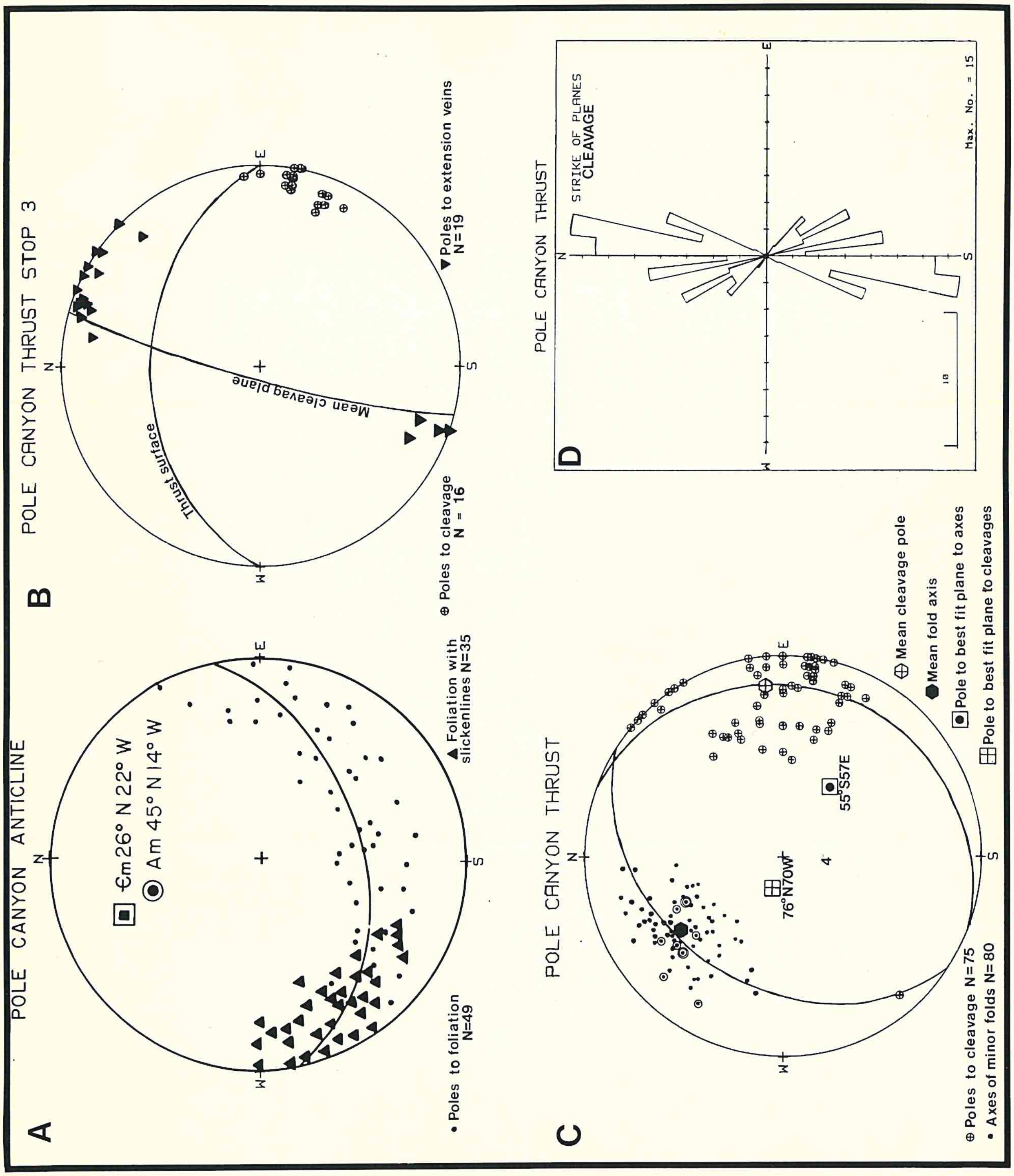

Figure 4. A) Lower hemisphere, equal area plot of foliation in the core of the Pole Canyon Anticline. indicates axis of folded foliation, is position of fold axis in Cambrian Meagher Limestone.

B, C) Equal area plots of mesoscopic data for the Pole Canyon Thrust Sheet.

D) Rose diagram of strike of cleavage planes, Pole Canyon sheet.

relationship of the Pole Canyon Anticline and thrusts of the southwest Montana Transverse Zone. Outcrops are limestones in the Amsden Formation (Pennsylvanian).

Discussion of Stop 2

The nature of the interaction between thrust belt structures and the Pole Canyon Anticline may be observed from this location. The principal thrust (Pole Canyon Thrust) may be traced along the base of the prominent cliffs of Mission Canyon Limestone (across Pole Canyon to the north). The footwall rocks are mainly Elkhorn Mountains Volcanics, Colorado and Kootenai Formations (all Cretaceous). Smaller splays from the main thrust are found in Cretaceous rocks which occupy the valley (Fig. 3). The Pole Canyon Thrust strikes northeast and dips 45° NW here. Its movement was right reverse (oblique) with nearly equal component of right slip and reverse dip slip (Schmidt, 1975; Schmidt and Hendrix, 1981). The Pole Canyon Thrust rejoins the structurally higher Tobacco Root Thrust at the hinge of the Pole Canyon Anticline.

The curved trace of the Pole Canyon Thrust, and that of the minor splays, has been inferred to have been a result of the impingement of the thrust against the Pole Canyon Anticline and the formation of a thrust ramp over the anticline (Schmidt and Hendrix, 1981; Schmidt and O'Neill, 1982). An alternative would be that the thrusts were folded during the formation of the Pole Canyon Anticline. The first hypothesis is more likely inasmuch as the hinge region is clearly truncated by the Pole Canyon and Tobacco Root Thrusts. Furthermore the hanging wall/footwall relationships in the Pole Canyon Thrust change abruptly across the foreland structure, from Mission Canyon Limestone over Cretaceous Elkhorn Mountains Volcanics to Mission Canyon on Mission Canyon at the hinge of the anticline (Fig. 3). This clearly supports the hypothesis that the Pole Canyon Anticline was present prior to thrusting and deflected the path of the Pole Canyon Thrust. The minor structures on the Pole Canyon Thrust sheet also support this interpretation. These minor structures will be examined and discussed at stop 3. Return to vehicles. Proceed to end of present Forest Service road. The road crosses the Pole Canyon Thrust near the outcrops of Mission Canyon Limestone about 2 mi (3.2 km) from Stop 2.

16.8 (26.9) 2.2 (3.5)

Stop 3. Park on grassy saddle at the head of Pole Canyon after crossing cattle guard. The saddle is underlain by the Three Forks Shale (Devonian) on the hanging wall of the Pole Canyon Thrust. The high peak immediately to the south is composed of Mission Canyon Limestone on the footwall of the thrust. The Jefferson River Valley and Highland Mountains may be seen to the west. The purpose of this stop is to examine the cleavage development in the micritic lower portion of the Mississippian Lodgepole Limestone and (time permitting) to examine several folds developed in the more sparry upper portion of the Lodgepole. The cleavage is easily seen at this stop, but a short hike along the ridge will be required to examine the folds.

Discussion of Stop 3

The cleavage here in the lower part of the Lodgepole is disjunctive, narrowly to closely spaced (0.3-1 cm)(0.12-0.39 in), nonsutured to slightly undulatory, generally continuous, and wavy to slightly anastomosing (see morphological classifications of Powell, 1979, and Engelder and Marshak, 1985). Cleavage at this particular location has a fairly consistent orientation (N10-20E, 75-90 NW) (Fig. 4b). Calcite veins occur at several orientations but the most persistent is N60-80W, vertical. In thin section this vein set appears to be coeval, with cleavage development suggesting a relationship between the two structural elements and indicating a pressure-solution origin for the cleavage. Pressure solution is also indicated by minor amounts of insoluble material within the cleavage domains.

Cleavage orientations to the east along the thrust change to N40-50W, vertical (Fig. 4c and d). This change of orientation of the cleavage is associated with a change of strike of the thrust and the hanging wall rocks, from nearly east-west to N45E, due to deflection by the Pole Canyon Anticline. Cleavage in the micrites of the Lodgepole Limestone is also present in the next structurally higher thrust sheet (Tobacco Root Thrust Sheet). However, it is a much weaker fabric with spacing between cleavage domains generally greater than 1 cm (0.39 in) (Fig. 3).

Cleavage is generally absent in the more sparry upper units of the Lodgepole Limestone here and completely absent in the overlying Mission Canyon Limestone (compare this with Stop 7). However, the upper part of the Lodgepole has been highly folded. The folded units may be traced for nearly 1 km (0.62 mi) along the northeast-trending ridge on the wall of the thrust. There are over 60 folds with a mean axial orientation of 38°, N35W (Fig. 4c). Shortening by folding was determined by direct measurement to be 28-35%. Apparently folding was the principal mechanism of shortening here, whereas cleavage formation was the principal mechanism of shortening in the micritic units.

Our preliminary interpretation of the important structures of the Pole Canyon area is as follows:

1) The Pole Canyon Anticline formed in association with Laramide movement on the Carmichael Fault of Proterozoic ancestry.

2) The same west-east compression which produced left-reverse movement on the Carmichael Fault was responsible for right reverse movement on the thrust faults of the transverse zone. The Tobacco Root thrust immediately west of the present location (Stop 3) changed stratigraphic position and formed a ramp across the Devonian and Mississippian strata on the footwall.

3) The footwall rocks shortened against the uplifted hanging wall of the Carmichael Fault where the Mission Canyon Limestone in the hinge and steep limb of the Pole Canyon Anticline formed an effective buttress to thrust movement. The initial shortening was in a WNW-ESE direction and is reflected in the formation of solution cleavage in the micrites and folds in the biosparites of the Lodgepole Limestone. Shortening was about 30%.

4) The Pole Canyon Thrust formed as a footwall splay of the Tobacco Root thrust ahead of the ramp. As the Pole Canyon Thrust propagated eastwardly it was deflected from an easterly to a northeasterly trend by the Pole Canyon Anticline. The Pole Canyon Thrust rejoined the Tobacco Root Thrust after being deflected over the hinge of the anticline. The deflection of the hanging wall rocks in the Pole Canyon sheet caused the initial N10-20E cleavage orientations and northerly trends of minor folds to be rotated counter clockwise (westerly) more than 45°.

Return to vehicles and retrace field trip route back to the South Boulder Road. Take the South Boulder Road to its intersection with the Armstrong Lane (T intersection where pavement begins). Return to Cardwell via South Boulder Road and County Road 359 if time does not permit the optional route and optional stop (Stop 4).

25.6 (41.0) 8.8 (14.1) -Optional Route-

Turn right (east) onto Armstrong Lane. Elkhorn Mountains Volcanics crop out on left (north). Tree covered ridges on the right are the Upper Paleozoic rocks on the gentle limbs of the Pole Canyon Anticline.

28.3 (45.1) 2.7 (4.3)

Turn right (southeast) on County Road 359. London Hills Anticline is on the left. The road follows the approximate trace of the major northwest-plunging Syncline (Summit Valley syncline of Schmidt, 1975)(Fig. 2) between the London Hills Anticline and Pole Canyon Anticline.

30.2 (48.3) 1.9 (3.0)

Turn left onto dirt road (property is currently owned by Melvin Armstrong, Cardwell, and individuals must obtain permission to enter this property). Steeply dipping to overturned Upper Paleozoic and Mesozoic units on the steep limb of the London Hills Anticline are on the left. We will take this road as far as current crop and road conditions permit. A short hike will be required after we stop.

31.5 (50.4) 1.3 (2.1)

Stop 4. Park in field overlooking the creek (Dogtown Sewer). Archean rocks in the core of the London Hills Anticline are covered by pine trees (the overlying Cambrian section is not) locally known as the "Black Forest." We will cross into the core of the structure via the main drainage to examine outcrops of foliated Archean rocks below the folded Cambrian Flathead Sandstone.

Discussion of Stop 4

The London Hills Anticline is very similar to the Pole Canyon Anticline except that it is in a somewhat more advanced stage of development (Schmidt and Garihan, 1983; Brown, 1983)(Fig. 5a). The western limb is strongly overturned between two fault segments and fault separation is greater than on the Carmichael Fault. Two of the more notable things about the fold is the presence of vertical to overturned Flathead Sandstone

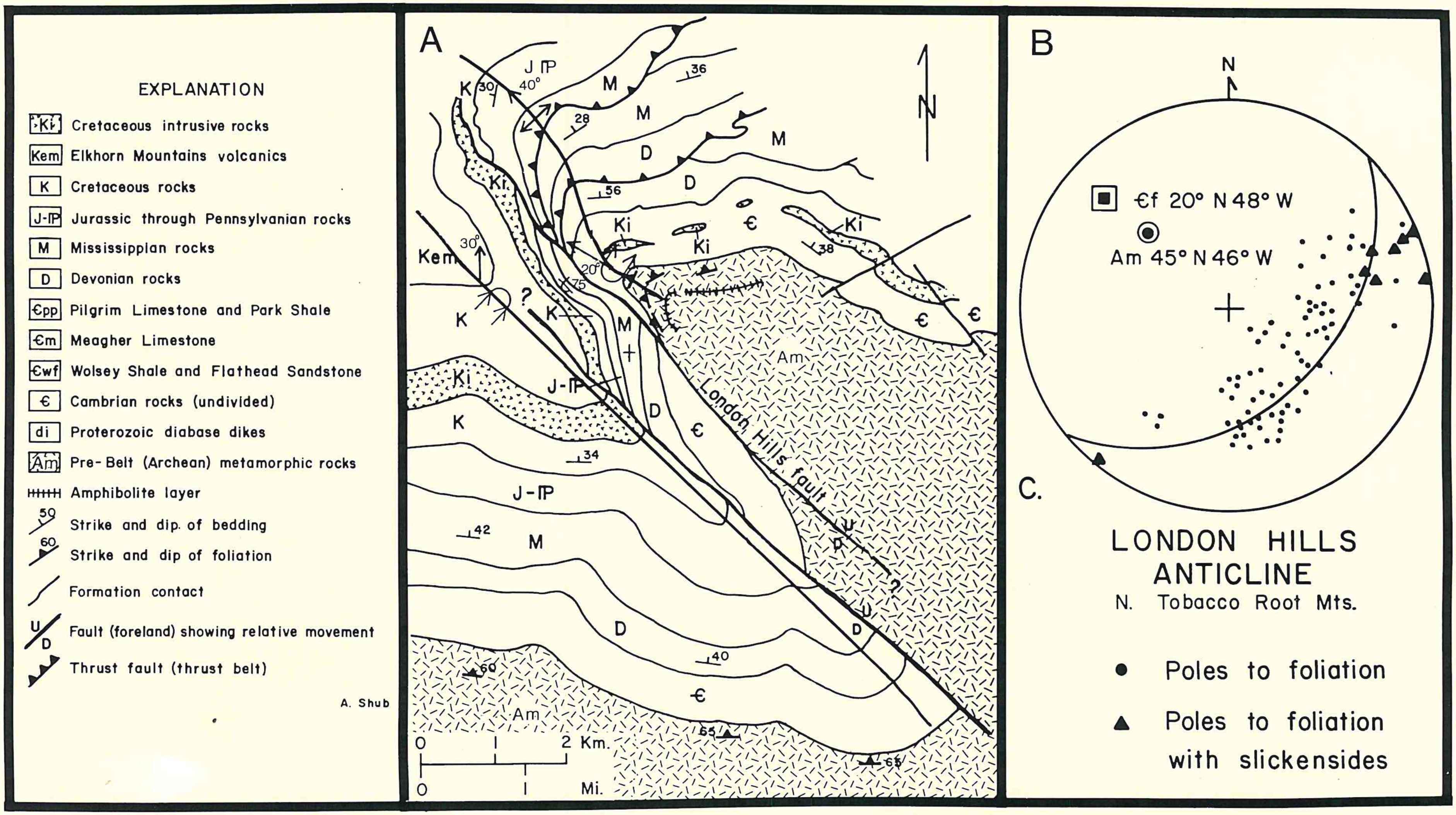

Figure 5. Sketch map of London Hills Anticline and lower hemisphere equal area plot of poles to foliation in the core of the London Hills Anticline.

beds on the steep limb of the anticline and very good evidence of folded foliation in the Archean rocks in the core of the fold.

Wagner (1966) first studied the Archean rocks in the core of this fold. Besides identifying and mapping the major lithologies (biotite gneiss, hornblende gneiss, quartzofeldspathic gneiss and amphibolite) she noted that the foliations followed the general shape of the major anticline and concluded that this early-developed fabric had been folded during formation of the anticline. Schmidt and Garihan (1983) showed that the axis of folded foliation has nearly the same trend as that for the folded Flathead Sandstone and a plunge which is steeper by 25° (Fig. 5b). This strongly suggests that foliation and bedding were folded together and that the difference in plunge represents the initial angular discordance between bedding in the cover and foliation in the Archean rocks. Foliation planes are strongly sheared in places and have developed slickenside striae and local chloritic alteration of biotite.

Return to vehicles and retrace route to County Road 359.

32.8 (52.5)   1.3 (2.1)

County Road 359. Turn right and proceed to Cardwell and intersection with U.S. Highway 10.

4.3 (68.8)   10.2 (16.3)

Intersection with U.S. Highway 10. Take U.S. 10 (under Interstate 90) and continue towards Whitehall and Twin Bridges. In Whitehall take Montana Highway 55 south. At the junction of Highway 55 and 41 continue south on 41 towards Twin Bridges.

75.2 (120.3)   33.2 (53.1)

Entering Twin Bridges.

75.9 (121.4)   0.7 (1.1)

Turn right at main intersection on Highway 41 towards Dillon.

76.0 (121.6)   0.1 (.2)

Crossing Beaverhead River.

76.4 (122.2)   0.4 (0.6)

Turn right (north) at the bend after passing a retirement home (former State Children's Center).

77.6 (124.2)   1.2 (1.9)

Bridge across Big Hole River. Pavement ends.

78.5 (125.6)   0.9 (1.4)

Road forks. The right fork goes into the Rochester Mining District in the southern Highland Mountains (see road log by Garihan and others, 1982). Take left fork toward Melrose. Rocks for the next 10 mi (16 km) are Archean gneisses capped with Tertiary sediments. McCartney Mountain on the south is a stock intruded into Mesozoic sedimentary rocks. Its age (73-75 Ma) (Brumbaugh and Hendrix, 1981) places a "youngest age" limit on the easternmost thrusting in the McCartney Salient.

89.3 (142.9) 10.8 (17.3)

Turn right (north) on unimproved dirt road. This road is frequently impassable due to wash-outs. However the first stop (Stop 5) is easily accessable. The turn-off is indicated on Figure 6.

89.7 (143.5) 0.4 (0.6)

Stop 5. Gather for a brief discussion of the Southern Highlands Fault Zone and McCartney Salient. Proceed a short distance down the creek bed to excellent exposures of Lodgepole Limestone. The principal purpose of this stop is to examine small scale deformation features (fractures and cleavage) associated with shortening of the Paleozoic section against a basement-cored foreland anticline.

Discussion of Southern Highlands Fault Zone and Features at Stop 5

Stop 5 is located at the southernmost exposures of a zone of northwest-trending folds and minor thrusts in Paleozoic rocks at the southwestern margin of the Highland Mountains. We will refer to this zone as the Southern Highlands Fault Zone (SHFZ) although the actual fault displacements are relatively small.

The SHFZ forms the northern boundary of the McCartney Salient of Brumbaugh (1973). It may be traced from McCartney Creek on the south (Stop 5) to Soap Gulch on the north (Fig. 6). It was discussed briefly by Brumbaugh (1977) and mapped by Smedes (1967). Brandon (1984) also mapped this zone and identified younger over older thrust relationships important to its overall interpretation. It is currently being mapped in detail by J. M. O'Neill, U.S.G.S., as part of a regional mapping project in the Highland Mountains. Our own work in the zone has focused on the development of fabric elements in the Lodgepole and Mission Canyon Limestones.

The foreland structure with which the SHFZ interacts is a northwest-trending anticline (herein called the Rochester Anticline) cored by Archean basement rocks. Like the other northwest-trending basement-cored anticlines in southwestern Montana it is related to a northwest-trending fault (South Rochester Fault) (Fig. 6), which had a long period of recurrent activity (Garihan and others, 1982; O'Neill and others, in review). The anticline is largely obscurred by late Cenozoic movement on the South Rochester Fault and subsequent erosion of the Phanerozoic rocks. However, between Camp Creek and Soap Gulch, the Rochester Anticline is well exposed and has a mean axial orientation of 21°N46W. The fold is asymmetrical and verges southwest. The Paleozoic rocks in the SHFZ are located on the southwest flank of this structure. Folds and minor thrusts in the zone trend generally N20-30°W and are therefore slightly oblique to the trend of the Rochester Anticline. The thrusts have only minor displacements. They generally dip between 20 and 45°SW, but in some cases they are nearly flat. The folds associated with the thrusts verge ENE. Transport direction, as inferred from fold hinges, cleavage, and calcite filled extension fractures in the Lodgepole and Mission Canyon Limestones was N65E (Fig. 7).

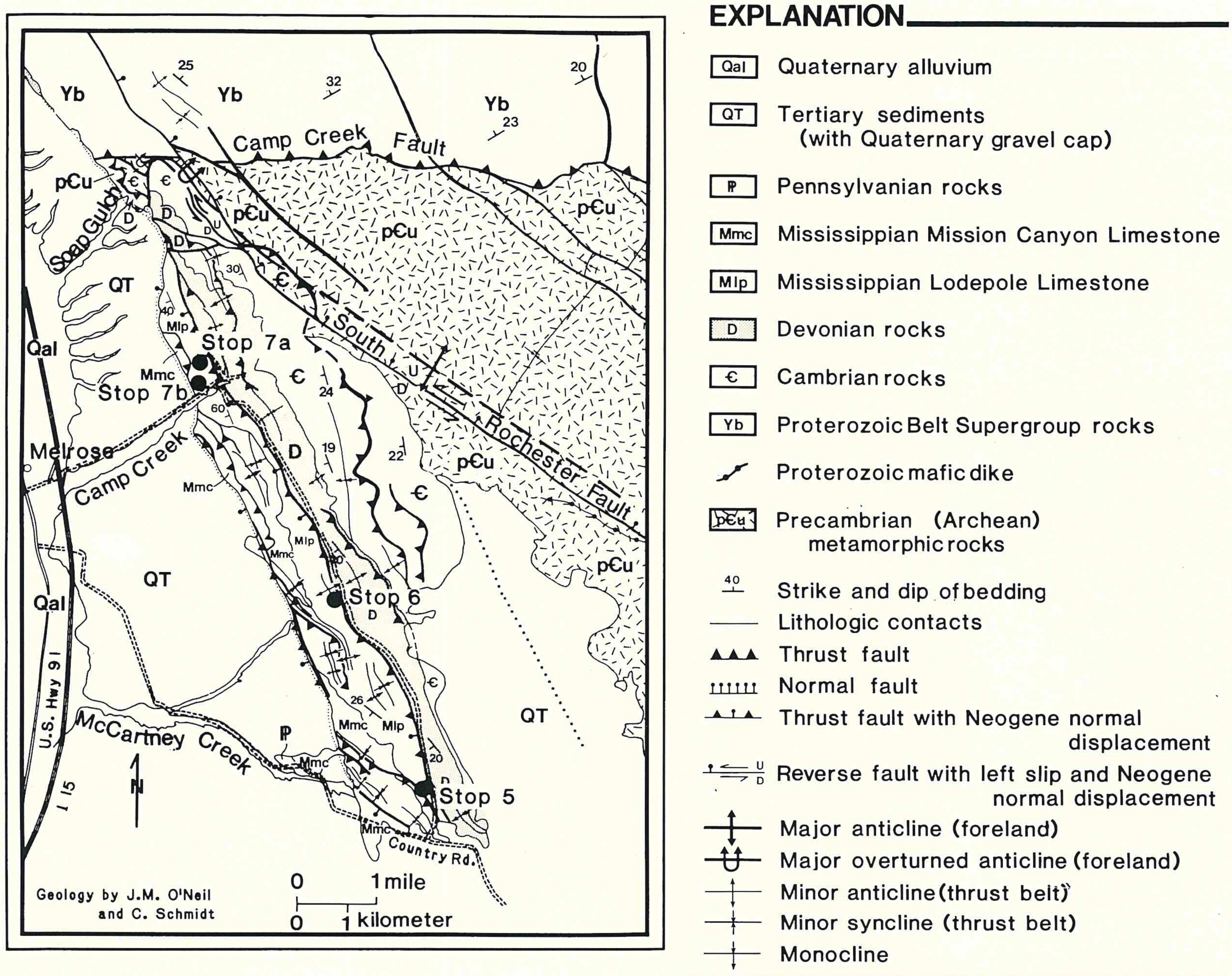

Figure 6. Sketch geologic map of the northern McCartney Salient (Southern Highlands Fault Zone) showing locations of Stops 5-7.

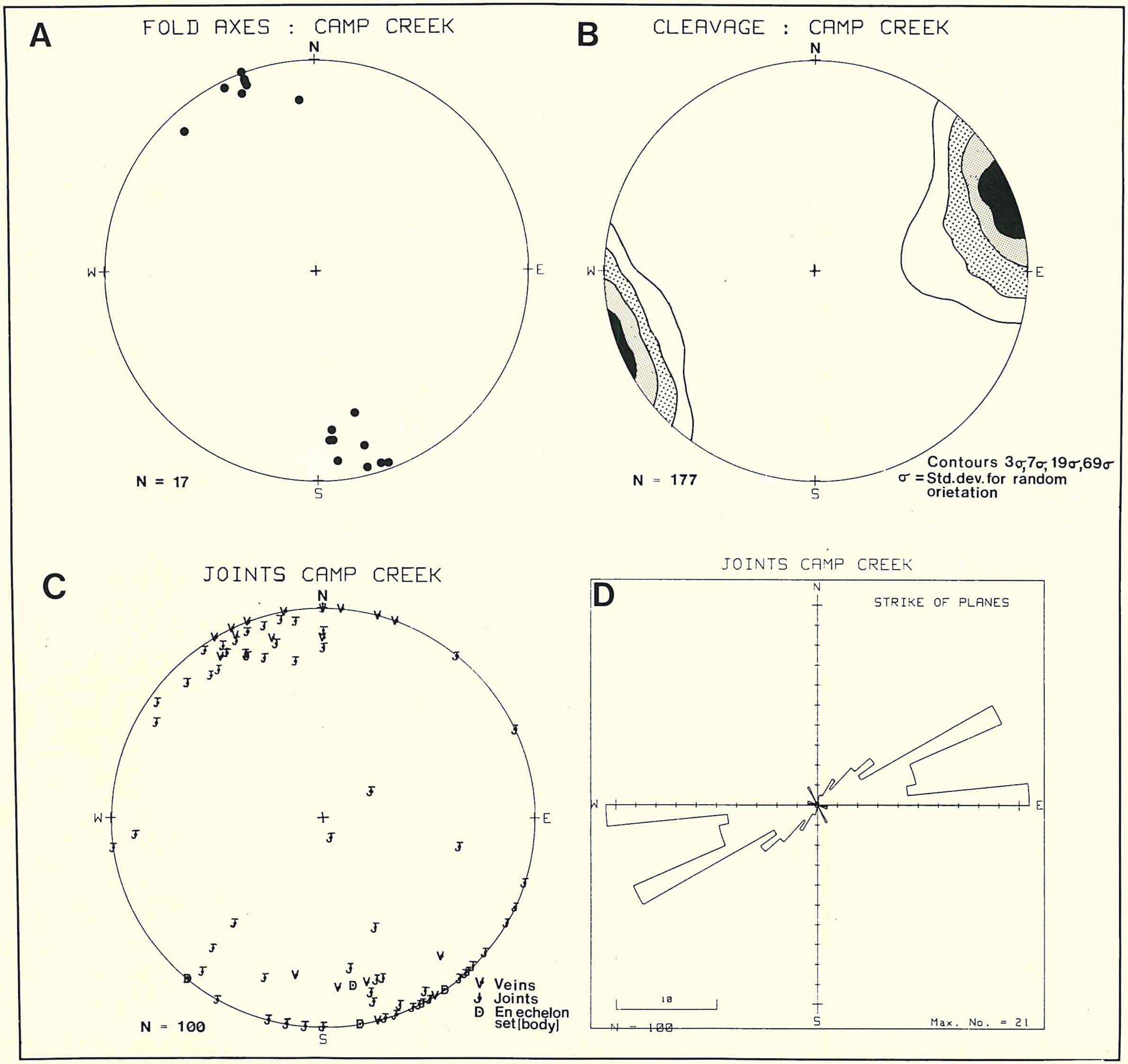

Figure 7. Fabric data for the Southern Highlands Fault Zone expressed lower hemisphere, equal area, plots (A-C) and rose diagram (D).

Several of the minor thrusts place younger rocks on older rocks (Brandon, 1984). They generally follow bedding but locally cut down-section across bedding in the transport direction. The outcrop width of the Devonian Three Forks Shale changes abruptly along strike, a fact we attribute to thrusting at the base of the overlying Lodgepole along an angle slightly more gentle than bedding (and subsequent erosion of the thrust sheet). The field trip route on the unimproved dirt road follows this thrust contact for several miles. The younger over older thrust relationships suggest that, like the Pole Canyon Anticline, the southwest-dipping panel of rocks associated with the Rochester Anticline had obtained its southwest dip prior to thrusting.

Like the cleavage in the Pole Canyon Thrust Sheet, we interpret the cleavage development in the Lodgepole Limestone and several other formations to be due to pre-thrust shortening of the section against a foreland buttress. We have chosen Stops 5, 6, and 7 to attempt to demonstrate this hypothesis and show a progressive northward intensity of cleavage development. Regionally within the McCartney Salient, cleavage intensifies toward the Highland Mountains. In the central part of the salient at the Pioneer Batholith, the Lodgepole Limestone is highly deformed by folds and thrusts, but cleavage is absent. Northeastward toward the Highlands, cleavage appears in the Colorado Formation and in limestones in the Kootenai Formation (both Cretaceous). In the SHFZ it is present in both the Lodgepole and Mission Canyon Limestones. A northward intensification of cleavage can be seen within the SHFZ itself. At the south end of the zone (Stop 5) the Lodgepole displays no penetrative fabric. Cleavage is widely spaced, generally sutured and found only locally within the micritic units. Between Stop 5 and Stop 6 cleavage is ubiquitous in the micrities, is closely spaced and is locally present in sparry units. From Stop 6 northward, cleavage is present in most of the Lodgepole lithologies and in the Mission Canyon Limestone, and it is locally penetrative. Folding and thrust displacement also increase northward in the SHFZ.

The northward intensification of deformation within the SHFZ appears to be related to the degree of shortening of the section against the Rochester Anticline. Because the shortening direction (N65°E) is oblique to the trend of the anticlinal buttress (N46W), the northern portion of the SHFZ is considerably more deformed than the southern portion.

At Stop 5, only the dark micritic units of the Lodgepole Limestone contain cleavage. Where cleavage is best developed, it consists of undeformed microlithons bounded by clay seams (insoluble residue) about 1 cm (0.4 in) apart. The cleavage domains are generally sutured to undulatory, discontinuous, and anastomosing. In most places, however, the micrite units show very weak cleavages or widely spaced tectonic stylolites (sutured domains). Where present, the cleavage is oriented N20-30W and is generally vertical (Fig. 8a).

There is an impressive array of calcite-filled veins, extension fractures and en echelon tension gashes (veins) here. The principal set of veins and extension fractures is oriented about N65E and is therefore normal to cleavage. Another set is oriented nearly east-west and, where

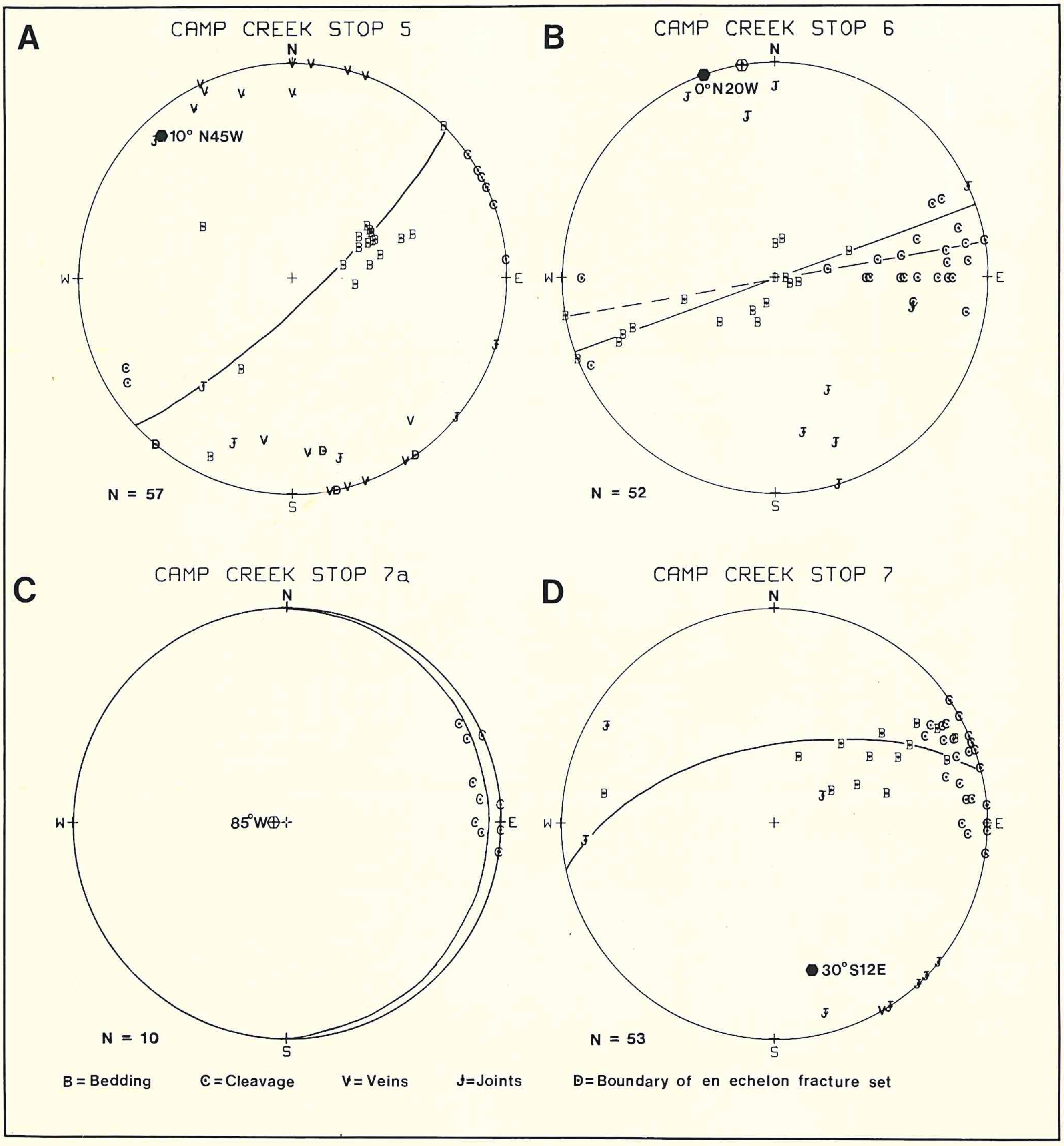

Figure 8. Fabric data for Stops 5-7 expressed as lower hemisphere, equal area, plots.

developed, is perpendicular to north-south oriented cleavages. These veins and fractures are generally transected by cleavage. The en echelon vein sets have several orientations but the most common trend of the zones is N70-80E. The pole to the best-fit plane to bedding is oriented 10°N45W. This is the general orientation of several folds in the Lodgepole Limestone in the vicinity of Stop 5. Folding (as will be demonstrated at Stop 6) post-dates cleavage formation.

Preliminary interpretation of the fabric data here indicates that the initial shortening direction was approximately west-east, parallel to the east-trending vein set. Later shortening in a N65E direction was recorded in the development of the principal cleavage and extension fracture orientations. The latest direction of shortening was N45E as recorded by the orientation of fold axes.

Return to vehicles and proceed to Stop 6.

92.9 (148.6) 3.2 (5.1)

Stop 6. Stop along the road a short distance north of the drainage divide and walk to the nearest outcrops of Lodgepole Limestone west of the road. The purpose of this stop is to examine the fabric elements in folded Lodgepole, especially cleavage fanning, and compare the intensity of cleavage development with that at Stop 5.

Discussion of Stop 6

This fold is one of several is the SHFZ which shows well-developed fanning of cleavage. The fold axis determined from poles to bedding is oriented 0° N17W (Fig. 8b). The fold is clearly asymmetrical and verges east. A zone of mineralization follows the faulted vertical limb, and we interpret the fault to be a small thrust near the base of the Lodgepole. Cleavage orientation varies from vertical to 20°W and therefore shows a rotation consistent with the rotation of bedding on the steep limb. The axis of rotation for cleavage (0° N10W) is only 7° different from that for bedding. A similar rotation of cleavage is shown throughout the SHFZ and is a clear indication that cleavage formed before folding and thrusting. Most of the fold axes measured or constructed for the SHFZ come from the central region near Stop 6. These folds have a fairly consistent N10-20°W orientation which differs from fold orientation near Stop 5 and indicates a shortening direction not significantly different from that recorded by maximum concentrations of cleavage poles and extension veins in this region (i.e., N70W). The east-trending joint and vein set seen at Stop 5 is also present here.

Cleavage exists in a variety of lithologies here. In the micrites it is closely spaced, undulatory, continuous, and wavy to anastomosing. In the more sparry units cleavage morphology is highly variable but is generally moderately spaced (1-3 cm), undulatory, discontinuous, and planar to wavy. A short distance to the west down the main creek bed from this location cleavage shows a pronounced refraction as it crosses sparry and micritic limestone beds, intersects bedding at an angle of about 50° in the sparry units, and is sharply bent to an angle of 20° to bedding in the micritic units. In the micritic units the cleavage is narrowly spaced

(< 0.5 cm), and a second moderately spaced cleavage, perpendicular to bedding, has been superimposed on the earlier fabric.

Return to vehicles and proceed north to Stop 7. This portion of the road can be very rough and impassable if washed out. If this is the case, we will retrace our path to the turn off to Stop 5, take a right and proceed to Melrose. There we can take the Camp Creek road to Stop 7. Following mileages assume road is passable in vehicles at hand.

95.7 (153.1) 2.8 (4.5)

Lodgepole and Three Forks Shale are very hydrothermally altered in this area. Small outcrops of basalt (of presumed late Cenozoic age) occur from here northward and the rocks, particularly the Three Forks, are highly affected locally.

96.1 (153.8) 0.4 (0.6)

Ford across Camp Creek

96.3 (154.1) 0.2 (0.3)

Intersection with Camp Creek Road. Stop 7 has two parts. The first is a short distance to the north (Stop 7a), the second (Stop 7b) is along the Camp Creek Road to the left. Turn right, then left and continue a short distance north.

96.6 (154.6) 0.3 (0.5)

Stop 7a. Outcrops of Lodgepole Limestone over the small hill on the left (west). The purpose of this stop will be to examine post-shortening effects on cleavage morphology and examine a very fine outcrop showing well-developed cleavage/bedding intersections.

Discussion of Stop 7a

Cleavage in this outcrop is narrowly spaced (< 0.5 cm or < 0.2 in) continuous and planar to wavy. Locally, bedding does not match across cleavage domains, suggesting considerable solution within domains. The domains are widened and modified by solution associated with a post-shortening hydrothermal event, and many of them have become narrow calcite veins with numerous void spaces produced by local dissolution.

A sample collected from the less altered outcrops in the creek bed shows two distinct vein orientations and a cleavage associated with each vein orientation. The earlier cleavage strikes N20E, is closely to moderately spaced and sutured (stylolitic). It is nearly normal to a vein set oriented S65E. The later cleavage is closely spaced, nonsutured and planar. It is nearly normal to a wide (1.5 cm or 0.6 in) vein and continues across the vein. The strike of this vein and associated cleavage is N60E and N20W, respectively. Development of the later vein set appears to have been associated with rotation of the earlier cleavage and veins. Therefore, like the fabric elements at Stop 5, there appears to be more than one direction of shortening, and the later shortening direction was rotated counterclockwise toward the foreland buttress, relative to the earlier direction.

Cleavages in this region also appear to be gently folded about a nearly vertical axis (Fig. 8c), but we have no explanation for this folding at present.

Return to vehicles and proceed back to the Camp Creek Road.

96.9 (155.0) 0.3 (0.5)

Turn right toward Melrose

97.2 (155.5) 0.3 (0.5)

Stop 7b. Outcrops of Lodgepole Limestone on the right (north) side of the road. The purpose of this stop is to examine the cleavage in coarse grained crinoidal limestone.

Discussion of Stop 7b

Cleavage here has the same variety as that seen at Stop 6 but is generally more intense. In the crinoidal limestone, cleavages are generally narrowly spaced to penetrative. Cleavage domains are very fine and do not occur as distinct seams. A few crinoids are deformed homogeneously with the rock, but generally they resisted shortening somewhat, as indicated by cleavage domains which bend around them. Some crinoids and other fossil fragments have well-developed coarse calcite pressure shadows.

The combined data from Stop 7a and 7b (Fig. 8d) are consistent with a fold axis for folded bedding oriented 30° S12E and supports our earlier observation that cleavage has undergone late stage folding on a nearly vertical axis.

Return to vehicles. Continue toward Melrose. Outcrops of limestone end at a Tertiary normal fault (Fig. 6).

99.1 (158.6) 1.9 (3.0)

Turn right onto I-15 toward Butte. End of roadlog. The last outcrops of the SHFZ are on the right (east) at Soap Gulch. Normal faults and associated reverse drag in the Tertiary deposits of the Melrose-Divide Valley here suggest that Tertiary normal faulting was closely associated with the earlier development of thrusts in the SHFZ (O'Neill and others, in review).

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FIELD GUIDE TO THE QUATERNARY GEOLOGY AND BIOGEOGRAPHY OF THE EAST FLANK OF THE CENTRAL BRIDGER RANGE, GALLATIN COUNTY, MONTANA

by

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This field guide/log presents both previously published work and unpublished work of the authors and others concerning the effects of lithology, structure, topography, and climate on the Quaternary geology and biogeography of the Bridger Range. The conclusions presented are tentative, and input from trip participants is strongly encouraged.

<u>Miles</u> Increment (Total)	<u>Description</u>
0.0 (0.0)	Assemble in the parking lot of The Inn of Bozeman, N. 7th Ave., Bozeman. The trip will leave promptly at 8 AM and return at approximately 5 PM. Transportation will be supplied. Leave parking lot and proceed left (north) on N. 7th.
0.4 (0.4)	Cross I-90 and take first right (east) on Griffin Drive. Proceed to the first major intersection.
1.8 (2.2)	Turn left (NE) on Bridger Drive. We are now traveling along the broad floodplain of Bridger Creek, which is incised into the post-Laramide valley fill of the Gallatin Valley. The Cenozoic section exceeds 300 m (1000 ft) in exposed thickness, and is tilted NE at 15 into the West Bridger Normal Fault, of which it forms the hanging wall. This fault is probably listric, and most likely joins and reactivates the Laramide thrust complex at depth (J. Zim, 1985, pers. comm.). Folded sedimentary and basement rocks of the Bridger Range are prominent on the NE skyline. The sedimentary rocks, nearly vertical in this area, dominate the geomorphology. The Mississippian (M) Mission Canyon Limestone (of the Madison Group) is the ridgeformer along the Bridger crest. The bedrock and geomorphology both affect the vegetative distribution on this flank of the range. In general, the vegetation of the Bridger Range is similar to that found throughout the central and northern Rocky Mountains. Overall, Douglas fir is considered to be the dominant cover of the Bridger Range. Its dominance, however, is most obvious where no snow movement exists. Changes in species dominance with substrate and altitude are significant (Weaver and Perry, 1978). Grasses are more dominant at lower elevations, whereas subalpine fir predominates at high elevations, especially in areas of snow movement. Changes in species dominance with aspect are significant below <u>ca</u> 2300 m (7500 ft), with forested north-facing slopes and non-forested south-facing slopes. Above that elevation, forest cover is more continuous on all slopes. Bare ground both increases with altitude and changes with aspect in the range, with 50% of bare ground occurring on ESE slopes and 25% on SSW

aspects (Weaver and Perry, 1978). Physiological or climatic timberline is not found in the Bridger Range.

Vegetational development is a product of a combination of environmental factors (including substrate, topography, microclimate, biotic stress, fire, avalanches, and human disturbance) rather than a product of one dominating factor.

1.9 (4.1) **STOP 1 (rolling).** Upon entering the sharp "V"-shaped canyon of Bridger Creek, the route crosses the West Bridger Normal Fault (obscured) and passes onto overturned, recumbent Paleozoic rocks. At road level on the left (north), highly sheared, thin-bedded Lodgepole Limestone is prominent. Several other late Paleozoic and early Mesozoic units on the right (south) side of the canyon are involved in a prominent landslide which was activated in 1961-2 by undercutting of the slope to build the new road at its base. An old quarry had been present at the toe of the slide block. Through a series of uphill retrogressions (both slump and glide), the slide has consumed half of the canyon wall. Failure occurs both along bedding planes and along shear planes generated during deformation, thus most of the slide mass consists of boulders and finer particles, rather than the massive slabs which might be expected, given the massive limestone at the top of the scarp.

New highway plans call for raising the road level, thus trapping future slide debris south of the road in the resulting moat. Nevertheless, the continuing instability here threatens to block both the highway and Bridger Creek, with potentially disastrous consequences.

Leaving Stop 1, the route passes stratigraphically upwards through the steeply dipping to overturned rocks of the east flank of the Bridger Range. Seismic evidence suggests that the east-directed Laramide compressional stresses caused translation of the Bridger Range along several overthrust faults. Although imbricated and overturned slices of this system come to the surface just east of Stop 1, the deeper and more horizontal faults remain at depth, continuing eastward for several miles. The surface structure here is a monocline, sloping with decreasing dip into the Crazy Mountains Basin to the east. The crest of the range is defined by the most resistant unit in the section (the M Mission Canyon Limestone), whereas the rounded topography in the valley is formed on the volcanogenic fine clastics of the Cretaceous - Tertiary (K-T) Livingston Group. This topographic relationship is typical of many of the ranges in this portion of northern Rockies. We will remain in Cretaceous and Paleocene rocks for most of the route to Fairy Lake.

North of the highway near the east range front, a colluvial surface mantle, consisting of angular boulders derived from the resistant units high on the flank of the range, covers pediment remnants. These late Quaternary boulder covers serve as an armor, preserving the evidence of a formerly-higher valley floor.

Along the east side of the Bridger Range, the low elevation, self-perpetuating tree taxon is Douglas fir. Lodgepole pine is serial to the Douglas fir, and is economically more valuable, thus logging practice is to clear-cut areas of 2 to 10 acres, allowing Lodgepole pine to establish a virtually monotype regrowth (Bradley and others, 1970). Recently clear-cut areas and those in various stages of reestablishment of forest growth can be seen along the Bridger road.

2.2
(6.3) South of the highway, the first bench is the site of the proposed Sohio Moats No. 1 petroleum well site. This controversial project is a rank wildcat, but may define a small oil or gas field developed in the outer reaches of the "Fold and Thrust Belt". Sensitivities run high among residents who value the relatively pristine nature of the valley, although most of the long-time residents are ranchers, who favor such development if it can be done with care and concern for environmental values.

The highway turns left (north) at this point, following the strike of the K Billman Creek Formation and the course of Bridger Creek. The creek is now subsequent to the structure; below, it is resequent. This bend in the stream may represent the "elbow of capture" of a mid-Cenozoic Bridger Creek, in response to the lowering of baselevel in the Gallatin Valley and the exploitation of a zone of structural weakness (Bridger Canyon) by a headward-eroding tributary to the East Gallatin River.

1.5
(7.8) Dark shales of the Billman Creek Formation appear in the road cuts to the left (west) where dense vegetation is absent. The purplish hues result from the weathering of andesite clasts, originally erupted from the Elkhorn Mountains, 50 km (30 mi) to the west, prior to the Laramide compression. To the right (east), across the creek, another prominent pediment remnant may be seen about 30 m (100 ft) above the present stream grade. The valley opening eastward in the distance is developed along units dipping slightly NE, and may be related to one of several former consequent streams originally draining the Bridger uplift. These streams were later captured by south-flowing, subsequent, Bridger Creek.

1.3
(9.1) Jackson Creek Road. The Bangtail Ridge to the east is the western margin of the topographic Crazy Mountains Basin. Sandy units found here, with a W-SW aspect, provide a good habitat for pine, whereas the more clay-bound units are barren. These differences show up as

alternating bands of barren and forest sites along strike of the beds(which here dip gently eastward.) The prominent ridge to the left (west) is held up by resistant units within the Cretaceous Kootenai Formation. The open prairies of the bottomlands of Bridger Creek were developed by clearing in an earlier day and are excellent dry-land hay fields.

2.4 (11.5) Stone Creek Road - popular cross-country ski trail.

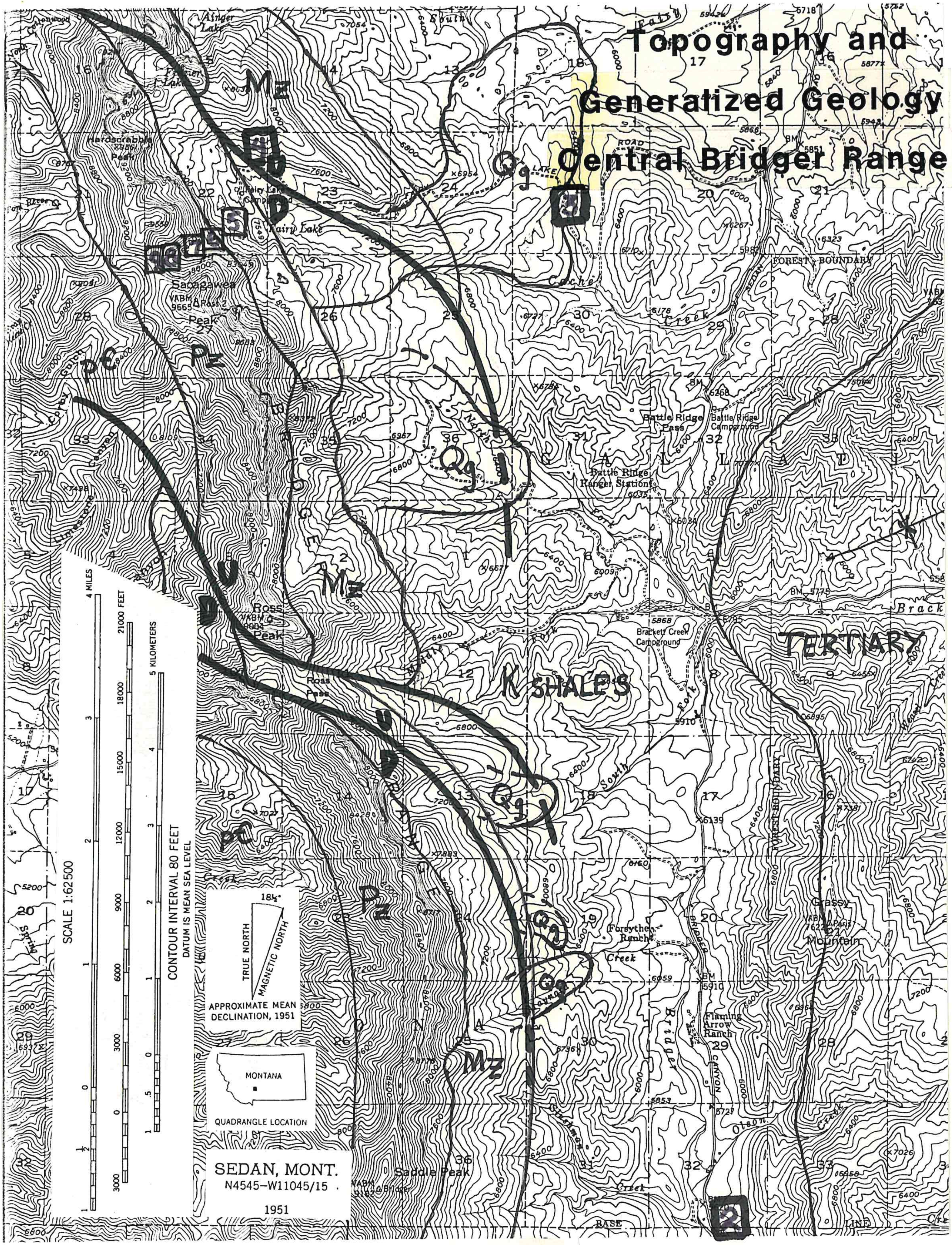
0.3 (11.8) **STOP 2.** Panoramic view of the east flank of the Bridger Range. Incipient glaciers occupied many of the bowls or cirques along the range front here during the Pleistocene. Larger glaciers occurred east of the higher peaks of the range, to the north (Figure 1). Erosional processes have etched out gulches in sharp relief, and vegetative patterns reflect the vigor of these erosional processes. Although the gulches were probably initiated by intermittent stream processes, they have been exploited by mudflows, sheetfloods, and periglacial phenomena.

Snow avalanche activity has further modified the gulches, uprooting trees and piling the debris at their downslope termini. Rarely, a particularly rugged Douglas fir can withstand this attack in the middle of an avalanche track. Subalpine fir, however, although flexible as saplings, never remain in avalanche tracks as mature trees.

Because of the orientation of the range across the prevailing winds, more snow tends to accumulate on this flank than on the west. Avalanches are more prone to occur on the east flank of the range than the west flank because of the excessive snow loading.

Most of the major vegetative communities of the Bridger Range can be seen from this stop. General patterns of distribution can be recognized, exclusive of microscale site effects.

Grasslands dominate at elevations below 1700 m (5500′), and continue to occur at low percentages throughout the elevational range of the mountains (Weaver and Perry, 1978). At high elevations a combination of grasses and herbaceous perennials form open mountain meadows, frequently interspersed amongst forest stands. The meadows are found on both concave and convex slopes. The causes of these open meadows are difficult to pinpoint, an it is likely that a variety of factors are responsible. Grasses occur preferentially on sandstone substrates, perhaps because of a lack of competing vegetation on the dry, well-drained soils. A grass-shrub mixture reaches its highest percentage cover on warm, dry, SE-facing slopes (Weaver and Perry, 1978). Potential problems associated with overgrazing of these meadows by the large population of indigenous mule deer and introduced



Topography and
Generalized Geology
Central Bridger Range
Mz
Pz
PC
Qg
U
D
K SHALES
TERTIARY
Hardscrabble Peak
Fairy Lake
Fairy Lake Campground
Sacagawea Peak
Ross Peak
Ross Pass
Saddle Peak
Battle Ridge Pass
Battle Ridge Campground
Battle Ridge Ranger Station
Brackett Creek Campground
Forsythe Ranch
Flaming Arrow Ranch
Grassy Mountain
FOREST BOUNDARY
SCALE 1:62500
4 MILES
21000 FEET
5 KILOMETERS
CONTOUR INTERVAL 80 FEET
DATUM IS MEAN SEA LEVEL
TRUE NORTH
MAGNETIC NORTH
APPROXIMATE MEAN DECLINATION, 1951
MONTANA
QUADRANGLE LOCATION
SEDAN, MONT.
N4545—W11045/15
1951

FIGURE 1

mountain goats are now being recognized.

Douglas fir dominates between 1700 m and 2600 m (5500 ft - 8500 ft), with smaller percentages of grasses, Lodgepole pine, and Subalpine fir. The greatest cover by Douglas fir is found on SW-facing slopes as well as on limestone (below 2440 m/8000′), gneiss, and interbedded sandstone and shale substrates. The stony, well-drained soils produced on limestone favor deep-rooted trees over the more shallow-rooted grasses. Additionally, it is suspected that the lack of phosphorous in the limestone-derived soils may favor the Douglas fir (Weaver and Perry, 1978). The Douglas fir is not found in sites subject to snow movement. At higher elevations (>2100 m/7000 ft) the Douglas fir is more sparsely distributed, occurring as single individuals or in small clumps.

Subalpine fir dominates within the Bridger Range at elevations greater than 2600 m. This fir is found on N-facing slopes below 2300 m (7500′), and on all slopes at higher elevations, especially those supporting a cool, moist environment (Weaver and Perry, 1978). Associated with this fir in small percentages are Engelmann spruce, Limber pine, and Quaking aspen. The site-specific characteristics of the Engelmann spruce are very similar to those of the Subalpine fir, and the spruce may be slightly dominant in areas of disturbed soil or high radiation receipts. Quaking aspen stands are frequently found in areas of recent disturbance and at sites with a high water table. Limber pine are restricted to well-drained sites, being more drought-tolerant than the spruce or fir. Additionally, the Limber pine are found on windy ridges. The bowl-shaped growth pattern on the pine′s branches are well adapted to high velocity winds. The pine is frequently dwarfed or contorted by strong prevailing winds.

Lodgepole pine is found at all elevations above 1700 m (5500′), being a serial, successional species, reproducing with both serotinous and non-serotinous cones. This pine reaches its greatest distribution on interbedded sandstone and shale substrates. Possibly this distribution is associated with the high nutrient and water availability within the soils developed on this substrate. More probably, however, the successional status of the Lodgepole pine allows its establishment on this substrate following the high number of fires which eliminate the dense stands of other tree species formerly dominant on these rich soils (Weaver and Perry, 1978).

2.2 (14.0) Olson Creek - a popular snowmobile play area.

0.9 (14.9) Flaming Arrow Lodge: excellent view of Bridger Bowl and avalanche paths to be examined this afternoon.

0.5(15.4) Bridger Bowl Road

0.5(15.9) End of pavement

0.9 (16.8) Bridger Creek/Brackett Creek Pass (1870 m/6140 ft). From this point the view to the NW encompasses Ross Pass, the craggy limestone monolith of Ross Peak (2745 m/9006 ft), and Sacajawea Peak (2749 m/9669 ft) to the north. This field trip will examine the deposits to the north and northeast of Sacajawea.

1.6 (18.4) Brackett Creek confluence. Brackett Creek and Road turn right (east), Highway 283 continues left (north). Brackett Creek appears to be anomalous. Bridger Creek occupies a strike valley in the T-K shales between the Eagle Sandstone and the sandy facies of the Tertiary Livingston Group. Cache Creek, a tributary of Flathead Creek to the north, occupies a similar setting. Brackett Creek, however, appears to cross the resistant Tertiary strata at this point. Actually, Brackett Creek is subsequent to the east, lying in a syncline within the Livingston Group, and consequent to the west, where it drains from the Bridger Crest. At this point the nose of the syncline has been breached, thus providing the opportunity to cross the Battle /Bangtail ridge. Quite likely this drainage developed through headward erosion and capture. To the east, Brackett Creek is superimposed on highly resistant sandstone ridges, probably during late Tertiary downcutting.

1.9 (20.3) Battle Ridge Campground and Pass (1930 m/6340 ft).

0.7 (21.0) Turn left (west) onto Fairy Lake Road. Highway 283 continues north towards Wilsall. The Fairy Lake road bears generally west up the valley of Cache Creek, through extensive clearcuts. The upper portion of the Cache Creek valley was apparently glaciated mainly by spillover from the Fairy Lake drainage which originated in a cirque on the SE side of Sacajawea Peak. The drainage from this cirque appears to have been captured recently by Fairy Creek, which is migrating southward.

1.5 (22.5) Road turns north, crossing Cache Creek. (NOTE: We are now entering the area of Figure 2, from the south.)

0.4 (22.9) Fairy Lake road turns sharp left (west), crude road continues north. The north road lies in a strike valley within the Livingston Group, and its lower portion lies between moraines from Fairy Creek and resistant units within the Livingston. Eastward it lies on outwash from the Fairy Creek glaciers.

The Fairy Lake road climbs across shales onto resistent sandstones, still within the Livingston Group, and continues northward. To the left (west) are rolling areas of Cache Creek till (Bull Lake I and II(?) equivalent) with Cache Creek and Fairy Creek (Pinedale I and II(?) equivalent) outwash between them (Figure 2).

SURFICIAL GEOLOGY OF THE FAIRY LAKE AREA, BRIDGER RANGE, MONTANA

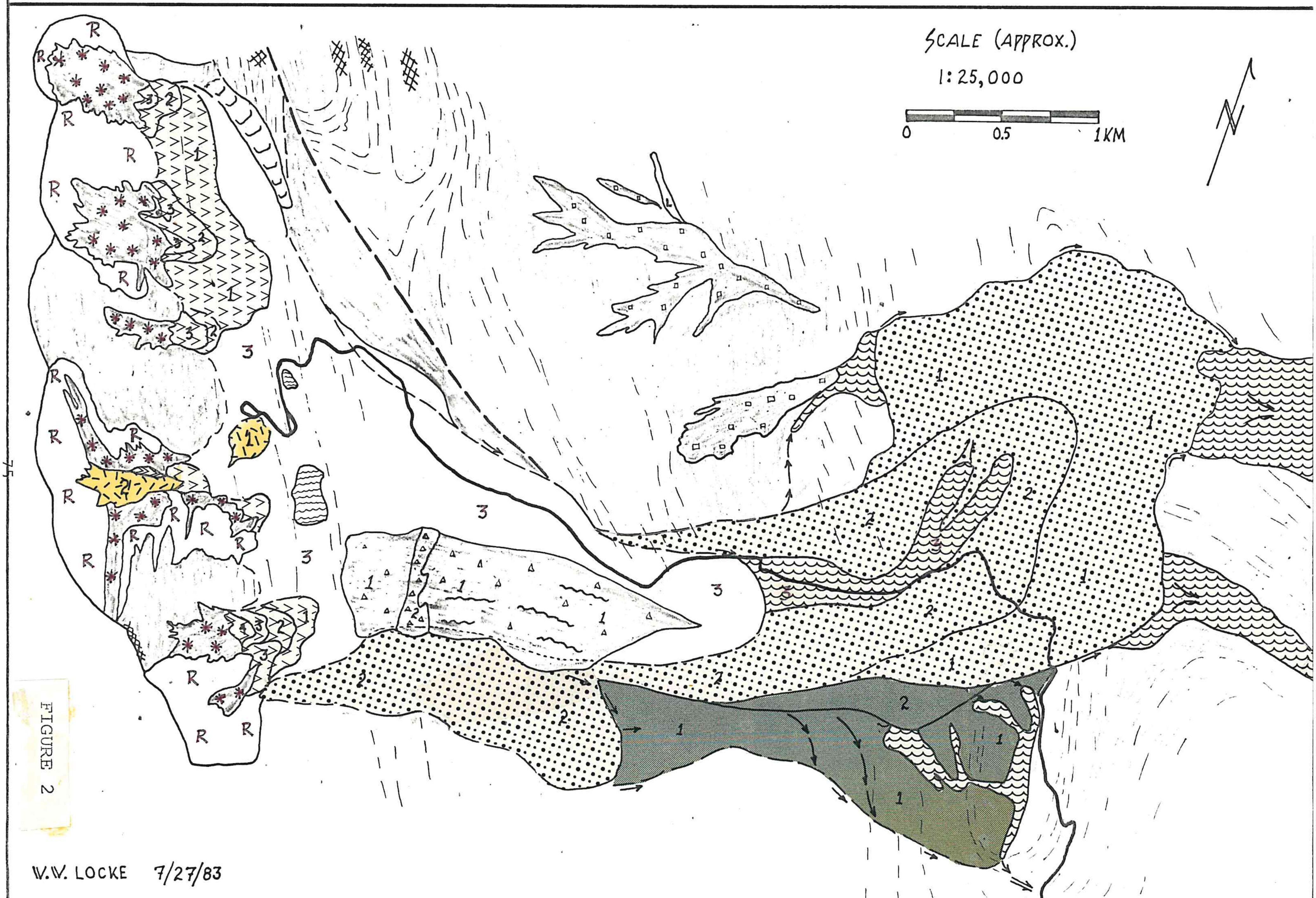
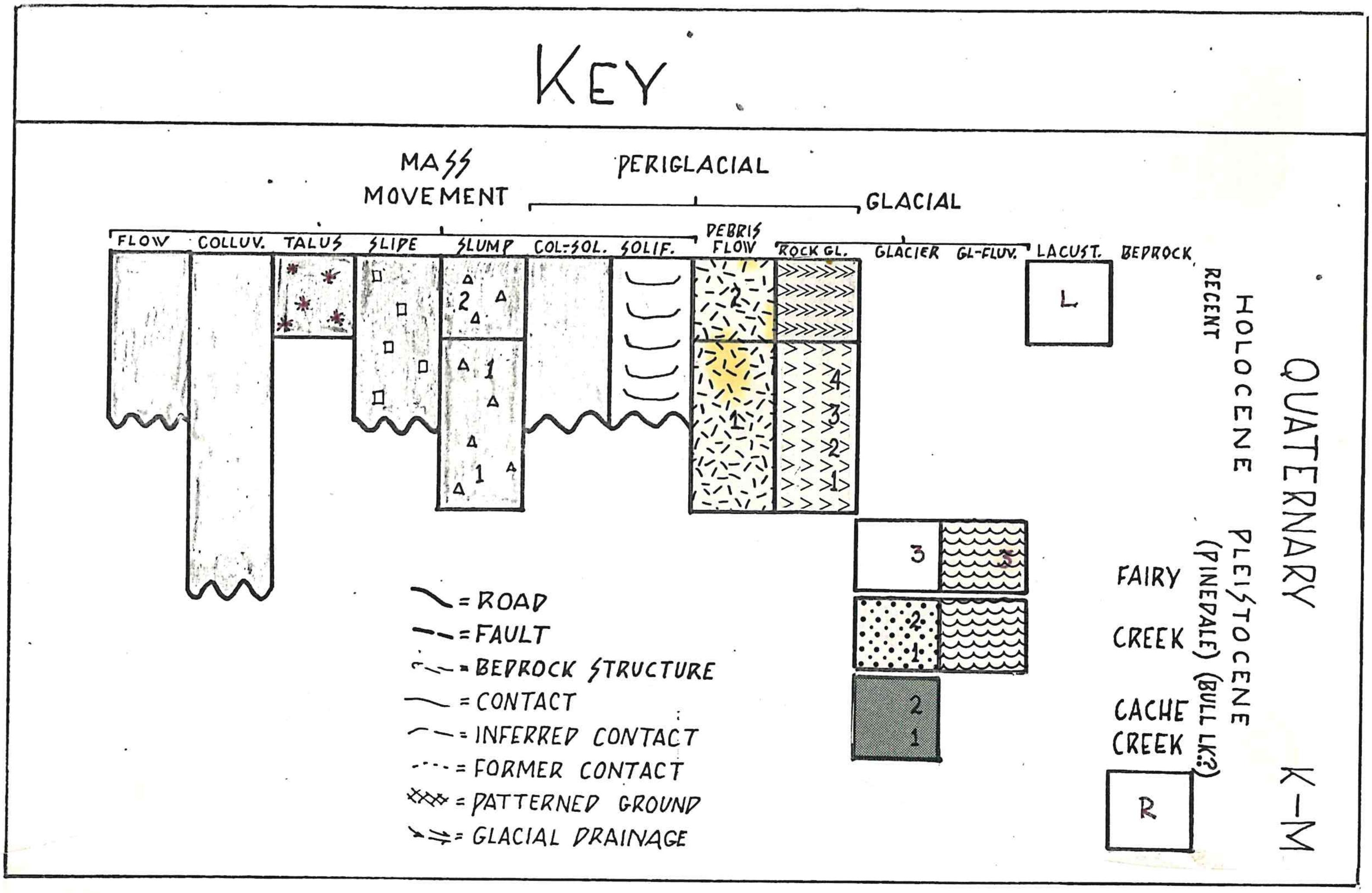


FIGURE 2

W.W. LOCKE 7/27/83



0.9 (23.8) **STOP 3.** At this point we enter glaciated terrain (Fairy Creek I). The most obvious evidence is the scattered but obvious erratics, mostly of Mississippian Mission Canyon Limestone, but also of Pennsylvanian Quadrant Quartzite and Cretaceous Kootenai Sandstone. Schrunk (1976) interpreted this material as Bull Lake-equivalent till, Locke (unpubl) interprets it as early Pinedale in age. The lack of notable field differences in soil morphology between this and proximal tills indicates a probable Pinedale age. If time permits, some soil pits may be open to examine.

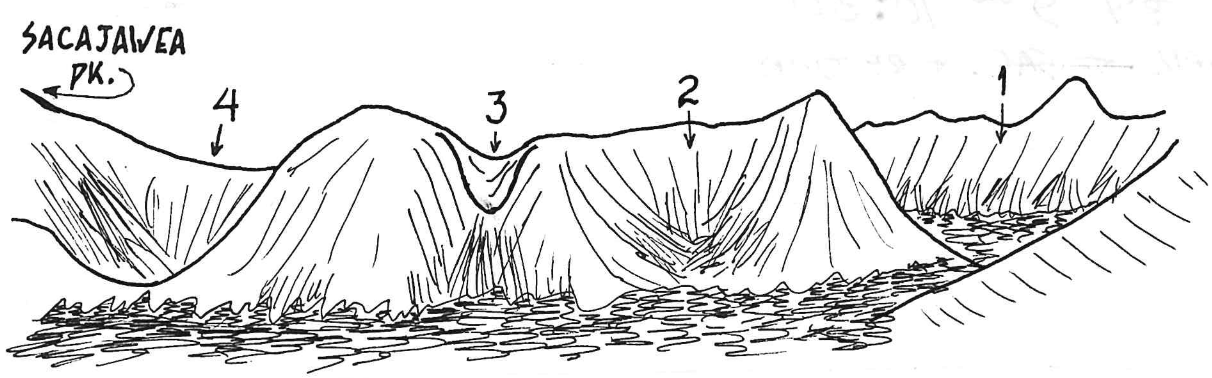
0.3 (24.1) Entering Fairy Creek II till (Pinedale I, Schrunk (1976); Pinedale II, Locke (unpubl.)). Erratics are more numerous on this unit than on more distal units, implying a significant age difference.

0.6 (24.7) Crossing Fairy Creek. We are now on Fairy Creek II recessional outwash. As we climb westward towards the imposing peak of Sacajawea, we cross the hidden contact between the T-K Livingston Group and the (undivided) Cretaceous shales. The climbing "S" curve (on K shales) is the end of 2WD travel in rainy periods. From this point westward there is a strong structural control on topography, on which glaciation has merely overprinted.

0.8 (25.5) The steep slopes on the right are cut into the resistent beds within the Kootenai, Jurassic Morrison, Jurassic Ellis, and Pennsylvanian Quadrant formations. The swales represent the shales within that sequence.

0.4 (25.9) Bearing NW and climbing, we are now traveling along strike within the shales of the Mississippian/Pennsylvanian(?) Amsden and Mississippian Big Snowy formations. As openings allow a view to the left rear (south), the south wall of the valley shows signs of a recent (1960) slump-flow and of a much older and larger (1 km^2) slump-flow. These features are typical of areas underlain by Cretaceous shales along the mountain front to the south, and incorporate till (in the 1960 slump) and till and bedrock (in the older failure). As we continue NW we cross the Cross-Range Fault, back onto the Cretaceous shales. This repetition of section, particularly the nonresistant shales, explains the N-branching nature of the Fairy Creek Valley, the broadening of the upper valley, and the lack of bedrock outcrops within the valley at this point.

1.0 (26.9) At this point the road flattens and bears left (west). To the right (NW) the shales continue as a broad, grassy swale to the crest of the divide between Fairy and Frazier Creeks. If the weather threatens or some people do not want to climb to the cirque floor, the hike to this ridge crest offers some moderately well-developed patterned ground and good views of the surrounding valleys, mountains, and plains.



STOP 4. The view straight ahead (west) shows the northern four cirques which supported the Fairy Creek glaciers. (The southernmost cirque is behind a shoulder

of Sacajawea, to the left.) Cirques 1 & 2 drain Hardscrabble Peak (2914 m/9561 ft) and are cut into Mississippian Lodgepole Limestone (with shaley inter-beds). Cirque 3 has only cut into massive Mississippian Mission Canyon Limestone, and is therefore poorly-developed and hanging. Cirque 4, the Sacajawea Cirque, is our goal. The talus which is evident on the walls of the northern cirques grades into rock glacier deposits. The last phase of deglaciation was probably characterized by copious rockfall as the cirques emptied of ice, with accompanying minor readvance of the shrunken, debris-laden (thus insulated) glaciers. The talus extending downward from Cirque 3 appears to be partly rock-glacierized; a poor example of one of the more prominent periglacial features of the area.

0.7 (27.6) Passing Elf Lake on the left (east). Unlike most glaciated valleys, the lakes in the Fairy Creek drainage are not controlled by glacial process as much as by lithology. Elf Lake and larger Fairy Lake, to the south, lie in basins scoured into the shales of the Jurassic Morrison Formation, between the resistant Kootenai sandstones on the east and the Jurassic Swift Sandstone on the west (exposed in the roadcut on the right side of the road). Only about 2 m (6 ft) of till covers the bedrock at the lake outlet.

0.5 (28.1) Fairy Lake Campground and Parking. Fairy Lake lies ahead and to the left, about 2 minutes down an easy trail. We will plan to have lunch here and to leave this lower parking lot at 1:00 PM for the second half of the trip and return to Bozeman. Please let one of the leaders know if you do not intend to accompany the rest of the group into the cirque.

0.2 (28.3) Sacajawea Peak Trailhead Parking. All ashore! If the vans are not here when you return, don't panic - they will be at the Lake parking lot.

The remainder of this log will cover the foot trail between this point and the cirque floor. The first few hundred meters (yards) of the trail climb gradually through dense forest across irregular terrain. It may not be immediately obvious, but this terrain is the result of major debris flow events issuing from the cirque. The age of the forest cover suggests that it has been some time (>250 yr) since a major event occurred.

The forest-alpine transition within the Bridger Range, and specifically near the Sacajawea Cirque, exhibits many variations resulting from microclimatic, topographic, and geomorphic differences. The conifer species are distributed along environmental gradients, with Limber pine (xeric) found along the dry, wind-swept ridges, and Subalpine fir and Engelmann spruce (mesic) found in sites of snow accumulation and protection from wind. Where spruce and fir are krummholz (dwarfed or distorted by environmental extremes), the Limber pine is often found erect, withstanding winter desiccation. Throughout the ecotone, as environmental conditions become more harsh, the trees take on an appearance or form less like "normal" trees. These changes in form reflect growth differences induced by the increase in severity of climate which accompanies increases in elevation and in exposure (Hansen-Bristow, 1981)

STOP 5. Within the closed forest, below timberline, the coniferous and deciduous trees have a characteristic shape: a vertical trunk with symmetrical branching. Shading of the understory is great, hence little understory vegetation exists in the dark, often damp, closed forest. The trees are relatively closely spaced; canopy cover nears 100%. A great increase in species diversity usually occurs in sunlit areas.

At timberline the trees continue to support branches on both sides of a main, vertical trunk. Occasionally, branches on the upwind side of the tree may be slightly less developed that the others. The conifer cover is less dense, often with a grove or island pattern. The trees are usually shorter and have narrower and more pointed crowns at timberline than within the closed forest. This may be a population response to heavy snowfalls, as narrow crowns hold less snow and reduce the damage of branch and leader breakage. Many of the trees have a wide trunk at the base, a characteristic of age or wind swaying.

As shading is much less within the timberline zone due to the wide spacing of trees, the understory species are more numerous. Additionally, with the abundance of light, the trees have living branches along the entire trunk. Wide, open meadows may also occur at timberline. The species found here are predominantly of forest affiliation, but occasional mixing with alpine tundra species may occur depending on local microenvironments and proximity to the alpine communities (Hansen-Bristow, 1981).

The trail breaks into the open as it begins to switchback up a steep slope. This slope is controlled by the dip of the Mission Canyon Limestone (ca 65 to the NE). As we climb this slope, which opens onto the cirque floor, we get views into a narrow (post-glacial? fluvial?) channel, controlled by jointing in the limestone. We also see the prominent red staining characteristic of the Mission Canyon-Amsden/Big Snowy contact. This is thus largely a stripped structural surface, with little of the resistant limestone removed, except within the cirque valley. To the north, talus dominates the near view with the Fairy/

Frazier divide in the middle distance.

STOP 6. At the mouth of the cirque the significance of periglacial processes in this area becomes obvious. Across the valley, to the south, is a massive accumulation of limestone boulders. This mass is apparently a valley-side rock glacier. The scattered, mature Subalpine fir shows that this feature is not presently active. Hopefully, some lichenometric data will be available to indicate the time since last movement of this feature.

STOP 7. In the floor of the cirque, above the narrow bedrock notch which is the outlet, the cirque floor is planar, dipping eastward at a moderate angle. There is notable irregularity to this planar surface, but vegetation and dissection obscures the pattern. This pattern will become more obvious as we climb higher into the cirque.

The mound to the right (north) of the trail appears first to be a bedrock knob, then a terminal moraine. It is too small to identify with certainty, but it gives a good location to examine microclimatic control on vegetation.

STOP 8. Still farther into the cirque, the front of an active or recently active rock glacier is evident on the right. The angle-of-repose front is quite clear. As we will see, the upper surface is in part vegetated, showing the ability of a rock glacier to act as a conveyor belt, moving outwards from the rock face and fringing (feeding) talus. Dating the stabilization of such a feature by colonization of the upper surface is, therefore, questionable.

Upslope from timberline the trees gradually become deformed due to a variety of factors, predominantly wind, winter desiccation, freezing injury, and mechanical abrasion. The first arboreal vegetation forms seen above timberline are groves of flagged trees. These trees, often found in elongated groves, create a special microclimate within the grove. Herbaceous understory vegetation is diverse, composed of both forest and alpine species. Because snow cover often exists late into the growing season, diversity and lushness of growth may be slightly diminished. Also, the late-lying snow allows the growth of a snow fungus (<u>Herpotrichia</u> <u>nigra</u> Hartig.) which parasitizes the needles on the lower branches. Many of the flag trees, therefore, have no needle growth within the lower 1 to 1.5 m (3 to 5 ft) of the trunk. The flag trees support a main, vertical trunk with branching mainly on the leeward side of the trunk. Slight windward branching is found on the lower trunk and occasionally along its midsections. The precise height of windward branching on these flag trees appears to be controlled by winter snow depth. In all cases, the upper section of the trunk supports no branching on the windward side. The upper flag is normally exposed above the winter snow cover (Hansen-Bristow, 1981).

STOP 9. Continuing on past the rock glacier, we cross the contact between the Mission Canyon Limestone and the Lodgepole formations. We also see the cirque floor in more detail. It appears to be formed through a debris flood/flow process, with a steep, braided morphology and lateral levees. The lateral levees are most clear near the cirque headwall - the farthest point of this trip. The source of most of the debris is clearly the couloir system in the NW corner of the cirque. Here, glacial, fluvial, and frost action and mass movement have eroded into dominantly shaley Devonian rocks.

With an increase in altitude and climatic severity, the form of the trees becomes even more stunted. The krummholz form upslope from the flag trees is termed a flag mat. The trees support only short, often scrawny, vertical trunks. Often the main trunk is bent and misshapen. Branching from the trunk close to the ground is a mat, cushion, or shrub growth. Although branching is found on all sides of the trunk, most branches extend to leeward. The mat varies in height from 1.5 to 2 m (5 to 7 ft) and the length may exceed 6 m (20 ft). Often the flag mat trees grow in elongated clumps, forming a ribbon extending in a wind-lee or west-east direction. The clump or island may actually be composed of several individuals. From the lower mat, frequently an upright branching leader is found. This upright leader, exposed to windy winter weather, may be called a "supranival flag". This leader is flagged, supporting branches only on the lee side. Although able to withstand to present environmental conditions, it appears rather stressed, suggesting that with climatic deterioration such leaders would most likely die back.

The second krummholz form, found at the highest elevations within the ecotone, is the mat or cushion. In this form the tree is dwarfed to a mat, usually no higher than 1.5 m (5 ft) but up to 5 m (16 ft) long. Rarely are vertical branching or upright leaders found. The trunk lies horizontal, close to the ground, and extends to leeward. The plant tissues on the upwind side of the mats are mostly a contorted mass of brown and broken needles and distorted branches. These deformations are mainly the result of mechanical abrasion by wind-borne particles. Little snow cover occurs on the upwind sides of the mats, but survival of the mat is largely dependent upon winter snow cover, as shoots that project above the snow are severely damaged or destroyed. The mat form, as viewed from the side, is asymmetrical or triangular. The windward side has a gentle slope, while the lee side is more nearly vertical. This is a result of shaping by the wind.

The mats are widely spaced and occur in a matrix of herbaceous and dwarf shrub alpine tundra vegetation. Rarely are sexually reproduced seedlings found in this subbelt.

At the uppermost ranges of the spruce and fir, the trees are found only where some type of shelter exists. The shelter provides protection from the dry winter winds which both desiccate and abrade with rock, snow and ice particles. Additionally, such shelters act as snow fences, causing an eddy on the lee side in which snow is deposited. The snow cover provides insulation to the soil and foliage during the winter and a moisture supply during spring and early summer. In cases where spruce and fir presently appear not to have adequate shelter, the windward side of the individual is damaged and the tree is often killed. These individuals either established long ago when climatic conditions were milder or began behind a shelter and migrated by layering to more exposed sites (Hansen-Bristow, 1981).

This is a good point at which to summarize the geomorphic features of this trip to date. Near the range crest, we find three different types of periglacial features dominating: water-lain debris deposits, talus, and rock glaciers. The debris deposits are characteristic of incompetent, fine-grained bedrock (Devonian shales), the talus of competent bedrock yielding blocks of moderate size (Lodgepole Formation), and the rock glaciers of competent bedrock yielding blocks of large size (Mission Canyon Limestone).

Below the cirques, structural control (Kootenai Sandstone) has localized erosion (Fairy and Elf Lakes) and deposition, mimicking moraines. Still farther downvalley, oversteepening of till-mantled slopes cut into shales has lead to massive slope failure. The morphology of the valley itself is controlled by structure (the Cross Range Fault and the regional strike [NW] and dip [65 NW]) and lithology (Cretaceous shales to the south, Paleozoic and Mesozoic rocks of varying resistence to the north).

Finally, there have been several major ice advances in the Fairy Creek valley. These advances have been tentatively correlated with both the Pinedale and Bull Lake Glaciations.

From here we will return to the vans and to the shore of Fairy Lake for lunch. If time, weather, lungs, and legs permit, you can climb to the col above us (about 30 minutes roundtrip) before returning to the lake (about 20 minutes from here). There is a spectacular view from the top, including an active rock glacier in Corby Gulch (to the west) and the Gallatin Valley with its mountain backdrop (the Tobacco Roots are to the west, the Madison and Gallatin Ranges to the SW and S, and the Elkhorns to the NW). To the east are the Crazy Mountains, which harbor small glaciers at present and were extensively glaciated during the Pleistocene. Depending on where you draw your boundary, this was the only area of the Great Plains physiographic province to be extensively glaciated by **local** ice.

WE WILL PLAN TO LEAVE THE FAIRY LAKE PARKING LOT FOR THE SECOND PORTION OF THE FIELD TRIP AT 1:00 PM SHARP!

Return to Bridger Bowl by the same route. As you cross Brackett Creek on the return trip, notice the outwash that Schrunk (1976) hypothesized drained from glaciers to the west. He mapped glaciation of cirques below the entire ridge from Sacajawea to Ross Pass, with moraines extending to within about 2 mi (3 km; west) of this spot.

12.9 (41.2) Turn right (west) onto Bridger Bowl Road.

0.5 (41.7) Parking lot at Lower Chalet, Bridger Bowl Ski Area. Cars will be taken beyond the parking lot to the Upper Chalet, or further if conditions permit. From here the group will hike into the subalpine area. Vegetation patterns and effects will be observed and evidence and results of snow avalanches discussed. The glacial and periglacial features of this area are less obvious than those of the Fairy Lake area but, using Fairy Lake as a model, it may be possible to draw valid analogies. Many questions remain to be solved and discussion by this group may bring forth new ideas and understanding. Rather than refer to specific stops, the following discussion introduces the topics to be covered here: snow and avalanches, ecology of snow movement zones, and subalpine geomorphology and hydrology.

This subalpine area in the headwaters of Maynard Creek is typical of east-facing slopes in the Northern Rocky Mountains. Early October snows begin to set the pattern for snow metamorphism and movement throughout the following winter. By November, snow is about 3 ft (1 m) deep here, and midwinter pack averages about 8 ft (2.5 m). Toward the upper slopes, if avalanching does not completely remove the snowpack, expectable depths are 12 to 14 ft (3.7 to 4.3 m). The easterly aspect of the slopes and the prevailing westerly winds causes snow accumulation in lee-side eddies.

Snowcover can be expected to last well into May on all slopes, and into June on N and NE slopes. Climax runoff, which occurs rapidly because of the short distance to the master stream, comes in early to mid-May. Saturation of the ground and rock debris at this time often leads to extensive modification of the landscape by mudflows and wet slab or loose snow avalanches. In the period of recorded observation, from

1960 to the present, the largest snow avalanche to occur in this part of the Bridger Range ran on May 7, 1963.

Snow avalanches have been recorded as early as mid-August and as late as early June, however, the "avalanche season" is not usually underway until late November. The usual course of events involves collapse slab avalanches, some climaxing to the bare ground, through the end of December. Occasionally a loose powder snow avalanche will run this early in the season, but only rarely. If the ground is exposed, the stage is set for further slides at the same location, since snow cover will remain shallow, thus conducive to temperature gradient metamorphism, instability, and failure. Heavy January and early February snows usually build up toward climax slides in early February. Slides may be expected regularly as each major snowstorm settles. Very little rock debris has been observed in these midwinter slides, but trees are fair game and much modification of the vegetation is brought about. Evidences of direct damage to tree trunks and branches by sliding masses of snow, ice, and rarely rock, are numerous in the obvious avalanche zones. However, frequently avalanches move through a grove of trees without appreciable damage. Indeed, contrary to some opinions, avalanches may even start in tree groves.

Interesting ecological relationships are found within areas of avalanche and snow movement activity. Troughs, gullies, and bowls appear to have been formed and maintained by these and related processes (snowmelt, glaciation). Herbs and grasses occupy these sites; there is no evidence that most of these areas supported a forest cover in the recent past. The lack of a forest cover may relate to geomorphic processes or to other factors, such as substrate, topography, microclimate, biotic stresses, and fire. _t is likely that each park is the result of a differnt factor or combination of factors.

Forests bordering snow movement areas and sites of snow deposition caused by avalanching show both ecological and structural relationships to the snow activity. In sites where large boulders (usually Madison Group limestones) are deposited by avalanches, tree seedlings have successfully established on the downhill, protected side of the boulders (Eversman, 1968). The tree establishment halts avalanche-carried debris, promoting further deposition, this providing a more protected environment for more seedlings. The subalpine fir, in particular, is found occupying such boulder accumulations.

Subalpine fir and the occasional Limber pine are also found on ridges bordering avalanche tracks, and may be successful in colonizing mounds of soil, deposited by flow but elevated above the normal path of snow movement. They commonly bear scars of avalanche activity.

Common evidence of avalanching includes trunk curvature within 1.5 m (5 ft) of the ground (to which soil creep may also contribute) and dense horizontal branch growth within 1.5 m (5 ft) of the ground, with sparse to no branching between 1.5 and 3 m (5 to 10 ft) and short, battered branches above 3 m (10 ft) (Eversman, 1968). The nearly bare midsection of the trees is due both to avalanche breakage and to mechanical abrasion by wind-blown ice crystals. The lack of a densely-branched midsection allows adequate summer sunlight to reach the lower branches, promoting denser-than-normal growth. The maximum height of the Subalpine fir and Limber pine in these sites rarely exceeds 5 m (15 ft), and the average diameter is 10 cm (4 in) By comparison, the trees on nearby ridges, which are not affected by snow movement, have sparse to no branches in the lower trunk region, normal branching in the mid- and upper portions of the tree, and reach heights of 14 m (46 ft) with diameters of only 14 cm (5 in) (Eversman, 1968).

There can be little doubt that the processes of snow accumulation so active today were accentuated to the point of ice accumulation during the earlier Quaternary. Below the cirque floors, arcuate forms composed of a till-like diamicton strongly suggest that glaciation extended as far eastward as the base of the Bridger chairlift. Although significant, this is far less extensive than glaciation in the Fairy Lake region to the north.

The nature and source of the rubble in the cirques such as the North Bowl is complex. Some rubble is avalanched down in both midwinter and spring snow avalanches. The arcuate form of incipient moraine is also present, as are pro-talus ramparts. Perhaps all of these processes contribute(d) to the present form of these deposits.

Possibly because of the powdery nature of midwinter snows at this location, and the fact that many if not most avalanches do not run clear to the ground, the boulder tails, cones, and other debris features aptly described by Rapp (1960) in Lapland and Potter (1972) in the Absaroka Range are not dramatically displayed here. This is despite the observed frequency of major avalanches at Bridger during the 25 years of monitoring.

As observed by McMannis (1955), the bedrock geology from the cirque floors to Bridger Creek is dominated by mostly non-resistant, highly clay-bearing rocks of the Cretaceous and Paleocene systems. Many of the hummocky landforms in this zone may be caused by ancient to modern mass movement of this material and its regolith. Such mass movements can easily be mistaken for morainal forms.

The final facet of subalpine geomorphology to be discussed is that of hydrology: the catchment and storage of melt- and rainwater in the porous sieve of Quaternary deposits along the mountain flank. Like other mountain-flank catchments, such as alluvial fans, they serve as valuable reservoirs for streams such as Maynard Creek. Not only is evaporation and runoff retarded, but some purification and filtering is induced.

The springs fed by this groundwater reservoir run dry from higher elevations to lower over the course of fall and winter, suggesting that a systematic lowering of water level is taking place. The time of flow cessation is dependent on the recharge of the previous spring. The same springs tend to recover progressively during the latter part of the spring runoff (May and June). Since the availability of potable water is one of the major controlling factors on the development of Bridger Bowl and the canyon, the necessity for the preservation of this reservoir cannot be emphasized enough. It should not be bulldozed, removed, used as a medium for the disposal of sewage, or tampered with in any way!

This ends the field trip! We will return down the canyon to the Inn by the same route we travelled this morning. Let us know if you have any comments or suggestions, particularly for future work. Thanks for coming!

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N O T E S