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Geology of a part of the southern margin of the Gallatin Valley Southwest Montana

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GEOLOGY OF A PART OF THE SOUTHERN MARGIN OF THE GALLATIN VALLEY, SOUTHWEST MONTANA

by

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A thesis submitted to the Graduate Faculty in partial fulfillment of the requirements for the degree

of

MASTER OF SCIENCE IN APPLIED SCIENCE

With a Major in Geology

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VITA

Martin David Mifflin was born on March 29, 1937, in Olympia, Washington. His parents are John P. Mifflin (deceased) and Eva Louise Mifflin. Mr. Mifflin attended public primary schools in Portland, Oregon; Olympia, Washington; and Silver Springs, Maryland. He graduated from West Valley High School in Millwood, Washington, in 1955. He then entered Washington State College and attended until early in 1957. The spring quarter of 1957 he attended Eastern Washington College of Education at Cheney, Washington. The fall of 1957 he entered the University of Washington and graduated in the spring of 1960 with a B.S. in geology. In the fall of 1960 he entered Montana State College, in pursuit of a master's degree in Applied Science with a major in Geology.

Professional experience other than this study consists of six months as an apprentice geologist for Pan American Petroleum Corporation in western Alaska in 1959, and three months as a geologist for the U.S. Geological Survey in the Lemhi Range in Idaho, in 1962.

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ABSTRACT

A detailed study of an 88-square-mile area along the southern margin of the Gallatin Valley where late Miocene tuffaceous sediments rest with sinuous unconformable contact on Precambrian metamorphic rocks and Paleozoic strata indicates that structural features control the present configuration of the south margin of the valley. However, much of the evidence pertinent to the Cenozoic history of the area is either indirect or so sparse as to be only suggestive. The sinuous contact, which would suggest sediments partly filled the present valley rather than that structure controlled the valley margin, is the result of a partially exhumed pre-late Miocene paleotopography which displays 2,500 feet of relief and eastward paleodrainage subparallel to Laramide structural trends. Total exhumation of the pre-late Miocene paleotopography apparently would yield mountainous topography with 3,500 or more feet of relief. Some of the old topographic high areas have broad summits which may be remnants of a pre-middle Eocene subvolcanic erosion surface stripped of Early Tertiary and/or late Miocene rock. Only locally has the erosion associated with the exhumation of the late Miocene strata modified the paleotopographic surface.

Bedrock geology is similar to that of the Gallatin, Madison, and Beartooth ranges and consists of northwesterly Laramide high angle faults which place Paleozoic rocks against Precambrian rocks. Early Tertiary basic volcanic rocks (Early Basic Breccia equivalents) occur as two isolated remnants with positions relative to equivalent rocks in the Gallatin Range suggestive of at least 2,400 feet of relative structural

depression of the margin of the Gallatin Valley since late early Eocene. Apparently the majority of this movement has occurred along the northeast-trending Gallatin Range Front Fault, heretofore unmentioned in geologic investigations. Late Miocene and/or later fault movements have occurred along some pre-existing north and northeast Precambrian structural trends and northwesterly Laramide structural trends. Post-early Eocene to pre-late Miocene movement along one Laramide fault may have also occurred.

Pardee describes a high,low-relief surface formed on bedrock in the map area as a remnant of a "Late Tertiary Peneplain". The distribution and composition of adjacent Pleistocene (?) pediment gravels indicate the surface is more likely a pre-late early Eocene erosion surface stripped of the last of its volcanic cover during the Pleistocene.

Lithologic distribution and structure near the mouth of the West Gallatin River canyon indicate that the position of the north flowing river is the result of superposition on late Miocene or younger strata over the presently incised bedrock areas. Valley margin structural relationships also suggest that antecedence may have played an important part in the formation of the West Gallatin River canyon.

Flanking the West Gallatin River are terraces and pediments developed on the semi-consolidated late Miocene strata. These features are believed to be Pleistocene in age, and are tentatively assigned to the Buffalo, Bull Lake, and Pinedale stages of glaciation. The increasingly higher gradients of the successively older terraces suggest the possibility of gradual valley margin warping throughout the Pleistocene and perhaps up to the present.

GEOLOGY OF A PART OF THE SOUTHERN MARGIN OF THE GALLATIN VALLEY, SOUTHWEST MONTANA

INTRODUCTION

This paper is the result of detailed geologic mapping of a portion of the southern margin of the Gallatin Valley. It was undertaken as a thesis project in partial fulfillment of the requirements for the Master of Science Degree in Applied Science, with a major in geology at Montana State College. Field work was done during the summer of 1961 and at intervals thereafter to the spring of 1963.

The objective of the study was to map in detail the geology of the area, with particular emphasis on geomorphic and structural features and distribution of Cenozoic deposits, in the hope that a more detailed knowledge of the Cenozoic history of the Gallatin Valley would become apparent. At the time the study was initiated, two general hypotheses for the origin of the intermontane valley had been proposed, but convincing proof for either hypothesis was not apparent from the literature on the subject. One hypothesis advances the idea of erosional valleys subsequently filled with basin sediments; the other advances the hypothesis that the valleys are the product of structural movements after the Tertiary sediments were deposited. Some geologists favor combinations of the two processes. None of the hypotheses have been proved to the complete satisfaction of all.

The area mapped in this study has been the subject of part of other geologic studies, sometimes specialized in their objectives, and has not been given the detailed coverage necessary to discover many of the relationships which might have a bearing on the Cenozoic history. Choice of the area was determined in part by the known exposures of bed-rock geology, abundant features undescribed in previous works, accessibility, and the suggestion of the Department of Earth Sciences staff at Montana State College.

Previous Investigators

The earliest geologic observations in the general region were made by Hayden (1861, 1884). The first extensive geologic mapping was by Peale, (1893, 1896). His study of the Three Forks quadrangle was of a reconnaissance nature but was remarkably accurate in the gross features.

Interest in corundum-bearing rocks in the map area and adjacent areas has prompted a number of publications. Included are Heinrich (1950), Reed (1951), Clabaugh and Armstrong (1951), Clabaugh (1952), and Foster (1962). Detailed studies of Precambrian rocks in areas near the map area have been conducted by Reid (1957, 1963) McThenia (1960), and Kozak (1961). Douglass (1899, 1903, 1909) described vertebrate remains in Tertiary strata of western Montana, some of which were collected nearby. Later significant work in this field has been done by Wood (1933, 1938), Wood and others (1941), Schultz and Falkenbach (1940, 1941, 1949) and Dorr (1956).

Various regional studies which include the map area have been made by Fix (1940), Pardee (1950), Alden (1953), Hackett, and others (1960), Robinson (1961 b), and McMannis (1962).

Unpublished data collected by W. J. McMannis, John de la Montagne, and R. A. Chadwick has been made available to the writer and has aided considerably in many interpretations of relationships between features in the map area and those of surrounding areas.

Acknowledgments

The writer is indebted to many persons for timely aid, numerous suggestions, and constructive criticism. W. J. McMannis, as major thesis advisor, bore much of the burden in this respect. Also helpful were John de la Montagne, R. A. Chadwick, C. C. Bradley, and M. J. Edie. Thanks are due to Edward Lewis, R. L. Konizeski, and C. Lewis Gazin for identifications of vertebrate remains. Also, the writer is indebted to his wife for both patience and manuscript typing and to Sharon Torgerson for manuscript typing. Appreciation is extended to the numerous landholders of the map area who granted unlimited access.

GEOGRAPHY

Location

The map area, occupying approximately 88 square miles, lies on the southwestern margin of the Gallatin Valley, and includes part of the valley proper and some topographically higher and rougher terrain. The Gallatin Valley is in southwest Montana, with the map area lying between the 111° 10' and 111° 25' meridians and the 45° 30' and 45° 40' parallels. The southern part of the map area essentially comprises foothills of the Madison Range, or Gallatin-Madison Range, perhaps a preferable term due to the structural contiguity of the two ranges. The small town of Gallatin Gateway lies within the area, and the town of Bozeman is approximately eight miles east of the northeast corner of the map area.

Topography

Elevation of the land varies from 4,600 feet on the northeast margin near the West Gallatin River, to greater than 7,200 feet on the south margin at the Salesville triangulation marker. The general landform of the area is one of low, rolling plains in the north and gradually increasing elevations southward toward the foothills of the Spanish Breaks, where the terrain becomes mountainous. Along the eastern margin of the area, floodplain terraces produce more or less continuous smooth surfaces. Locally throughout much of the map area, broken terrain occurs where small streams are incised.

Climate

The average precipitation in the area is not known exactly, but according to measurements made in 1952 (Hackett, and others, 1960, p. 18) the precipitation varies from less than 15 inches per year in the northern part of the map area, to greater than 20 inches in the Spanish Breaks on the south margin. Up to 3 inches per month fall in the spring months, and the rest of the year averages slightly less than 1 inch per month. The daily and seasonal fluctuations in temperature range widely, which is typical of most of the higher areas of the Rocky Mountains. The average temperature in the area is probably slightly less than 40° F., based on comparison with the town of Bozeman, which has an average of 42° F. and extreme temperatures range from above 110° F. to below -50° F. Snow depth in most of the map area during the winter months rarely exceeds more than two feet for prolonged periods of time. The best means of access for geologic field work are by four-wheel-drive vehicle or by horse during the summer months.

Land Economics

Most of the map area at present is in agricultural use, either as farm land or pasture. Areas of low relief and good soil cover are utilized in dryland grain production or, with the aid of irrigation, in alfalfa or other hay crops. Most of the high areas and areas of broken terrain are used as grazing land for beef cattle. Except for minor amounts of federal land, most of the northern part of the area is owned by farmers and ranchers, and the southern part by two large livestock companies.

At the present time there is no economic development of mineral deposits. In the past, corundum of industrial quality was mined, but synthetic abrasives have made these deposits uneconomical. Possible economic materials include Flathead sandstone for building stone, some local areas of Precambrian gneiss in which freshness of the rock would permit usage for building stone, extensive deposits of alluvial gravels for road beds and concrete aggregates, and Paleozoic carbonates for Portland cement manufacture.

Vegetation

Vegetation of the map area where undisturbed by man consists of evergreen timbered areas on the higher northern exposures, sagebrush and grass on other high exposures, and various deciduous shrubs and trees along drainages and on well-watered soil. Grasses and occasional juniper shrubs occupy the lower and drier, uncultivated areas. For a more comprehensive description of plant types and distribution, the reader is referred by DeYoung and Smith (1936).

STRATIGRAPHY

Precambrian

About one-half of the map area is occupied by high grade metamorphic rocks very similar to those termed Pony gneiss in adjacent areas. This metamorphic complex consists generally of the following rock types: gneiss, schist, amphibolite, and metaquartzite, listed in order of abundance. Detailed studies in adjacent areas have indicated that the rocks are a product of a number of stages of high grade and at least one stage of retrograde metamorphism. In adjacent areas determinations on radiogenic minerals indicate ages which range from 3.1 billion years to less than 2 billion years (R. R. Reid, personal communication, 1963). Most authorities believe that the rocks were originally a thick sedimentary sequence that was metamorphosed and intruded by igneous magmas, and subsequently underwent additional phases of metamorphism.

In the map area most of the surface exposures of the metamorphic rock are deeply weathered, generally displaying scattered outcrops of more resistant folia and intervening areas covered by varying thicknesses of gruss and soil. Typically, metaquartzites, amphibolites, and massive gneisses form most of the resistant outcrops. Only in sharp and relatively recent stream incisions are exposures continuous.

Compositional layering is generally well developed, with schistosity commonly parallel to compositional layering over wide areas. Locally the foliation reverses itself, forming large isoclinal folds with amplitudes ranging up to miles in length. In the crestal part of these folds schistosity generally crosses the compositional layering. Attitude of compositional layering was mapped in this study, but upon analysis of structures it became apparent that the schistosity may be critical in recognizing faults which displace only metamorphic rocks. The crestal parts of isoclinal folds, unless well enough exposed to be recognized as such, can create the illusion of post-metamorphic deformation if the relationships of schistosity are not known.

The outcrops of two unusual lithologies plus the similarity of lithologic sequences throughout most of the map area suggest that the very thick metamorphic complex may be repeated by displacements along northwesterly trending faults. The two lithologic types, a corundum rich zone and a siliceous magnetite layer, each occur in two localities on opposite sides of northwest-trending faults. The magnetite layer is exposed in sec. 26, T. 2 S., R. 3 E. and sec. 3, T. 3 S., R. 3 E., on opposite sides of Camp Creek and Salesville Faults. If the unusual lithologies are unique, similar thick sequences are exposed in the northern and southern parts of the map area.

The mineral assemblages of the metamorphic rocks support Reid's (1963) interpretation of a number of phases of metamorphism. Minerals characteristic of high to low grade metamorphism are present. In two parts of the map area, the northeastern and southeastern exposures of metamorphic rocks, the rock types differ from those of the rest of the map area in that there is more massive gneiss present and in having a higher percentage of migmatite veins and less constant attitudes. Elsewhere in the map area layering is fairly constant and includes alternating zones of the following rock types: gneiss-schist-amphibolite, gneiss-metaquartzite, and gneiss-metaquartzite-amphibolite. In these sequences the compositional layers range from a few inches to more than 10 feet in thickness, commonly being about 2 to 4 feet thick.

For more detailed lithologic and petrographic descriptions of similar rocks, the reader is referred to the works mentioned under previous investigations.

Paleozoic

Paleozoic sedimentary strata are preserved primarily in down-faulted and small synclines in the southern half of the map area. Formations as young as Mississippian are exposed, but the majority of the exposures consist of Cambrian and Devonian rocks. Although some of the formations are fairly well exposed, no detailed stratigraphic studies were made. A brief description of the rock units follows:

Middle Cambrian: The Flathead Quartzite, resting unconformably on the Precambrian metamorphic complex, consists of three general zones of differing lithologies. The basal unit consists of alternating maroon and white medium- to coarse-grained quartz sandstone and orthoquartzite, in places containing thin granule to pebble conglomeratic lenses and cross bedding. The middle unit consists of red to maroon beds of medium- to coarse-grained, cross-bedded orthoquartzite. The upper unit is glauconitic and somewhat shaly, and grades into the overlying Wolsey shale. Total thickness, as determined by map distribution, appears to vary from area to area, but probably does not exceed 200 feet. In most of the map area the depositional contact with the underlying metamorphic rocks is subparallel to foliation. Locally, due to pre-Middle Cambrian and post-metamorphism structural movements, the contact is quite angular.

The Wolsey Shale is poorly exposed in the map area, with one exception near the mouth of the West Gallatin River Canyon in sec. 32, T. 4 S., R. 4 E. There greenish micaceous shale and siltstone are exposed and contain numerous worm trails. The estimated thickness is as much as 250 feet in places, but accurate formational boundaries are usually impossible to obtain due to dip slope exposures and extensive cover.

Lying directly above and in gradational contact with the Wolsey shale is the Meagher Limestone which typically forms hogbacks and ledges. In the Bridger Range the Meagher has been divided into three units: a lower thin-bedded dense gray limestone with intercalated greenish shale, a middle unit of resistant thin-bedded dark gray fine-grained limestone, and an upper unit of interbedded shale and fine-grained limestone (McMannis, 1955, p. 1393). In the map area only the lower two units are extensively exposed. The upper unit, if developed in the map area, is on dip slope exposures and is commonly covered by rubble from the underlying unit. On Ruby Mountain the ledge-forming unit is approximately 150 feet thick and is abruptly overlain by what is assumed to be the Park Shale interval (almost entirely covered throughout the map area). Therefore, it is possible that upper less resistant unit has been mapped as part of the Park Shale, or is not present in the map area. Total thickness of the Meagher seems to be from 200 to 400 feet, but no complete exposures were seen.

Directly overlying the Meagher, probably gradationally, is the Park Shale. Only a few feet of poor exposure of this formation were observed in a shallow road cut on the southern boundary of sec. 13, T. 3 S., R. 3 E. Here red soil and a minor amount of micaceous red and green shale are present. The total thickness is unknown, but map relationships of a typically covered saddle on the unit suggest a thickness of approximately 200 feet.

Upper Cambrian: Overlying the Park Shale, apparently in conformable contact, is the Pilgrim Limestone. In the map area this formation is dominantly thick-bedded, medium to light gray, medium- to coarsely crystalline limestone, in places glauconitic and slightly sandy. No limestone pebble conglomerates, common in the lower Pilgrim of adjacent areas, were noted. Where fairly well exposed on Ruby Mountain, the formation displays massive ledge-forming outcrops, whereas to the southeast it forms low hogbacks. The thickness is about 200 feet, with the basal part unexposed.

Above the Pilgrim on Ruby Mountain, in apparent parallel contact, is a saddle-forming, nonresistant thin interbedded yellow to buff sandy and dolomitic limestone and red shale interval 60 to 80 feet thick. This is believed to represent the Upper Cambrian Dry Creek Shale Member of the Red Lion Formation.

In gradational contact with the Dry Creek on Ruby Mountain is a thin-bedded, light gray, medium- to coarsely-crystalline limestone. This contains abundant irregular tan chert stringers and nodules. This ledge-forming limestone is about 42 feet thick, and on the basis of lithologic similarity, it is called Red Lion Limestone (Hanson, 1952).

However, near Goose Creek in sec. 20, T. 4 S., R. 4 E., poor exposures of the same interval (between Dry Creek Shale and Jefferson Limestone) indicate that the limestone present at this locality is similar to the Sage Pebble Conglomerate Member of the Snowy Range Formation. Here, sparse exposures of limestone pebble conglomerate are present. These exposures are adjacent to the Salesville Fault and could possibly be fault slivers of the Pilgrim Formation, which also is known to have similar lithologies. However, it is possible that somewhere between the Ruby Mountain and Goose Creek exposures there may be a facies change in the Red Lion - Snowy Range interval.

Devonian: On Ruby Mountain the Jefferson Limestone directly overlies the Upper Cambrian with an abrupt change in lithology. There is little physical evidence of this major hiatus that represents all of Ordovician and Silurian, as well as part of Devonian time. At Goose Creek the contact between the Sage Pebble Conglomerate Member(?) and the Jefferson Limestone is not exposed. The Jefferson is composed of alternating limestone and dolomite with thin to thick beds of gray to grayish brown fine- to coarsely crystalline textures. It weathers with a characteristic medium to dark brown color. The total thickness of the Jefferson at Goose Creek is approximately 420 feet. On Ruby Mountain the Jefferson may be thicker, but the uppermost beds are either stripped away by erosion or are unexposed because of vegetative cover on dip slopes.

At Goose Creek the Three Forks Shale gradationally overlies the Jefferson Limestone. The Sappington Sandstone is also present. The poorly exposed interval between the Jefferson and Lodgepole formations appears to consist of green and red shales, yellow calcareous siltstones, and yellow calcareous sandstones. The paced interval indicates approximately 140 feet of combined thickness for the Three Forks-Sappington unit.

Mississippian: Exposed above and apparently conformable with the Sappington Sandstone are 175 feet of the Lodgepole Formation. The rock exposed is thin-bedded, dark gray; finely crystalline limestone with intercalated shaly partings. Younger Paleozoic rocks, if present, are covered by Miocene and/or Quaternary deposits.

Tertiary: Erosion prior to middle Eocene time removed a large amount of Paleozoic, Mesozoic, and possibly Paleocene strata before the widespread Gallatin Range Volcanic rocks were formed. Erosion after this phase of volcanism and prior to late Miocene deposition removed most of the volcanic cover and additional unknown amounts of older rock.

Eocene: Volcanic rocks of probably middle Eocene age occur in the map area as two erosional remnants of slightly more than one-fourth square mile each. One is poorly exposed one-half mile southeast of the mouth of the Gallatin canyon. There the rock is mainly oxidized and vesicular basic andesite. The dip is probably southeast at a gentle angle on the basis of vesicule orientation and areal distribution, and thickness is probably less than a few hundred feet.

The other remnant, three miles west of Gallatin Gateway, dips approximately 30 degrees to the north on the basis of the physiographic expression and orientation of flow structures (Fig. 1). The exposures are poor, but the following generalized section was obtained on a traverse of the remnant.

Top:	Unconformably overlain by late Miocene tuffaceous sedimentary strata.
300'	Dominantly vesicular medium gray to oxidized finely crystalline andesite and scoria. Vesicule pipes suggest northward flow. Some vesicules filled with calcite and opal.
350'	Mostly covered. Appears to be interlayered gray, slightly porphyritic finely-crystalline andesite and oxidized scoria.
100'	Oxidized scoria, dark gray scoria, and vesicular basalt.
30'	Finely crystalline medium gray andesite, mafic phenocrysts appear to be altered to golden brown biotite, also the plagioclase phenocrysts are altered.
50'	Flow breccia with vesicular basalt fragments up to 6" in diameter.
50'	Oxidized vesicular basalt.
100'	Black to dark gray glassy and porous basalt, some faint flow bands visible. In some zones plagioclase laths appear altered.
Base:	Precambrian metamorphic rock in angular contact.

Total Thickness: About 1,000 feet exposed.

The age of the volcanic sequence is uncertain but is assumed to be Middle Eocene on the basis of the following evidence. Six miles to the south of the map area near Garnet Mountain W. J. McMannis (personal communication, 1962) has found a thin carbonaceous siltstone directly underlying similar volcanic strata. These deposits rest on the subvolcanic erosion surface. Spore identifications from the siltstone indicate a late early Eocene age; therefore, the volcanics must be no older than that. To the south in Yellowstone Park similar and probably correlative volcanics accumulated until "well into Middle Eocene" Dorf (1960). Therefore, it is probable that the bulk of the sequence is middle Eocene in age.

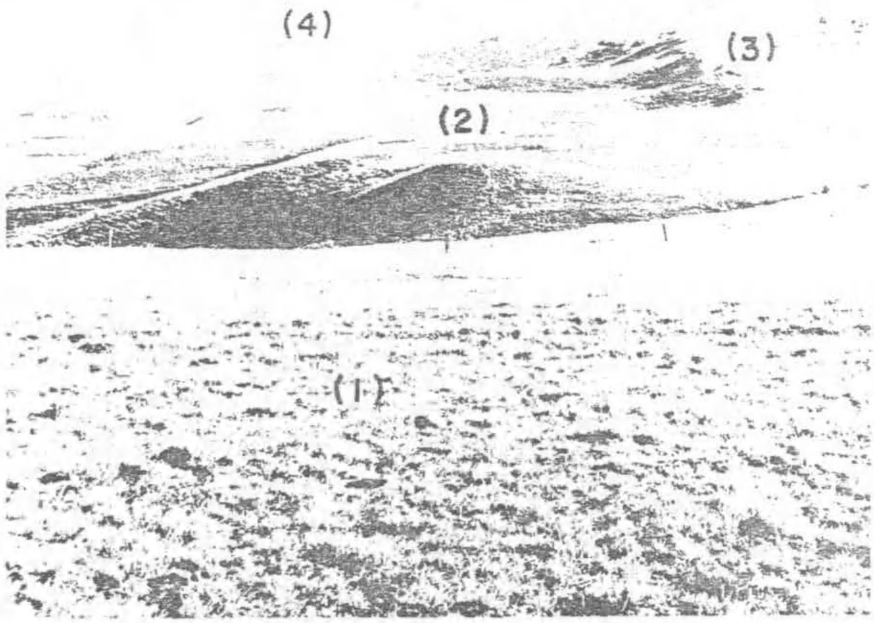


Fig. 1 View to the southeast from sec. 7, T. 3 S., R. 4 E., of: (1) Qpg^2 gravel surface overlying late Miocene strata, (2) remnant of basic volcanic rocks dipping 30° toward the foreground, (3) Precambrian metamorphic rock topographic high, (4) high peaks of the Gallatin range in the background.

Near High Flat in the northern part of the map area in sec. 28, T. 2 S., R. 4 E., a poorly exposed possible remnant of volcanic rock occurs. Angular blocks of andesite are concentrated in one small area and are generally covered by soil or loess. Adjacent to them are rounded lag pediment gravels of the same lithology. However, due to the size and angularity of the blocks, the writer interprets them as inplace remnants. If so, this remnant is probably less than 50 yards square and very thin, resting on and surrounded by metamorphic rocks.

Late Miocene: Sedimentary strata of dominantly light colored, fine-grained tuffaceous nature are widely distributed throughout most of the map area and are believed to be entirely late Miocene in age on the basis of vertebrate fossil remains, apparent similarity of lithologies, and probable contiguity. However, the strata south of the Camp Creek paleotopographic valley and south of the bedrock high three miles west of Gallatin Gateway have not yielded identifiable fossils.

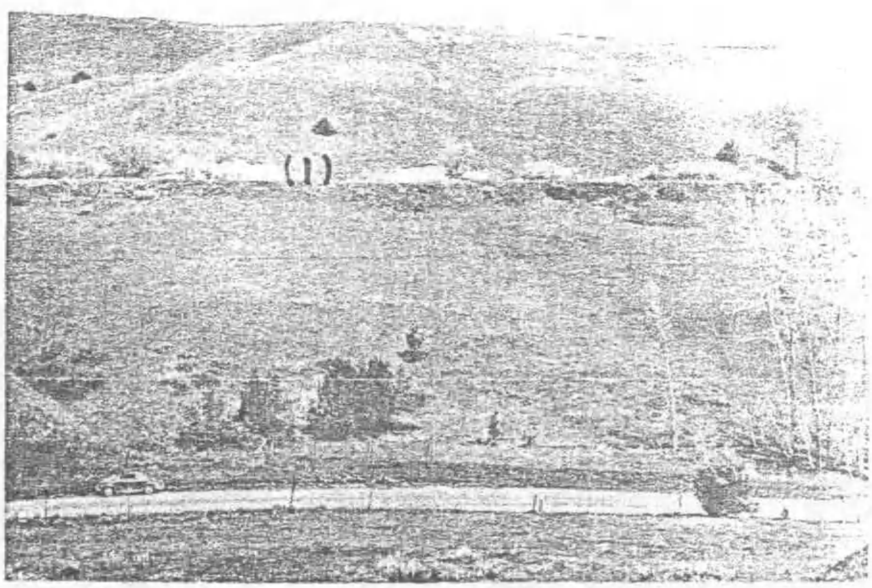


Fig. 2 Typical lack of exposure of late Miocene strata with the exception of ditch cuts (1). Skyline is formed by a Qpgl surface. Qpgl derived gravels occur as float on the entire ravine slope, sec. 33, T. 2 S., R. 4 E.

Throughout the area underlain by the tuffaceous sediments exposures are poor because of the low degree of induration of the strata and extensive surficial deposits of loess and gravel (Fig. 2). Most of the exposures which do occur crop out on steep slopes, and even at these localities only the most resistant beds stand as outcrops (Fig. 3). It is probable that most of the finer-grained strata are covered, but they may constitute the majority of the late Miocene sediments. In general, outcrops consist of less than 50 feet of sequence, and only the resistant coarser grained partially cemented beds are exposed.

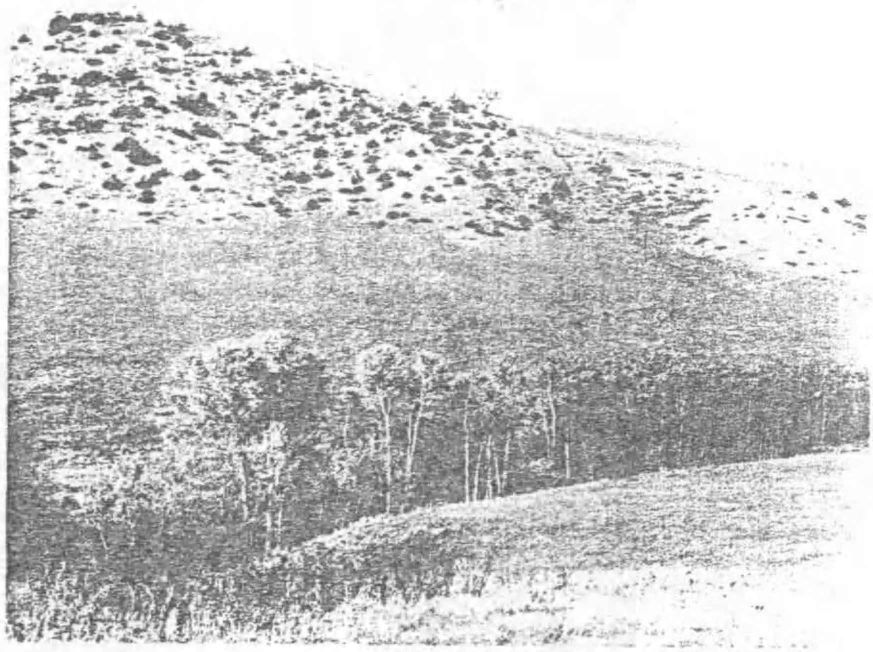


Fig. 3 Northerly view from sec. 29, T. 3 S., R. 4 E., across Brown Hollow. Late Miocene strata dip east at 15-25 degrees. Typical of sporadic exposure of the more indurated beds.

Two representative sections are included here to indicate the two typical sequences noted. The first is less commonly exposed and is probably stratigraphically higher than most of the late Miocene of the map area. This belief is based on fault relationships and character of the beds. The second section is more or less typical of most of the late Miocene exposures that occur adjacent to the flanks of paleotopographic bedrock highs.

Section No. 1, measured along Highline Canal, NW ¼ of sec. 13, T. 2 S., R. 3 E.

Top: covered by soil and loess.

26' +Crudely bedded 1-3' zones of very fine to medium-grained tuffaceous sandstone, with lateral irregular cementation by calcite. Interbedded 8-10" light gray lenses of silt and clay beds, 1' lenses of dark gray calcite cemented siltstone, and a few lenses of yellowish-brown siltstone. Calcareous root casts, Tertiary pebble fragments, and irregular cementation occur throughout the unit.

34' Alternating ½-2½' zones of light gray tuffaceous very fine- to medium-grained massive sandstones, and thin-bedded 1/2 to 3/4" light gray to white, cross-laminated ashy silt and clay beds, in places slightly calcareous. Upper 4' is a massive sandstone well cemented with calcite. Top of one sand bed ripple marked, with overlying white clay filling troughs, but not reaching the crests of the troughs.

10' 6-8" massive beds of gray tuffaceous very fine-grained sandstone, some lightly calcite cemented. A few zones of thin-bedded 1/8 to 1" light gray silty ash beds. Some of the sandstones seem to be rich in carbon.

15' Thin-bedded 6" to ¼" cross-laminated, light brownish gray tuffaceous medium-grained sandstone and siltstone, mostly nonindurated. Carbon content varies laterally.

2' Bed of medium-gray, cross-laminated, medium- to fine-grained, tuffaceous sandstone, slightly calcareous in places.

23' Thin-bedded ½ to 3" cross-laminated and laminated light gray to white very fine-grained sandstone and siltstone, nonindurated, some zones stained with iron oxide.

3' Thin-bedded ¼-1" laminated and cross-laminated micaceous, tuffaceous, buff colored very fine-grained sandstone. Some 1/8-1/4" limonite-cemented layers.

Base: Small fault, probably related to the Middle Highline Canal Fault.

Total thickness: 113'

Section No. 2, along Goose Creek road, in W ½, SE ¼ of sec. 17, T. 3 S., R. 4 E.

Top: Covered by Qpg_1 gravel.

12' Massive 1-3' beds of tuffaceous, buff colored, very fine-grained sandstone and siltstone. Some carbon or manganese spots and veinlets.

2' Bed of granule conglomeratic buff colored fine-grained tuffaceous sandstone. Calcareous, slightly more resistant than adjacent beds. Granules of pumicite.

4' Covered.

22' Partly covered, thin-bedded, very fine-grained pumicite, with pumicite granule to pebble fragments. Color varies from off white to buff. Partly calcareous.

42' Massive 1-5' beds of buff to light gray tuffaceous, fine-grained sandstone, some calcite cemented beds form ledges.

34' Covered, probably nonindurated very fine-grained sandstone and siltstone of altered tuff.

36' Massive 2 to 12' -thick beds of light gray to buff tuffaceous, very fine-grained sandstone and siltstone. Beds vary in amount of calcite cement. One well cemented 6" very coarse sand lense. Top capped by a 6 foot well indurated calcite cemented bed.

6' Bed of coarse sandstone with granule to pebble conglomerate in cross-bedded channels and light gray tuffaceous sandstone. Conglomerate composed of subround to subangular clasts of quartz, basic volcanic rock, metamorphic rock, chert, silicified wood, and some tuffaceous siltstone fragments. Varying amounts of calcite cement, but the upper part is tightly cemented.

36' Mostly covered, appears to be massive buff to white tuffaceous, very fine-grained sandstone and siltstone. Some irregular-shaped calcareous concretions, also carbon or manganese spots and veinlets.

6' Massive buff colored bed of highly calcareous tuffaceous medium-grained sandstone and pebble conglomerate. Pebbles occur as crude lenses and consist mostly of basic volcanic clasts, with some metamorphic clasts.

Base: covered with soil and loess.

Total thickness: 200'

Note: Pumicite refers to sandstone made up dominantly of glassy fragments. Tuffaceous is used in the sense that the sandstone is light colored and punky, but the individual shards are not identifiable due to alteration.

A number of characteristics of the Miocene strata persist throughout the map area. Near the depositional contacts with bedrock, the sandstones typically become coarser, and the bulk of the detrital material is bedrock derived. Many beds are conglomeratic to various degrees, but most typical are beds displaying small to medium-grain size. Pebbles are usually well rounded, of low sphericity, and are dominantly composed of basic volcanic rock, Precambrian lithologies, and in some beds clasts of reworked Tertiary tuffaceous material. The bulk of the volcanic clast-bearing beds occur in two localities: in the west-flanking beds adjacent to High Flat in secs.25 and 36, T. 2 S., R. 3 E., and in sec. 1, T. 3 S., R. 3 E., and in secs. 5 and 8, T. 3 S., R. 4 E., east of the basic volcanic remnant west of Gallatin Gateway. Some pebble conglomerate channels of similar but coarser debris occur to the north of High Flat.

Also widespread in the Camp Creek-Brown Hollow paleotopographic valley is a zone of coarse siliceous arkosic conglomerate and sandstone that forms the only lithologic unit constituting a marker horizon in the Miocene section (Fig. 4). This zone has local thickness up to 30 feet (at the south end of the High Flat bedrock high) and appears to be directly overlain by typical poorly indurated finer-grained tuffaceous beds.

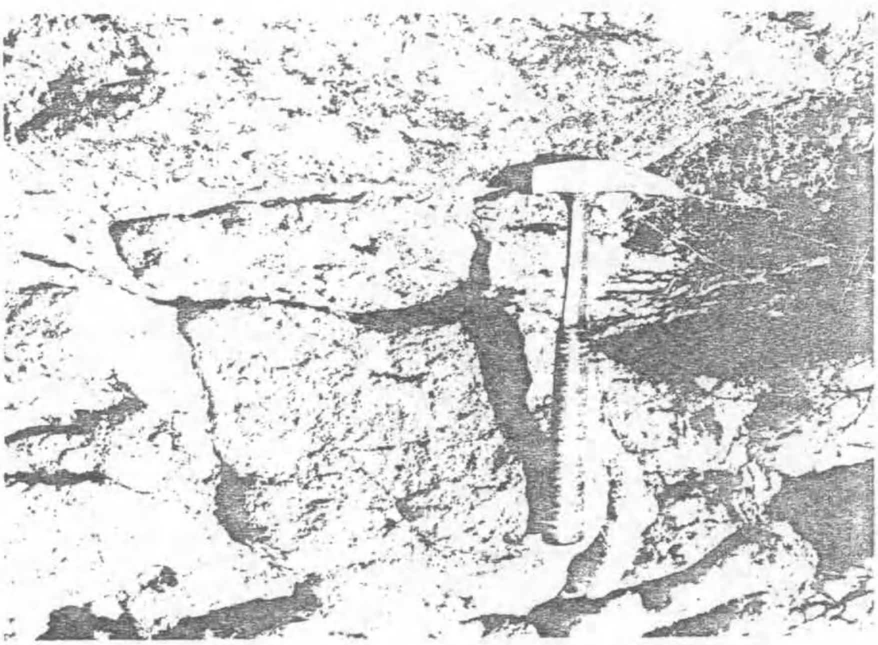


Fig. 4 Well indurated late Miocene arkosic conglomerate with clasts to cobble size and matrix of granules. Crops out on the southwest flank of High Flat, sec. 6, T. 3 S, R. 4 E.

The highest exposure of the marker is found 20 feet east of E. Brown Hollow fault at 5,760 feet elevation and is traceable eastward three-fourths of a mile to an elevation of 5,400 feet. North of these exposures scattered outcrops of the marker occur just to the east of Ruby Mountain, where it is at 5,500 feet elevation. Along the southwest flank of High Flat a similar unit varies from 5,680 feet at the contact with the southernmost High Flat bedrock exposure, to approximately 5,300 feet one and one-half miles to the northwest. One and one-half miles southeast of Anceney a small hill is capped with the same resistant marker zone at an elevation of 5,040 feet. The distribution and elevation of remnants of the marker zone may indicate subsequent tectonic movements, but exposures are not continuous, and there is no guarantee that the beds are contemporaneous. Typically this rock unit forms massive blocks which lag behind as underlying fine-grained tuffaceous material is erosionally removed. The interesting and not clearly understood aspect of the rock is its composition. Shards or tuffaceous debris are rarely present, and most grains are fresh, extremely angular quartz and feldspar fragments. Possibly the deposit represents an interval of time during which no airborne ash reached the area, or perhaps an interval of time when precipitation was markedly increased. The position and thickness of outcrops suggest the zone originated as slope flanking colluvial deposits and not as stream deposits.

Most of the other Miocene siltstones and sandstones contain about 50 per cent of grains derived from Precambrian metamorphic rock and 50 per cent of light-colored altered ash. The percentages vary from bed to bed, but they are believed to represent an average based on microscopic examination of a number of samples. Most of these beds are massive and lenticular, and individual beds are not extensive enough to be correlated from outcrop to outcrop except in a few local areas. Small-scale bedding structures are generally absent. The beds are commonly massive, a few feet in thickness, and could conceivably have been deposited in slow, sluggish streams, lakes, or as eolian deposits. Some small-scale channeling of underlying beds suggests stream deposition for some of this material. In places where lateral exposures are more extensive, a few beds were noted that had broad undulatory upper contacts, with calcite cementing the upper horizons. These beds, commonly 6 to 8 feet thick, are similar in general character to present loess deposits in the map area. Commonly throughout the area the fine- to medium-grained sandstones show an upward gradational increase of calcite cement, which suggests caliche development under subaerial dry conditions.

Vertebrate remains are concentrated in localized areas and also in the coarser beds as occasional water-worn bone fragments. The concentrations of bones generally occur where there are nearby fluvial features, such as channeling or coarser material in concentrated zones. However, the bones are usually found in very fine-grained, nonindurated, altered, ashy deposits. This suggests the possibility of dry periods and animal concentration at waterholes, with mud trapping numerous animals in localized areas. The bones are not commonly articulated but are concentrated adjacent to one another. Vertebrate remains were collected at the following localities:

SE¼, NW¼, NE¼, sec. 8, T. 3 S., R. 4 E.
NW¼, SE¼, NE¼, sec. 36, T. 2 S., R. 3 E.
NE¼, SE¼, NW¼, sec. 13, T. 2 S., R. 3 E.
SW¼, SW¼, NE¼, sec. 13, T. 2 S., R. 3 E.
SW¼, NE¼, SE¼, sec. 14, T. 2 S., R. 3 E.

Other features of the commonly exposed siltstones and sandstones are calcareous and siliceous root casts, generally of small diameters, and less commonly, what appear to be differentially cemented rodent burrows.

Near Goose Creek, in sec. 20, T. 3 S., R. 4 E., a 1- to 2-foot thick, tuffaceous, fresh-water limestone bed crops out near and east of a zone of faulting. One and one-half miles to the south, a poor exposure of thin chert beds and intercalated fine-grained, altered, ashy material was noted on the same side of the north-trending fault previously mentioned. These units are on the downthrown block and are believed to be stratigraphically higher than most of the Tertiary strata in the map area. To the north of High Flat, in a similar downthrown position with respect to postlate Miocene faults, are strata of Stratigraphic Section No. 1. Both zones display features compatible with lake deposition and are thought to be related to stratigraphically higher strata which were deposited after the adjacent bedrock highs were buried and, therefore, in relatively more valleyward environments.

In summary, the late Miocene strata varies considerably and formed from both local and exotic material. The ash was probably brought into the area as ash fall which in places remained where initially deposited and in other places was partly reworked by small stream action. In general the altered condition of the ash may support this hypothesis. Small root casts and localized areas rich in vertebrate bones suggest at least seasonally dry conditions. The type of animal remains (e.g. horses, camels, rhinoceros, and rodents) also suggest a grassland environment. However, the mixture of terrigenous debris with the altered tuffaceous material suggests that precipitation was sufficiently great to keep the bedrock high areas generally washed free of eolian ash. As burial of the pre-late Miocene paleotopography became complete, at least locally, lake deposits interbedded with ashy stream material may have become dominant.

Intrusive Rocks

In the SW¼ of sec. 30, T. 3 S., R. 4 E., just west of the eastern Brown Hollow fault, a dike of basaltic rock intrudes Precambrian rocks. This dike, trending northwest, is exposed for a few hundred feet and averages about 20 feet thick.

In NW¼, sec. 23, T. 2 S., R. 4 E., a large vari-textured gabbro dike of unknown age intrudes Precambrian rocks. Finely-crystalline chill margins indicate intrusion into cold rock, and post-intrusion fracturing indicates involvement in later tectonic movements. Post fracturing alteration of the plagioclase and augite has occurred. This rock, in thin section, is composed of approximately 45 per cent plagioclase (An_{55}) and 45 per cent augite, and the rest of the rock is composed of alteration minerals. The augite is unlike the augite characterictic of the Eocene basic volcanic rocks, and therefore this dike is apparently unrelated to them. The dike is mapped as a Cretaceous-Tertiary(?) intrusive because it could be related to the volcanic debris abundant in the Late Cretaceous and Paleocene Livingston Formation. However, it is possible the rock may be as old as Precambrian.

In several places there are generally finely-crystalline andesite porphyries intruded as concordant sills up to a few feet thick. These are essentially restricted to the Wolsey Shale and overlying Meagher Limestone. These sills are similar lithologically to those in similar stratigraphic position in the Squaw Creek area to the south, though not as coarse grained or as well developed.

Surficial Deposits and Related Features

Alluvial deposits are very common in the map area and relate to late Cenozoic history of the Gallatin Valley and surrounding areas. The following units were noted and appear on the geologic map where extensive enough to be practicably mapped. Figure 5 indicates probable correlation of these features.

Present West Gallatin River floodplain (Qal)
Terrace No. 1 (Qtg_1)
Terrace No. 2 (Qtg_2)
Terrace No. 3 (Qtg_3)
Younger pediment remnants (Qpg_1)
Older pediment remnants (Qpg_2)
Mass wasting deposits (Qm)
Loess deposits
Alluvial fans (Qfg)

The matched terraces of the map area are believed to represent periods of increased discharge and detrital increment of the West Gallatin River, directly related to periglacial climatic conditions in the drainage basin. The basic assumption is that each terrace level is related to the climatic change which accompanied the glaciation. The assumption is based on contemporary observations of glacial outwash streams (Flint, 1947) and known glaciation in the West Gallatin River drainage basin. Most observed glacial outwash streams are either in a reduced stage of incision or are aggrading, depending on the local circumstances, whereas previously, during interglacial periods, they were actively downcutting. The West Gallatin River was at least in part fed by outwash waters

Years B. P.		Probable Stade	Surficial features in map area		
0	Recent	Neoglaciation	Present floodplain (Qal) - tributary alluvium (Qal)		
5000	Wisconsin	Pinedale	Terrace No. 1 (Qtg_1)		Mass wasting features (Qm)
30,000	Wisconsin	Bull Lake	Terrace No. 2 (Qtg_2)	Loess	Large alluvial fans (Qfg)
75,000	Pre-Wisconsin	Buffalo	Terrace No. 3 (Qtg_3)		Pediment remnants (Qpg_1)
1 to 3 million	Pre-Wisconsin	Early Pleistocene or Late Pliocene	Bedrock benches		Older pediment gravels (Qpg_2)

Figure 5

Tentative correlations of surficial features during the Wisconsin and probably earlier phases of ice accumulation, but details are as yet little known in this region. The conditions which formed the thinly alluviated, paired terraces and terrace remnants are what might be expected of a stream which is loaded to the point that it cannot cut down but is vigorous enough to laterally abrade during maximum runoff.

West Gallatin River Floodplain (Qal): A floodplain is essentially absent along the southern reaches of the West Gallatin River in the map area and is restricted to a maximum of a half mile wide along the northern reaches. The river is braided slightly from the vicinity of Gallatin Gateway to the northern map boundary, but it is not known whether the braiding is active or relic at the present time. In places there is a suggestion that the widely separated channels are not the result of contemporary processes, which in turn suggests the present braids could be incised from a very recently abandoned terrace. Much of this terrace is three feet higher than the common high water mark.

Composition of gravel in the present West Gallatin River was determined by a count of the clastic material greater than 1 inch in average diameter in three square feet of area on a gravel bar near Sheds Bridge. The results are tabulated below:

- 45% Precambrian metamorphic lithologies (gneiss, meta quartzite, and amphibolite)
- 40% basic volcanic rocks (basic andesite, andesite porphyry, basalt, and vesicular basalt)
- 12% dacite porphyry (some clasts texturally border on fine-grained granodiorite, diorite and quartz diorite porphyry)
- 3% other lithologies, including the following in order of decreasing abundance: Flathead Quartzite, Quadrant Quartzite, Cretaceous sandstone, and Paleozoic chert.

The clasts are mostly well-rounded cobbles averaging 8 inches in diameter and displaying low sphericity. The matrix constitutes about 10 per cent and is composed of fine gravel and sand for the most part. However, the percentage of matrix varies considerably depending on the type of deposit. A few two-foot boulders and many one-foot boulders are present. Most of the igneous lithologies have weathering rimes 2-5 mm thick.

The rock types found in the alluvium all occur in abundance within the West Gallatin River drainage basin. However, rhyolitic volcanic rock (Hall, 1961), which is abundant along the upper Gallatin River, is not represented in the gravel. This is probably related to the lack of durability of the semi-welded tuffs, distance of travel, and their susceptability to chemical and mechanical weathering.

The present (?) floodplain is typically mantled with a varying thickness of silty deposits, presumably laid down during flood stage overflows and flooding due to ice jams. During the winters of 1961-62 and 1962-63, the writer observed numerous examples of this phenomenon just south of Sheds Bridge. Two processes were noted: (1) Silt-charged waters overflow on river ice during sudden temperature rises after prolonged below-zero weather and cause flooding beyond the normal high water level, and (2) floods form during prolonged temperature rises due to ice jams at natural constrictions. These processes may be more important than the normal early summer high-water stage floods, when overflow of banks onto the floodplain was not observed by the writer during two high-water stages.

Due to the fact that there is suggestion of incision by the present channels, the floodplain may not be actively forming at the present time. Based on position, it could conceivably correlate with the last phases of the Neoglaciation stade, believed to have climaxed in the Rocky Mountains in the mid 1800's (Richmond, 1960).

Terrace No. 1 (Qtg_1): Five to fifteen feet above the present floodplain is a more extensive terrace, averaging about a mile in width. This terrace, as well as older and higher terraces of the West Gallatin River, increases in vertical distance above the present river elevation from north to south, indicating either a different river gradient during formation or tilting after formation. These relationships are discussed in terrace slope analysis.

Silt and sand, with a thin in situ soil, cover most of the coarser underlying alluvial material. The gravel composition and size is similar to that of the present river gravels. The abundance of vegetation commonly found suggests subirrigation of the terrace, which is mainly used as pasture land.

The only direct evidence suggesting an age for this surface is located outside the study area, in the vicinity of Hell Roaring Creek, which flows from the Spanish Peaks area into the West Gallatin River south of Spanish Creek. W. J. McMannis (personal communication, 1963) has found outwash deposits from the youngest of extensive alpine glaciation in the Spanish Peaks graded to a terrace which appears to be equivalent to Terrace No. 1 because of its position relative to present river level.

These glacial features are tentatively considered as Pinedale equivalents but have not as yet been extensively studied. Terrace No. 1 is the first surface at higher elevation than the possible neoglacial terrace and is more extensive. On the basis of the preceding evidence, it is suggested that Terrace No. 1 is a Pinedale feature.

Terrace No. 2 (Qtg_2):

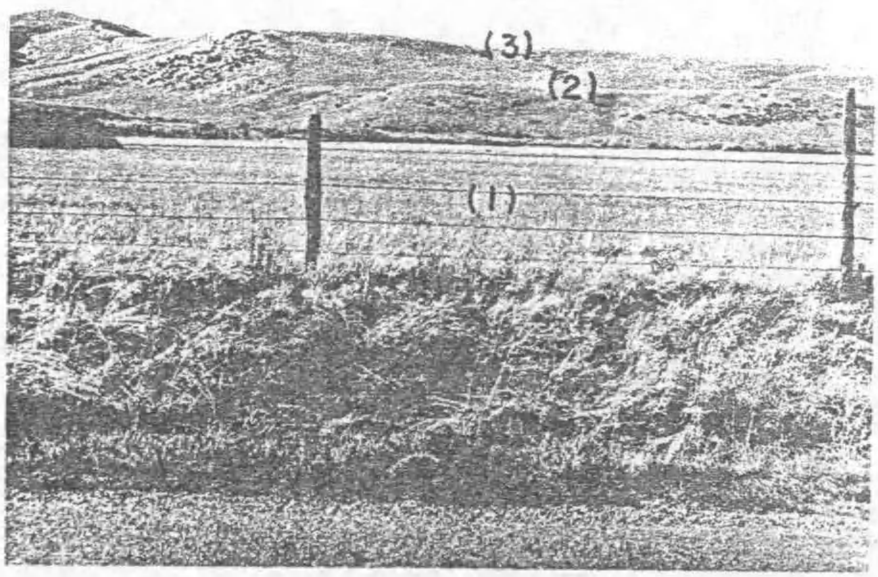


Fig. 6 Westerly view from sec. 28, T. 3 S., R. 4 E., toward the mouth of Brown Hollow. (1) a cultivated Qtg_2 terrace, (2) an undifferentiated river terrace, and (3) a northeasterly sloping Qpg_1 surface. All features are developed on late Miocene strata.

A higher, even more extensive floodplain terrace is preserved on the west side of the river (Fig. 6). Crop cultivation exposes very little coarse clastic material, apparently due to loess or river silt cover. One good exposure of the gravel deposits of this terrace is in a shallow gravel pit approximately one-half mile west of Williams Bridge, at the edge of the erosional scarp between the high Qtg_2 terrace and the Qtg_1 terrace. The percentage of various rock types represented in the clasts are as follows:

55% Precambrian metamorphic lithologies
25% Basic volcanic lithologies
5% Dacitic volcanic rock
5% Chert
5% Paleozoic sandstones
5% Mesozoic sandstones

The clasts are well rounded, of moderate sphericity, and range up to one and one-half feet in diameter and have an average diameter of about seven inches. The maximum thickness of the deposit exposed is approximately eight feet, but the total thickness is believed to be between twenty and thirty feet on the basis of well data. In places up to two feet of loess-like material overlies the coarse alluvium. This material displays no fluvial structures and is composed of an upper ten-inch humus rich zone (A horizon) and a locally absent lower one-foot caliche zone (B + C horizon). The caliche zone contains siliceous or calcareous root casts and is markedly similar to some late Miocene beds with respect to texture, composition, and general appearance.

Near the mouth of the West Gallatin River canyon, the difference in elevation between the Terrace No. 2 (Qtg_2) and the lower Terrace No. 1 (Qtg_1) is approximately 25 feet. One mile north of Williams Bridge the difference decreases to 12 feet and then remains approximately the same or less for the next five miles northward to where Terrace No. 2 is terminated by later erosion near the confluence of Fish Creek and the West Gallatin River. This change in relief between terraces poses a problem which is discussed in terrace slope analysis.

On the basis of position and the assumed Pinedale age of Terrace No. 1, Terrace No. 2 is believed to correlate with the Bull Lake stage of glaciation. As Terrace No. 2 was more extensive than younger terraces, and extensive alluvial fans appear to grade to this level on the east side of the West Gallatin River, it appears that this phase of glaciation was either longer in duration or had more profound periglacial effects.

Terrace No. 3 (Qtg_3): Along the present eastern dissected edge of the pediment system (Qpg_1) sporadic exposures of high level alluvial gravels occur which are similar to the younger terrace gravels and to present West Gallatin River gravels (200 ft. above Qtg_2). Representative data was obtained from the northernmost exposure of these deposits in the map area, located one mile west of Sheds Bridge, just north of the highway in a barrow pit (Fig. 7). The following is the percentage composition determined by a count of the clasts greater than 1 inch in average diameter in three square feet of exposure.

- 52% basic volcanic rocks (andesite porphyry, basalt, basic andesite, vesicular basic andesite)
- 26% Precambrian metamorphic lithologies (amphibolite, meta quartzite, gneiss, and schist)
- 14% Dacite porphyry (approaches quartz diorite, diorite or grandiorite)
- 4% Quadrant Quartzite (Pennsylvanian)
- 2% Flathead Quartzite (Cambrian)
- 2% Other lithologies, dominantly Paleozoic chert and Cretaceous sandstone.

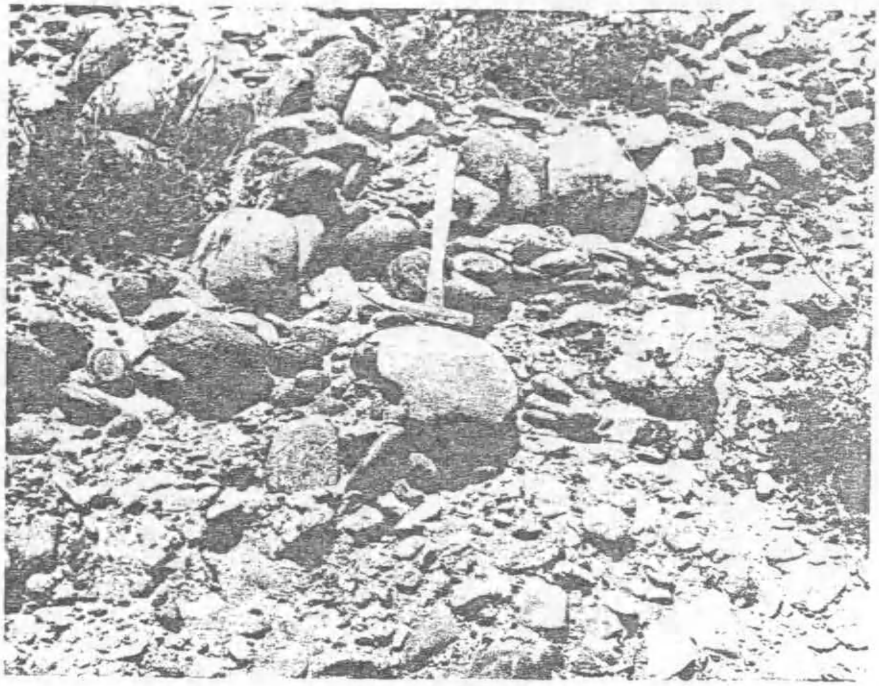


Fig. 7 Gravel deposits of an old West Gallatin River terrace, Qtg_3. Exposure in a gravel pit one mile west of Sheds Bridge in sec. 10, T. 2 S., R. 4 E.

Clasts are medium to well rounded, with moderate to low sphericity. Average size is about seven inches in average diameter , and the maximum is about twelve inches. The matrix is estimated as 25 per cent of the total volume and consists primarily of coarse sand granules and pebbles of the same lithologies as above. Poorly developed polish is present on some gneiss and basic volcanic clasts. The dacite porphyry and some basic volcanic clasts are deeply weathered. Imbrication indicates a northward direction of flow during deposition.

By comparison these deposits are closely related to the present West Gallatin River gravels and to the pediment gravels. Only deposits related to the West Gallatin River have dacite porphyry clasts and measurable percentages of Cretaceous debris. In secs. 27 and 33, T. 2 S., R. 4 E., loess- like lenses up to 10 feet thick occur between the coarse gravels, indicating contemporaneous deposition. The position of these gravels at the pediment shoulders indicates the pediments were graded to this terrace level.

These river deposits, on the basis of position, are tentatively assigned to the Buffalo stade of glaciation. Further discussion of age is presented in the section dealing with the equivalent pediments.

<u>Other River Terrace Remnants:</u> Near the mouth of the West Gallatin River canyon, on the west flank of the canyon, are a number of terrace remnants, some of which are too small to be practicably mapped. At higher elevations in the same area are small possible river benches with no foreign material resting on them. Terrace remnants with alluvial material consisting of cobbles and boulders of metamorphic lithologies, basic volcanic lithologies, and some limestone, sandstone, and dacite clasts, occur in sec. 5, T. 4 S., R. 3 E., and sec. 32, T. 3 S., R. 4 E. The highest gravel-covered terrace remnant occurs at 5,490 feet elevation, about 400 feet above the present West Gallatin River. This was the only gravel-covered remnant found at this elevation. The next lower terrace remnant on the flank of the canyon is at 5,415 feet elevation. Approximately one-half mile north on a ridge of Meagher Limestone a veneer of

river gravel occurs at the 5,410-foot elevation, and these probably form a terrace remnant equivalent to one of the canyon remnants. These lie on the upstream projection of the profile of Terrace No. 3 (Qtg_3) (Pl. II, Fig. 4). About one-fourth of a mile apart, at the 5,305-foot elevation on the canyon flank and at the 5,235-foot elevation on the Meagher Limestone ridge, small gravel-covered terrace remnants occur; these also apparently are equivalent to each other. The most extensive terrace remnant, mapped as Qtgu, occurs at the 5,210-foot elevation, about 120 feet above Terrace No. 2 and about 130 feet below the projected Terrace No. 3.

At elevations higher than the highest gravel-covered terrace remnants, small bench-like surfaces are notched into the west flank of the canyon. These are commonly of small areal extend and occur at various elevations up to 6,400 feet. Similar surfaces do not occur on the very steep east side of the canyon. Because the benches appear to be unrelated to the trend of foliation in the underlying gneiss a stream cut theory of origin is favored. Assuming these features are river cut benches, the question arises as to why gravels are absent. In places the high benches are of greater areal extent than the lower gravel-covered remnants and, therefore, presumably provide ample area for gravel preservation.

Three plausible explanations seem apparent, and all may be contributing factors. They are (1) The high benches probably were cut before one or more stades of glaciation during which there were periods of increased precipitation, solifluction, and generally accelerated erosional processes, which have stripped the benches of alluvium. (2) The remnants are small in areal extent and bounded by steep slopes; therefore, normal erosional processes over a longer period of time have stripped these areas of alluvium, whereas lower and younger terrace remnants retain the gravel. (3) Prior to glacial activity, streams may have been engaged in removing large amounts of fine-grained, poorly indurated Tertiary strata and thus may have been carrying only minor amounts of bedrock derived gravel. Possibility (3) has some merit in view of the abrupt vertical change from gravel-covered terraces to barren ones. If possibility (2) were solely responsible for the barren benches, one might expect a gradual decrease in amount of gravel on successively higher stream-formed features. Such gradation is not apparent from the field studies.

Position with respect to Terrace No. 3 (Buffalo ? age) suggests an early Pleistocene to late Pliocene age for the bedrock benches. Some of the lower bedrock benches may be contemporaneous with the highest gravel deposits (Qpg_2) found further north, believed to be old pediment remnants which are discussed in a following section.

Analysis of Terrace Slopes: On Plate II, Fig. 4, fall line profiles have been plotted for the river terrace remnants as well as for the present West Gallatin River for comparison. Also plotted are the high bedrock-cut benches of the West Gallatin River Canyon. Even though some of the lines shown are not directly adjacent to others, all are in true lateral position in that they subparallel each other in a northerly direction.

PLATE II

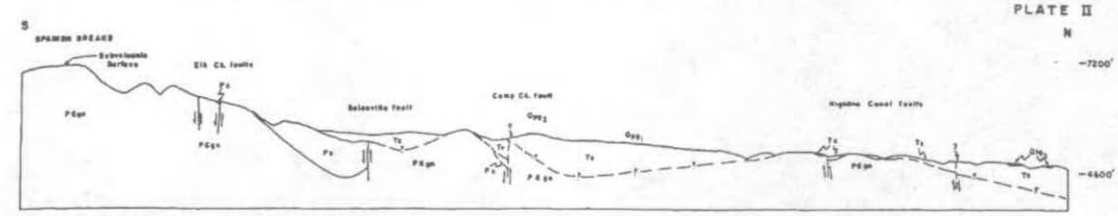


Figure 1. N-S cross section of map area, A-A' on plate I

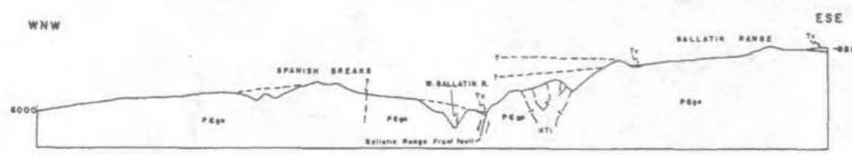


Figure 2. E-W cross section along subvolcanic surface remnants, includes areas adjacent to map area.

GENERAL KEY

- Qtg1 river terrace
- Qtg2 river terrace
- Qtg3 pediment equivalent river terrace
- Qpg1 pediment gravel
- Qpg2 older pediment gravel
- Ts late Miocene tuffaceous strata
- Tv Eocene mafic volcanic rock
- KTi Cretaceous-Tertiary intrusive rock
- Pz Paleozoic sedimentary strata
- PЄgn Precambrian gneiss

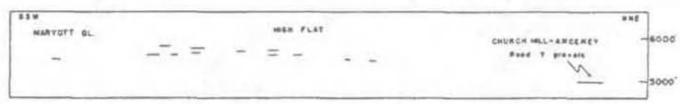


Figure 3. Profile distribution of Qpg2, with varying elevations suggesting pediment remnants.

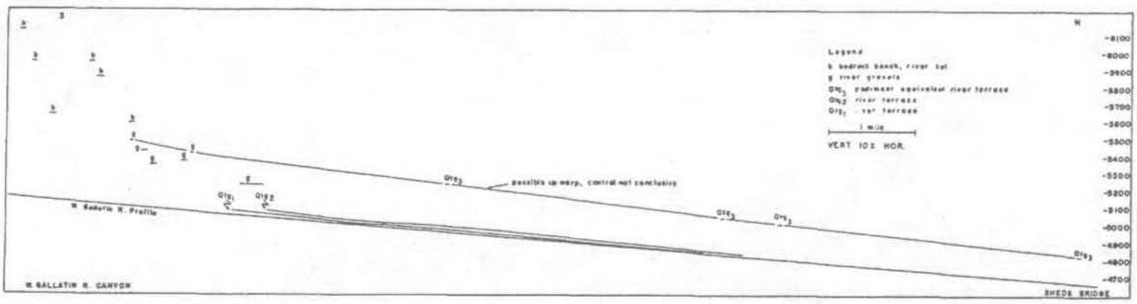


Figure 4. Terrace profiles, and restored Qtg3 profile.

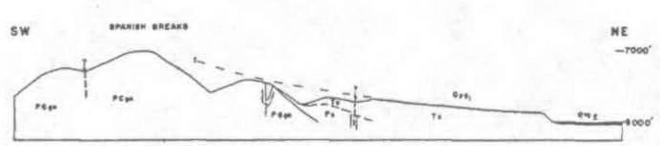


Figure 5. Pediment remnant and probable source, B-B' on plate I.

It is apparent that Terrace No. 3 has the highest gradient or slope, and the successively lower profiles have successively lower gradients or slopes, with the present river having the lowest gradient. The two terrace remnants on the southern end of Terrace No. 3 profile unfortunately are not definitely correlatable with the main part of the terrace but are believed to be essentially equivalent on the basis of their positions relative to the projected profile.

The cause of different slopes for noncontemporaneous river terraces is at present very difficult if not impossible to ascertain due to indeterminate factors which may have influenced the gradients of the deposits. Mackin (1948) has pointed out numerous factors which may control the gradients, most of which can not be determined from deposits alone. Therefore, even though the deposits of the various terraces in the map area may be essentially similar in character from the standpoint of grain composition, size, and sorting, there is no reason to believe that ancient West Gallatin rivers necessarily had slopes or gradients identical to those of the present river. On the contrary, if the terraces represent periglacial features, different gradients might be more likely.

Two possible interpretations can be made from the diagram: (1) the profiles reflect the gradients of the streams at the time of formation, or (2) the slopes of the profiles are not the gradients of the streams at the time of formation, but have been rotated or tilted. If (1) is assumed, gradient controlling factors of the earlier streams must have been considerably different from those of the present West Gallatin River in that they display higher gradients. As the alluvial material associated with the higher remnants is very similar to the present West Gallatin River alluvium in size and composition, one might assume smaller discharge or greater detrital increment to account for a higher gradient. Recent stream capture at some upstream point could cause a lesser discharge in the past, but no known evidence is present to support this possibility. A more arid climate is also a possibility, and perhaps is suggested by the polish found on some clasts in the pediment gravels and equivalent river deposits. However, the polish indicates only that semiarid or arid conditions have existed since deposition and not that conditions were such at the time of deposition.

Evidence in the Gallatin Valley strongly favors possibility (2). Hackett, and others (1960) have noted abnormally thick deposits of Quaternary alluvium eight miles north of the map area near the town of Belgrade. In the map area between Gallatin Gateway and Sheds Bridge the low terrace gravel deposits are 70 to 80 feet thick. Northward from the map area the deposits thicken gradually, and near Belgrade reach at least 400 feet in thickness and may be as thick as 800 feet (based on an oil test well: State well #1). Where the river leaves the Gallatin Valley at Logan test drilling indicates 20 to 30 feet of alluvium resting on bedrock (Madison Group). Therefore, because the base of the alluvium near Belgrade is at lower elevations than the bedrock threshold where the river issues from the Gallatin Valley, tectonic subsidence is necessary to produce such a thick alluvial deposit of Quaternary age. If the hypothesis of tectonic instability is assumed, the relative increase in slope of each terrace remnant with respect to the younger and lower terrace remnants might reflect a tilting of the map area, acting gradually throughout the period of time represented by the alluvial features. On the basis of the tentative ages assigned to the terraces the tilting has been active from early Pleistocene to perhaps the present. In a relative sense, the Spanish Breaks area may have been rising or the center of the Gallatin Valley subsiding. However, the entire map area has been structurally downdropped an unknown amount relative to the Gallatin Range since late Miocene time; therefore, the logical assumption would be that the Spanish Breaks area should have subsided with the rest of the Gallatin Valley, perhaps at a lesser rate than the center of the valley.

If the tilt mechanism is accepted, then the most logical place for the northward tilt to cease is along the Gallatin Range front fault, assuming that valley margin structures and the recent tilting are related. South of the point where the tilt ceases, the terraces should essentially parallel each other southward if the original gradients were not too different. No detailed work has been done on terraces to the south in the Gallatin River Canyon, but terrace remnants are present. W. J. McMannis and C. C. Bradley (personal communication, 1963) have noted shoulders with coarse alluvial material at several widely separated points on the flanks of the upper West Gallatin River canyon, at approximately 350 to 400 feet above and more or less parallel to the present river level. Positions of these remnants with respect to the present river from Gallatin Gateway to Dudley Creek, some 25 miles upstream, indicate the remnants are probably equivalent to the highest Gallatin River terrace of the map area (Qtg_3).

Further to the south, just north of the Yellowstone Park border near the upper reaches of the West Gallatin River, Hall (1961) has mapped and described gravels overlain by flows of rhyolitic tuff that he believes to be Pliocene in age (based on indirect evidence). These gravels are not continuous, but appear to be restricted to areas adjacent to and at approximately 300 to 400 feet above present river. The description of the gravels suggests they were deposited by a fairly large stream or streams, possibly similar to the present river. Morainal material of two possible stades of glaciation rest on the rhyolitic flows, indicating the flows and underlying gravels are at least older than some of the glaciation and perhaps older than Pleistocene.

These gravel deposits are critical to the regional tectonic history of the Gallatin-Madison Range. From the area in which these gravels occur, the West Gallatin River flows northward through a steep-walled canyon 1,000 to 2,000 feet deep, carved for the most part in gneiss, to the Spanish Creek Valley. If the Pliocene (?) gravels are ancestral Gallatin River gravels, and if no relative tectonic uplift of the Gallatin Range-Spanish Peaks structural block has occurred between Pliocene and the present, then the canyon must have been present during Pliocene in essentially its present configuration and since has been modified only by 300 feet of deepening. This also would imply that the terraces at 300 feet above the present river in the map area (Qtg_3) are Pliocene (?) in age. However, the evidence presented in this study suggests pre-Wisconsin (Buffalo stade) equivalency for Terrace No. 3. On this basis, and the configuration of the canyon, the writer favors relative uplift of Gallatin-Madison Range since Pliocene and the related tilting of the canyon mouth area. However, it should be made clear that this is an opinion based on tentative datings and indirect evidence, and as yet no firm conclusion can be stated.

Pediment Remnants (Qpg_1): Several partially dissected northeast-slopping pediment remnants flank Terrace No. 2 on the west. These remnants are formed on late Miocene tuffaceous strata and terminate toward the northeast at elevations 200 to 400 feet higher than Terrace No. 2. In places, at their dissected end, gravels of the pediments intertongue with deposits of Terrace No. 3 (Qtg_3). The pediments occur in the area between the mouth of the West Gallatin River canyon and High Flat. Pediments of this generation, north of High Flat, were either originally of poor development or have been subsequently modified by erosion and have lost much of their characteristic identity. Fall-lines on the pediment remnants trend northeast parallel to present linear dissecting ravines and incised streams and have an average gradient of 160-200 feet per mile. The surfaces are under dry land cultivation made possible by a thin layer of loess lying on the pediment gravel. The gravel may reach a maximum thickness of about 15 feet but appears to average a few feet in thickness. The loess cover varies from 0 to about 10 feet in thickness generally being about 1 foot, and large areas of the surfaces yield pebbles and boulders upon cultivation.

Gravel characteristics vary from the southern to northern pediment remnants with respect to average grain size and composition. The southernmost exposed pediment gravel has an average diameter of approximately 8 inches, though one 3.5 foot boulder was noted. Most of the clasts are well rounded and of moderate sphericity. The approximate percentages of major rock types represented in the gravels of the southern remnants is as follows:

65% basic volcanic lithologies
10% Precambrian metamorphic lithologies
10% Paleozoic carbonates
10% Cambrian Flathead Quartzite
5% blue and tan chert (probably Paleozoic).

Farther north, on pediment remnants just south of High Flat, (Fig. 8) the gravel includes approximately 50 per cent basic volcanic rock and 50 per cent Precambrian metamorphic rock, and the average diameter of the clasts is about 3 inches. Clasts derived from Paleozoic strata are rare to absent and silicified wood fragments (probably derived from the basic volcanic sequence) are common. The pediment remnants south of High Flat are at slightly varying elevations. However, there is no noticeable difference in gravel type at different elevations.

The composition and position of these gravels indicate the remnants are pediments and not tilted river terraces (the latter might be plausible in view of very recent tectonic activity in some areas of southwestern Montana). These gravels differ in composition and grain size from those of the present West Gallatin River and the ancestral West Gallatin terrace gravels. Locally ancestral West Gallatin River gravels of Terrace No. 3 (Qtg_3) lie adjacent to and appear to intertongue with the eastern limits of the pediment gravels, indicating the two were deposited essentially at the same time.

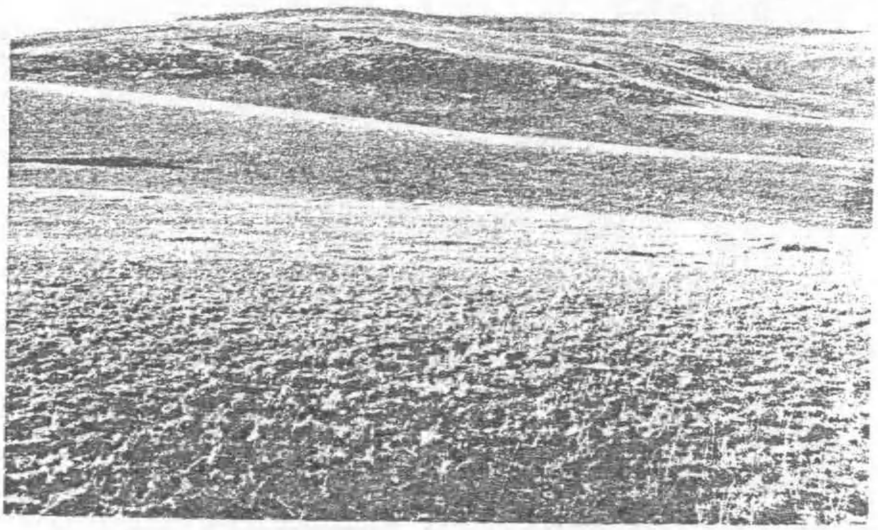


Fig. 8 View northward from sec. 6, T. 3 S., R. 4 E., toward the south flank of High Flat Precambrian bedrock high (background) and gravel (Qpgl) covered late Miocene strata in the middleground and foreground. Gravel surface is near the head of Qpgl pediment system.

The southern pediments are beheaded, and the pediments south of High Flat in part seem to head in the older pediment (?) gravel deposits (Qpg_2). Figure 9 illustrates the typical relationship between Qpg_1 and Qpg_2 deposits. Composition of the gravel (Qpg_1 and Qpg_2) poses a problem, but this subject is deferred to the following section on the higher gravel deposits. If the interpretation of Qpg_1 heading in Qpg_2 is correct, the

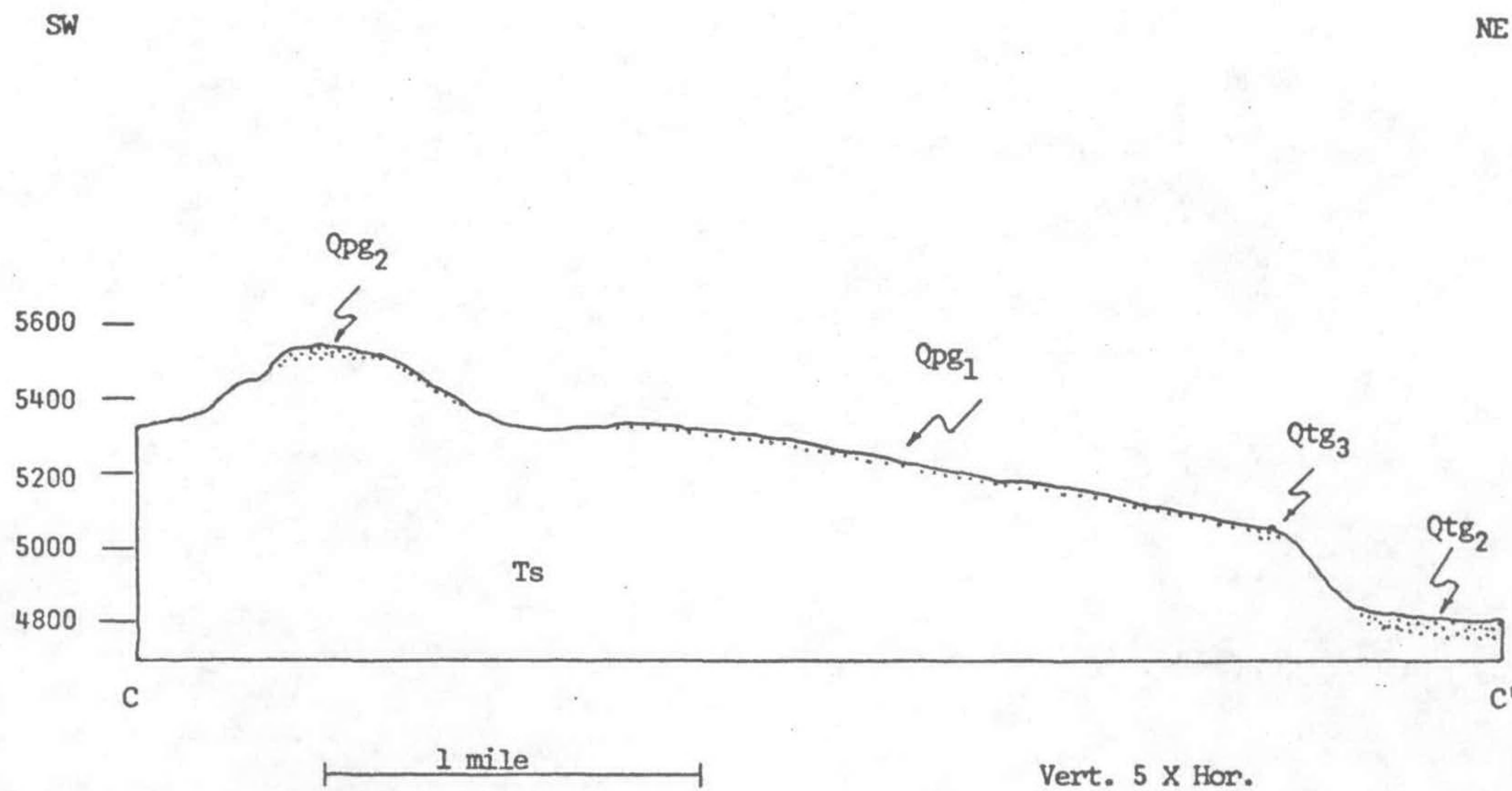


Figure 9 Typical relationship of pediment gravels in the map area. Qpg$_2$ is the probable source for Qpg$_1$.

pediments just south of High Flat are essentially preserved in their entirety, and their original configuration is altered only by longitudinal dissecting ravines.

The age of the Qpg_1 pediments is not directly determinable from field evidence. The surfaces may correlate with similar pediment remnants common in many areas in southwestern Montana, but no direct correlation appears possible. Furthermore, there is no certainty that the highest and consecutively lower surfaces in one area are contemporaneous with the highest and consecutively lower in adjacent areas, though in many cases investigators have made this assumption. The highest surfaces have been considered to be Flaxville equivalents, or called bench #1 and tentatively dated as Pliocene, Alden (1953), Horberg (1940). Studies by Bluemle (1962) and Pardee (1950) suggest that the higher surfaces in areas adjacent to glaciated terrain are overlain by or contemporaneous with moraines of undetermined Pleistocene stades.

The age of pediment system (Qpg_1) as suggested by position and intertonguing relationships with Terrace No. 3 is tentatively believed to be equivalent to pre-Wisconsin glaciation. Other evidence which supports correlation with some stade of glaciation, but not necessarily pre-Wisconsin, follows:

In the map area, course clasts, a few of which are 2 to 3 feet across, indicate the pediments were cut by streams with much greater competence than that of the present-day dissecting streams. Also, loess deposits interbedded with contemporaneous ancestral West Gallatin River alluvial deposits suggest that at the time of pediment formation eolian processes known to occur in periglacial environments were in progress (Flint, 1947).

Furthermore, conditions which appear necessary to the formation of pediments can be associated with periods of glaciation. Many investigators favor lateral planation to explain pediment formation (Gilbert, 1877; Johnson, 1932; Bluemle, 1962; and many others). If pediments formed on weakly resistant sedimentary strata are produced by lateral stream planation, extensive pediments must indicate a pause in downcutting of the master stream which controls local base level of the tributaries. In a periglacial situation, the master stream carrying abundant glacial melt water and heavy bedloaded is rarely downcutting (Flint, 1947). Generally the stream is either aggrading or in a reduced rate of incision, depending on the local circumstances.

The pediment remnants in question are extensive, and therefore the master stream (the West Gallatin River) must not have been actively downcurring. The loess deposits, the coarse pediment gravels, and the extensive pediment surfaces lead the writer to suggest that the pediments formed during a period of glaciation. Development during the Pliocene seems to be incompatible with expected climatic and erosional changes that should have occurred later during Pleistocene glaciation, for it is difficult to rationalize erosion of only 200 feet of material from local valley areas and minor pediment dissection during the one to three million year duration of the Pleistocene. Basically this minor erosion is all that has happened since the pediments formed; therefore, a Pleistocene age appears probable for the pediments (Qpg_1) and equivalent river terraces (Qtg_3).

High Gravels (Qpg_2) Scattered throughout the northern and central parts of the map area on interdrainage divides are gravel deposits that may be older and are slightly higher than the gravels preserved on the well developed pediment remnants. Locally, these deposits appear to be up to 25 feet thick, though accurate measurement of thickness is impossible because of poor exposures. Most of these remnants occur on high parts of the divide between the Camp Creek and West Gallatin River drainages (Fig. 10). Also, north of High Flat outcrops of similar gravels occur on the highest

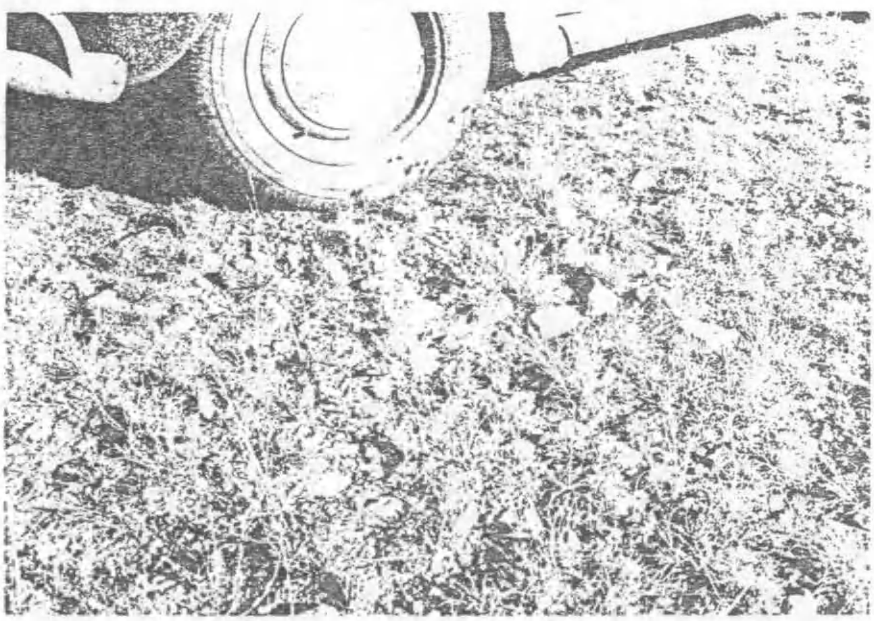


Fig. 10 Typical of Qpg_2 gravel remnants south of High Flat. Sec. 6, T. 3 S., R. 4 E.

interstream divides, in various degrees of preservation and exposure (both usually poor). In the latter area one deposit is well exposed just northwest of the Anceney-Church Hill road junction at elevations of 4,960 to

5,000 feet. These exposures consist of conglomeratic sandstone and conglomerate (Figs. 11 and 12). The twelve feet of section exposes buff to medium-brown sandstone and conglomerate in beds ranging from 6 inches to 2 feet in thickness. The sandstones are commonly conglomeratic, but otherwise are very similar to late Miocene strata. The conglomerate contains an estimated 80 per cent basic volcanic clasts, 10 per cent Precambrian metamorphic rock clasts, and a few clasts of Quadrant Quartzite, Cretaceous sandstone, silicified wood fragments, chert, and reworked Miocene (?) conglomerate. Size of clasts ranges from granule to cobble, with an average 2-to 4-inch diameter. Imbrication indicates stream flow was slightly to the east of north. One bone was collected from the exposure by W. J. McMannis and sent to C. Lewis Gazin at the Smithsonian Institution. The following is Gazin's opinion: "The bone you sent me for identification is a phalanx or toe bone of either an antilocaprid (pronghorn) or cervid (deer). It cannot be determined generically. In some ways it seems somewhat closer to an antilocaprid. This gives us almost no information on geologic age, as these forms could have been found in almost any part of the Pleistocene, or possibly a little earlier." This is taken to mean that the gravels are more likely Pleistocene, but could be late Pliocene in age.

The remnant gravel deposits adjacent to and south of the High Flat area generally increase in elevation from 5,480 feet on the north to nearly 5,800 feet on the south. The areal distribution and elevations of these gravels suggest similarity to the younger pediment (Qpg_1) surfaces in mechanism of origin. The gradients of the older gravel-covered surface

(1)

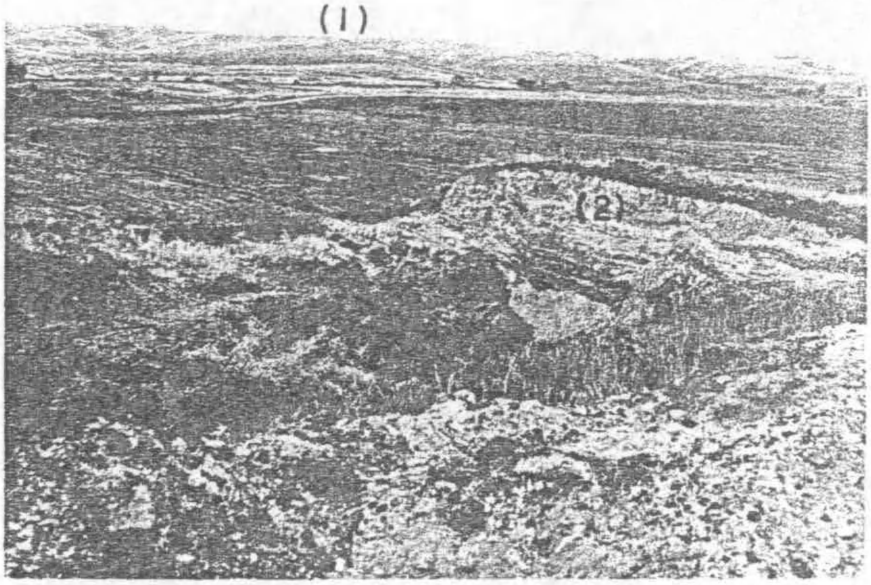


Fig. 11 Southwesterly view from sec. 8, T. 2 S., R. 4 E., of: (1) High Flat and (2) Qpg2 exposure in gravel pit.

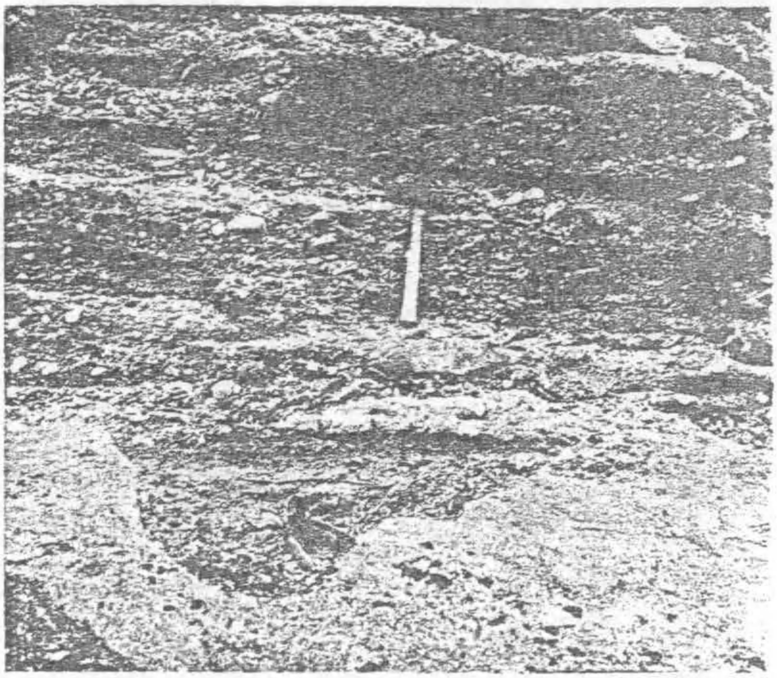


Fig. 12 Channeling and imbrication in Qpg2 deposits north of High Flat in sec. 8, T. 2 S., R. 4 E. It was not determined whether the lower deposits in the picture are late Miocene in age or part of the Qpg2 deposits. If late Miocene in age, the deposit is abnormally conglomeratic as compared to adjacent late Miocene strata. North to right.

appear to slope northeast across High Flat, and in the south, northeast to an ancient West Gallatin River (Plate II, Fig. 3). This relationship is similar to that of the younger well-preserved pediments, differing only in that the younger surfaces butt against rather than cross High Flat. The composition of the high gravels is similar to that of the younger pediments; that is, the bulk of the clasts are 2 to 4 inches in diameter and composed about equally of Precambrian metamorphic lithologies and basic volcanic lithologies. Some silicified wood fragments and Flathead Quartzite clasts also occur.

The percentage of basic volcanic clasts is greater in the Qpg_2 gravels north of High Flat than in gravels south of there. This and other evidence discussed under Tertiary stratigraphy suggests that the High Flat area may have supported fairly extensive basic volcanic remnants, the last of which was removed when the late Miocene cover was removed during pedimentation.

There is a possibility that the high gravels (Qpg_2) are part of the northern Qpg_1 pediment remnants and not old gravel remnants of some earlier stream system. However, the thickness of the remnants seems greater than typical pediment veneers (Qpg_1), even though thicknesses are estimates based on indirect evidence and poor exposures. Perhaps the best evidence for two separate deposits is the topographic relationships between the high gravel deposits (Qpg_2) and the source of the extensive pediment remnants (Qpg_1) (Fig. 9 and Fig. 13).

It therefore seems probable, from the elevation of the gravel remnants, as well as the nature of the gravel, that these deposits may be remnants of an earlier pediment system or systems. The gravel remnants (Qpg_2)

supplied some of the clasts found on the younger pediments (Qpg_1). However, the original source for the high gravel deposits is a problem because these older surfaces apparently head to the south or southwest, where late Miocene strata lap directly onto bedrock other than basic volcanic rocks.

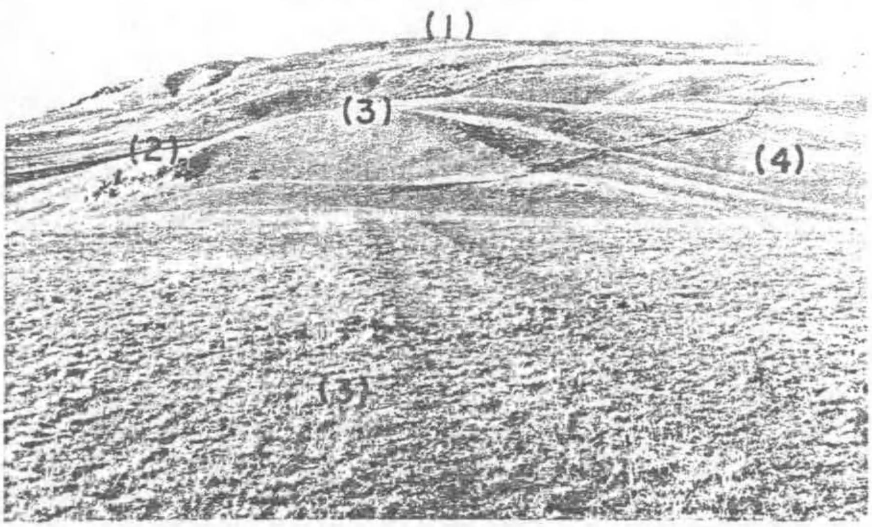


Fig. 13 View northward from sec. 6,T. 3 S., R. 4 E., of: (1) High Flat, (2) exposure of late Miocene strata, (3) Qpg2 gravel remnants, and (4) poorly developed head of Qpgl pediment system.

Projection of the pediments (Qpg_1) southwestward shows that they head in and against the topographically higher Spanish Breaks where only Pre-cambrian metamorphic rocks and early Paleozoic strata are now exposed (Plate II, Fig. 5). Lapping directly onto these rocks are the tuffaceous sedimentary strata of late Miocene age. Projection of the older and higher pediment system (Qpg_2) creates a similar relationship. The summit of the Spanish Breaks ridge is a flattish surface, suggesting it was formed by erosion unrelated to the present dissecting cycle. Therefore, coarse basic volcanic clasts, which are the most abundant rock type in the pediment gravels, apparently have three possible sources:

(1) The gravels were derived from the Spanish Breaks area; the volcanic source rock being stripped during and/or after the phase of pediment formation.

(2) The basic volcanic clasts in the gravels came from conglomerates in pre-existing late Miocene strata, or from older and higher gravel surfaces now removed by erosion.

(3) The gravels were not derived from the Spanish Breaks topographic high and thus must have originated to the south or southeast of that area.

Possibility (3) is the least plausible for the following reasons: The pediment remnants in question appear to be graded to an ancestral West Gallatin River that flowed approximately 300 feet higher than the present West Gallatin River. The latter has Spanish Creek as a tributary paralleling the Spanish Breaks area on the south in a partially exhumed tuffaceous sediment-filled valley, similar to those in the area here studied. If this entire region were still blanketed by thick Miocene deposits at the time of pediment formation and the volcanic debris were brought northwestward from the Gallatin Range across the Spanish Breaks and on to the pediments, then the pediments could not have been graded to an ancestral West Gallatin River only 300 feet higher than present river level. This possibility is contrary to evidence in the study area, for grade and trend of the pediment deposits merge with terrace deposits (Qtg_3) that, in turn, indicate the ancestral river was present, flowing northward as it is now, and carrying essentially the same rock types as in the present bedload.

Possibility (2) might be ruled out for the following reasons: Remnants of Cenozoic conglomerates which could have supplied the basic volcanic clasts of the pediment gravel are lacking along the southern margin of the Gallatin Valley. All the late Miocene sedimentary strata in this part of the Gallatin Valley are dominantly poorly indurated fine-grained tuffaceous sedimentary strata. In the area studied, some zones of coarser material were noted; one example consists of a widespread siliceous conglomeratic arkose lense locally up to 30 feet in thickness but lacking volcanic debris. Others are local channel deposits in the northern part of the map area, which contain pebbles up to a few inches in diameter, many of which are basic volcanic rock. Neither type of deposit could provide the voluminous source necessary for the much coarser and widespread pediment gravels.

The nearest coarse-grained older conglomerates crop out on Beacon Hill, just east of the town of Bozeman, 12 to 15 miles east of the map area. These late Miocene and/or Pliocene beds contain conglomerates which include clasts up to boulder size, with imbrication indicating deposition by west-flowing streams. These conglomerates contain basic volcanic clasts but also contain distinctive igneous rock types not present in the gravels of the map area. Derivation of pediment gravel from possible reworking and westerly transport of these conglomerates is therefore ruled out.

Possibility (1), a source in the Spanish Breaks area, seems to be the best hypothesis for the origin of the high pediment gravels and perhaps in part for younger pediments. This interpretation is supported by a small downfaulted block of basic volcanic rocks adjacent to the Gallatin Range front fault, where these volcanic materials rest on an isolated remnant of the crestal Spanish Breaks surface. The latter remnant has been isolated from the Spanish Breaks proper by incision of the West Gallatin River Canyon (Fig. 22). From a point about 3 miles west of the canyon, the flattish crestal surface descends gradually southeastward toward the volcanic remnant and fault but is interrupted by the canyon notch (Plate II, Fig. 2).

If possibility (1) is correct, then the crestal surface of the Spanish Breaks area is an exhumed pre-middle Eocene erosion surface and not a remnant of a "Late Tertiary Peneplain" as proposed by Pardee (1950). This data is critical in clarifying part of the regional Cenozoic history of southwestern Montana and automatically raises the question: How many similar bedrock cut surfaces in southwestern Montana and northern Wyoming are exhumed erosional surfaces, preserved at the present time because of relatively recent stripping of protecting covers?

The age of the high pediment gravel (Qpg_2) is debatable. Pre-Wisconsin glaciation probably consisted of several glacial advances, and the high gravel remnants are less than 100 feet higher than pediments tentatively assigned a mid-Pleistocene age. Therefore, it seems probable that the high gravels (Qpg_2) are also mid-Pleistocene in age, but it should be pointed out that the antilocaprid or cervid bone from these gravels could be as old as late Pliocene.

Other Terraces: The incised ravines of both intermittent and perennial streams of the pediment area bear clasts derived from the pediment surfaces. Also common are very small terrace remnants capped by similar gravels. These alluvial features were not mapped because of their restricted and localized nature.

Alluvial Fans (Qfg): On the east side of the West Gallatin River, and lying for the most part beyond the map area, are several alluvial fans which are not actively forming at the present time. These fans were formed by Big Bear Creek and Cottonwood Creek, both of which flow north from the Gallatin Range into the Gallatin Valley. The northwest part of the fans have been truncated by Terrace No. 1 (Qtg_1) river action, and the fans appear to have originally graded to the Terrace No. 2 (Qtg_2) level. However, on the east side of the river the latter terrace is not clearly differentiated from the fans. Middle Creek flows from the Gallatin Range on part of a large fan (the Bozeman fan of Hackett and others, 1960) just east of the map area. This fan also appears to grade to the Terrace No. 2 level, though the relationship is not clear cut.

The lower fan deposits are coarse and somewhat similar to the West Gallatin River terrace gravels, with the following exceptions: The percentage of Paleozoic carbonate debris is much higher, no dacite clasts are present, and average clast size is apparently much smaller, averaging about 3 or 4 inches in diameter.

Because of the relationships of the fans to Terrace No. 2, they are tentatively assigned a Bull Lake age. It is possible that Pinedale glacial processes added debris to the headward parts of the fans, but if so fluvial activity was not vigorous enough to dissect and regrade the fan to Terrace No. 1.

One very small fan on the west side of the West Gallatin River grades to Terrace No. 1, north of the mouth of the West Gallatin Canyon but is not actively building at the present time.

Mass Wasting Deposits (Qm): In areas adjacent to steep slopes on Paleozoic strata, numerous mass wasting deposits occur. On the geologic map the more extensive deposits have been mapped as Qm and include deposits which may have originated as landslide, talus, solifluction, and colluvial accumulations. For the most part these deposits are generally composed of carbonate and quartzite rubble covered in part by soil. At one locality in the southeast 1/4 of sec. 2, T. 3 S., R. 3 E., just north of Ruby Mountain, scattered large blocks of gneiss rest on Tertiary strata, and the mechanism of emplacement is obscure. One-half mile to the south of the blocks typical mass wasting deposits of carbonate occur on the flanks of Ruby Mountain, essentially covering gneiss bedrock, which is in fault contact with the topographically higher Paleozoic carbonates. Adjacent gneiss outcrops to the west of the blocks are not much higher and are freshly exhumed, which suggests some other source for the angular gneissic blocks. If the gneissic blocks were derived from the Ruby Mountain flank to the south they must be older than the carbonate debris and may have been let down essentially in place from a higher surface by gradual erosion of the fine-grained underlying Tertiary strata.

Based on topographic relationship to the pediments (Qpg_1), most of the mass wasting must be Wisconsin or younger in age because of their positions below projected pediment profiles.

Loess: Loess deposits cap areas of low relief throughout most of the lower parts of the map area. Known thickness of the loess rarely exceeds 15 feet, and probably averages about 2 to 3 feet. However, exposures of the loess generally are in cuts on topographic highs; therefore, the thickness estimate may be quite conservative (Fig. 14).

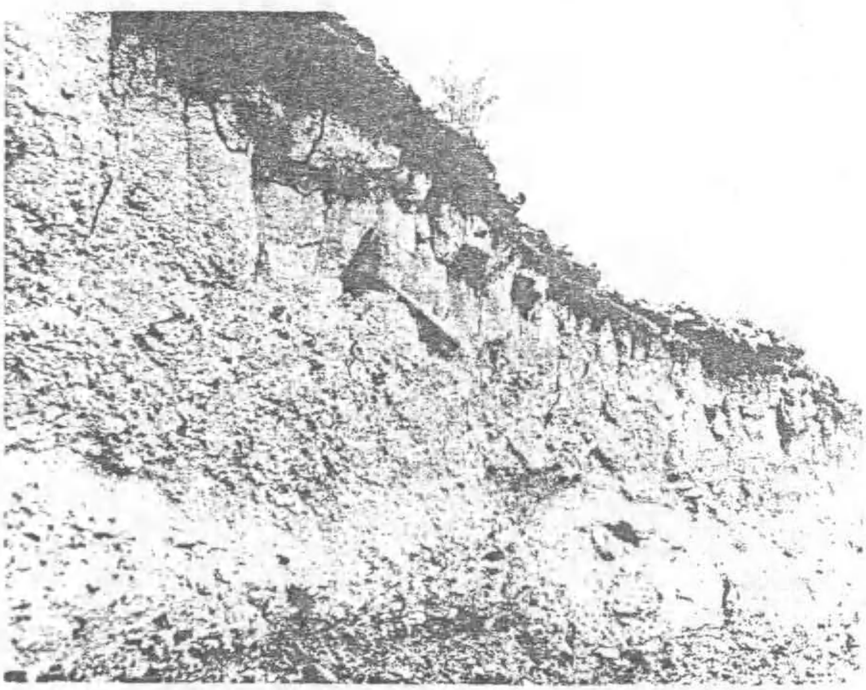


Fig. 14 Loess deposit resting on Qpg_2 gravels north of High Flat in sec. 8, T. 2 S., R. 4 E. Hammer rests on the lower contact of the loess.

In places, the loess has A, B, and C horizons developed in the upper few feet of exposures. The C horizon is almost identical to portions of many beds of the late Miocene strata, and in many localities of poor exposures (e. g., areas of rodent burrow mound mapping), it is very difficult to decide which type of deposit is present.

At the present time loess-like material is accumulating in the map area. Evidence for this is based on observation of partially melted snow drifts in early spring which are commonly black with wind-blown soil. Observations by the writer indicate that this is related to the present cultivation in the Three Forks basin and probably not characteristic of conditions before the region became cultivated. It is suggested that older loess deposits of the map area may not be related to interglacial semi-arid climatic conditions such as the present climate but to periglacial climates when extensive outwash plains void of vegetation were present. Even in the driest localities of the Three Forks basin, grasses, when not overgrazed by domestic stock, effectively retard eolian movement of soil during the present climatic conditions.

PALEOTOPOGRAPHY

Pre-Late Miocene Paleotopography

The late Miocene semi-consolidated strata rest unconformably on older, much more resistant rocks. Upon exhumation these older rocks display the rugged topography present when they were buried by onlap of the late Miocene sediments, and for most of the map area that exhumed topography is little altered by later erosion. This is due to the resistant nature of the bedrock and also to the fact that much of the exhumation has occurred since mid-or early Pleistocene. Age relationships are indicated by the position of probable mid-Pleistocene pediments with respect to paleotopographic features. From the northern to southern map boundaries the exhumed topography exhibits 2,500 feet of relief at the present time, and some of the valleys are still partly filled with the late Miocene tuffaceous sediments. By projection of paleotopographic bedrock valley slopes, it is estimated that the total pre-late Miocene relief in the map area may have exceeded 3,500 feet. However, it should be pointed out that probable northeastward tilting of the map area has also created apparent total relief not actually present during the late Miocene. It is also possible that the paleotopography exposed in the northeast part of the map area may have been several hundred feet higher relative to the Spanish Breaks and is at its present elevation due to post-late Miocene downfaulting. In general, the topography was hilly to mountainous during late Miocene and was probably very similar to the present totally exhumed topography in the southern part of the map area.

The pattern of the contact between old bedrock valley flanks and late Miocene sedimentary fill is of importance in the interpretation of Cenozoic history. The pattern in two of the partially exhumed bedrock valleys is dendritic, with the barbed pattern indicating southeast flowing streams cut the valleys in the bedrock subparallel to Laramide structural trends. However, just east of Anceney, two barbs on the north side of the old valley are reversed in orientation, possibly controlled by foliation orientation. These old valleys, now partially occupied by northwest-flowing Camp Creek (Fig. 15) and Elk Creek, join approximately four miles west of Gallatin Gateway.

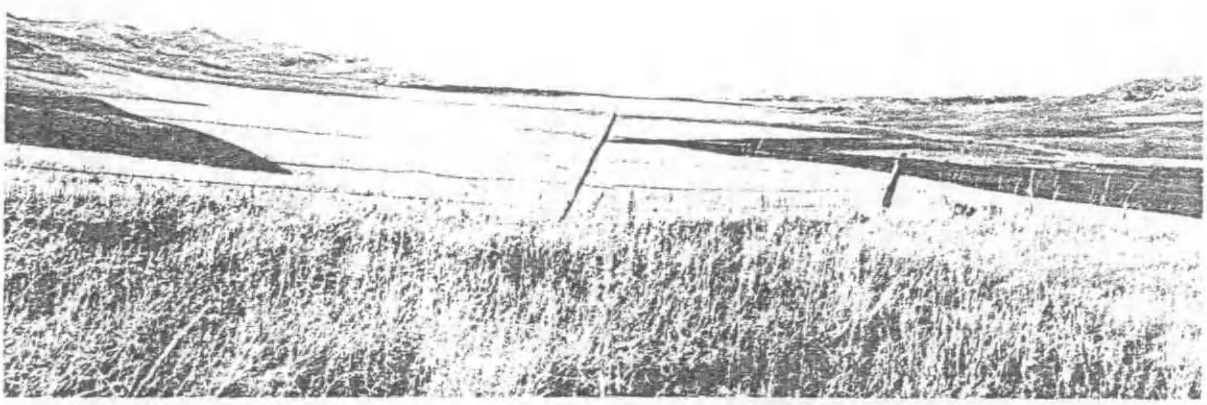


Fig. 15 View to the northwest from sec. 7, T. 3 S., R. 4 E. showing partly exhumed Camp Creek paleotopographic valley. Valley margins are Precambrian metamorphic rocks, the cultivated valley lowland is underlain by late Miocene strata. Foreground is a topographic high of late Miocene strata.

Just east of this junction is a bedrock high composed of gneiss with flanking Paleozoic strata and Tertiary volcanic rock. The pre-late

Miocene streams may have flowed either north or south around the bedrock high, or post-late Miocene fault movement may have raised the bedrock to its present position (see discussion of Camp Creek Fault). The size and length of the old southeast-flowing streams are not known with certainty, and no coarse clastic deposits clearly related to the original valley-forming streams have been found. Because detailed mapping of the basal Tertiary contact has not been done in areas to the south and east where similar contact relations exist, it is not known with certainty whether the southeastward drainage pattern is regional. Study of air photos of the area to the west toward the Madison River suggests the following: The Elk Creek paleotopographic valley probably heads a few miles west of the western boundary of the map area; therefore, it formed an eight-mile-long tributary to the Camp Creek paleotopographic valley. The Camp Creek paleotopographic valley branched at Anceney, and what appears to have been the main stream extends northward toward the center of the Gallatin Valley, where the bedrock valley flanks are now buried. The other tributary valley extends westward from the Anceney area at least as far west as the Madison River, where side valley barbs indicate drainage was to the east. From the known western extremity of this tributary to the bedrock high west of Gallatin Gateway, the total length of the paleotopographic valley is 14 miles. To the south and southeast of the map area the relationships of the paleotopographic valleys are not as well known. Spanish Creek Valley appears to head halfway between the Madison and West Gallatin Rivers, and the known distribution of Tertiary strata suggests southeastward flow. However, a similar valley heads just to the west of Spanish Creek Valley and may have graded northwest down what is now the Cherry Creek drainage. The significant relationship of the Spanish Creek paleotopographic valley is the fact that if it flowed southeast it should have run directly into the Gallatin Range front fault. The further continuation of the valley was either where the Gallatin Range is at present, or it joined a late Miocene West Gallatin River approximately in the same position as that of now. There is no known evidence for such a river in late Miocene either at the junction of Spanish Creek or in the map area to the north. On the basis of the preceding evidence, the Gallatin Range may not have been present during development of the paleotopography during pre-late Miocene time. However, other evidence presented in the section on structure strongly suggests that post-Middle Eocene Gallatin Range Front Fault displacement is not of the magnitude to have permitted pre-late or late Miocene drainage across the Gallatin Range. The displacement of the subvolcanic surface is at least 1000' short of that necessary to allow drainage southeast across the Gallatin Range, assuming all movement occurred post Miocene.

Other features of the paleotopography are important to interpretations of Cenozoic history. Ruby Mountain stands as an exhumed mountain in which the structurally downthrown Paleozoic carbonate rocks stand topographically higher than the adjacent Precambrian gneiss. This relationship suggests that during the late Miocene climatic conditions were such that carbonate lithologies were relatively more resistant to erosion than the gneiss. Therefore, the late Miocene climate may have been similar to that of the present, that is arid or semi-arid. Contrarily in other parts of the map area carbonates stand as exhumed resistant highs but generally are lower than nearby metamorphic rock.

Pre-Middle Eocene Paleotopography

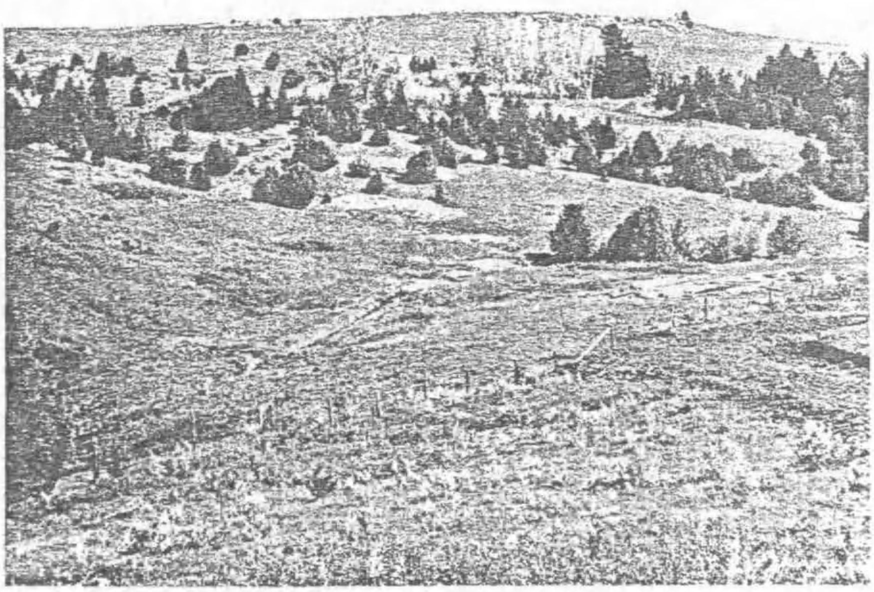
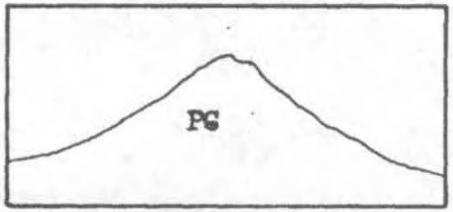
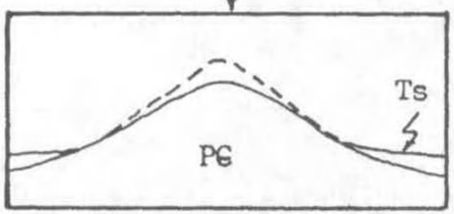


Fig. 16 Northwesterly view from sec. 27, T. 2 S., R. 4 E., of low relief developed on the metamorphic rocks of High Flat. Foreground is loess-covered late Miocene strata.

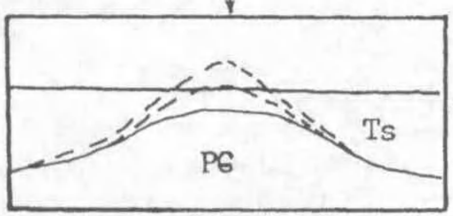
High Flat, another old paleotopographic mountain in the area, has a rather broad, low relief-summit (Fig. 16) and relatively steep flanks. Three possible explanations of its landform are presented: (1) The summit is a modified pre-middle Eocene subvolcanic surface; (2) as late Miocene onlap proceeded the summit portions were worn down due to longer exposure to erosion and then were flattened still further by pedimenta-



Miocene topographic high before sediments begin to accumulate.



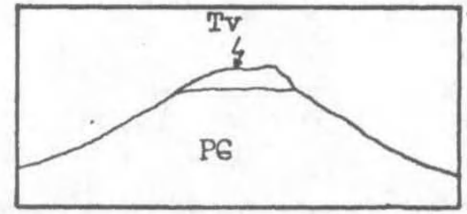
Intermediate stage of onlap, exposed portion of 'high' contributing clastics.



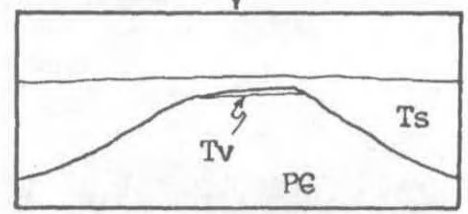
Summit considerably modified when late Miocene sediments overlap.

Tv gr.	basic volcanic gravels
Ts	late Miocene sediments
Tv	Eocene basic volcanics
PG	Precambrian metamorphics

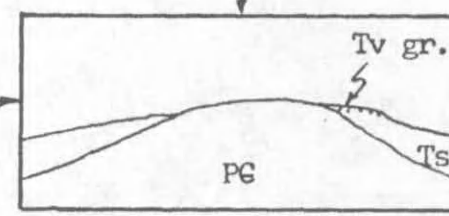
Tv gravels from Spanish Breaks area, summit beveled by pediment forming streams.



Miocene 'high' capped with basic volcanics before onlap of sediments.



Most of the basic volcanic cap eroded during late Miocene onlap.



The remainder of the Tv cap stripped during Pleistocene erosion.

Figure 17 Two interpretations of the low relief summit of High Flat, a bedrock topographic high.

tion during exhumation (Fig. 17); or (3) a combination of both (1) and (2) created the present configuration of High Flat. The writer favors the first interpretation because: Late Miocene channel deposits containing basic volcanic clasts occur to the north and west of High Flat; a very small poorly exposed remnant of basic volcanic rock crops out on the northeast edge of the sub-summit surface of High Flat; and there is a higher percentage of volcanic clasts in gravel on pediment remnants north of High Flat than in pediment gravels to the south.

Evidence in favor of (2) is as follows: Where preserved, late Miocene strata lap directly onto the Precambrian metamorphic bedrock of High Flat; pediment gravels have volcanic clasts up slope from High Flat which indicate source areas to the south; and the flanking late Miocene strata are dominantly composed of acidic ash and debris derived from metamorphic rock. At the southern and western bedrock exposures on High Flat, a siliceous arkosic conglomerate, derived entirely from metamorphic lithologies, laps directly onto bedrock. However, the evidence in favor of (2) only indicates that the basic volcanic rocks were not extensively preserved in late Miocene time and not that they were entirely gone or never present. On the basis of available evidence, hypothesis (3) is also plausible.

To the south, on the summit of the Spanish Breaks, a similar surface of low relief is present. Here pediment gravel relationships and a preserved volcanic rock remnant strongly indicate basic volcanic rocks were present during early Pleistocene or late Pliocene as a capping remnant. These relationships are dealt with in more detail in the section on pediment gravels (Qpg_2).

The regional configuration of the pre-middle Eocene topography is only partly known due to limited preservation and exposure, and as yet limited study. A northwest profile showing the low relief on the Spanish Breaks and the adjacent Gallatin Range remnant is given in Plate II, Fig. 2. Some evidence suggests High Flat had a similar low relief subvolcanic surface, but Pleistocene pedimentation may in part be the process which formed the present low relief summit area. In the Gallatin Range W. J. McMannis (personal communication, 1962) has mapped valleys in the subvolcanic surface with steep flanks having at least 3,000 feet of relief. It is the writer's belief that the pre-volcanic topography was composed of large interdrainage areas of gentle slopes and relief, with intervening, deeply incised canyons.

The distribution of preserved basic volcanic rocks in the map area and the extensive preservation in the Gallatin Range suggest that major differences between the two areas existed either during or after volcanism. Either the volcanic rocks never accumulated to the considerable thickness found in the Gallatin Range, or intensive erosion proceeded in the map area until late Miocene time, permitting drainage to be re-established subparallel to Laramide structures. This also would suggest the Gallatin Range was not, for some reason, subjected to erosion of similar intensity. Perhaps the more plausible explanation is that the volcanic rock never accumulated to as great a thickness in the map area.

STRUCTURE

Regional Structural Setting

The study area, which is located along the south central margin of intermontane Gallatin Valley, includes the Spanish Breaks, which are the northern foothills of the Madison Range (Spanish Peaks segment) and the Camp Creek Hills, which are considered a part of the Gallatin Valley (Hackett and others, 1960). The general structural features found in this area are closely related to the structure of the mountainous areas to the south, southeast, and east, including the Madison-Gallatin Range and perhaps the Beartooth Range. In these ranges Laramide deformation (Late Cretaceous to early Tertiary) has produced west-northwest and northwest-trending high angle faults which cut both basement metamorphic rocks and overlying Paleozoic and Mesozoic strata. The basement rocks are widely exposed in these mountainous areas, due to the erosional removal of the Paleozoic and Mesozoic strata during and since Laramide deformation, and remnants of these strata occur along the downthrown side of high angle reverse faults. Exceptions occur, as in the upper West Gallatin River drainage, where an extensive area is underlain by Paleozoic and Mesozoic strata preserved in a relatively downthrown block of the crust between two of the northwest-trending fault zones (Hall, 1961).

The ranges consist mainly of Precambrian metamorphic rocks or, as in the Gallatin Range, of early Tertiary volcanic rocks which rest unconformably on the metamorphic rocks and Paleozoic-Mesozoic strata. The volcanic layers are relatively undisturbed, generally having gentle dips and only minor deformation (W. J. McMannis and R. A. Chadwick, personal communication, 1963).

Master drainages flow northward from and in places through these ranges with no apparent regard for existing structures; however, many of the major tributaries are subsequent streams developed along the west-northwest structural trends. These relationships suggest that the major streams may have originated as north-flowing consequent, antecedent, or superimposed streams.

To the north and northwest of the map area, on the northern margin of the Three Forks basin (which includes the Gallatin Valley), the gross regional structural development is entirely different from that of the area to the south and southeast (Robinson, 1961; Ross, and others, 1955). Here the presence of thick Precambrian Belt strata is perhaps the reason for Laramide compressional forces forming relatively low angle thrusts and folds of north and northeast trends. The boundary between the two tectonic areas passes east-west through the Three Forks basin and is believed to be a fault zone that was active in Precambrian Belt time, permitting thick, very coarse Belt sediments to accumulate on the northern side (McMannis, 1963). North of the old east-west fault zone more extensive areas of Paleozoic and Mesozoic strata are preserved, indicating less uplift and erosion.

The only real similarity between these major tectonic elements is that of post-Laramide range front faults and associated intermontane valleys which occur in both and are characteristic of the entire northern

Rocky Mountain physiographic province (Robinson, 1961; Pardee, 1950).

The northeastern boundary of the Gallatin Valley is formed in part by the Bridger Range, which is more or less structurally unique in southwestern Montana (McMannis, 1955). This narrow north-south oriented range is mainly composed of Paleozoic and Mesozoic strata dipping to the east, and northwest-trending faults cross the range. Figure 18 illustrates the gross relationships of the Gallatin Valley with the diverse structural trends that surround it. The valley occupies the zone of intersection of these diverse structural trends and probably, in part, owes its existence to this fact.

The structure of the map area is grossly similar to that of the Madison-Gallatin and Beartooth Ranges, but has been modified by movements after Laramide deformation. The margin of the Gallatin Valley within the map area is marked by one of the major northwest-trending reverse faults, and evidence suggests the presence of another such fault buried by Tertiary sediments. Most of the Paleozoic and all of the Mesozoic strata have been stripped from the area, and the only extensive occurrence of Paleozoic strata is on the downthrown side of the Salesville fault. Precambrian metamorphic rocks form the highest hills along the southern boundary of the area (the Spanish Breaks), and remnants of Paleozoic strata form adjacent slightly lower hills.

The major difference between this area and the region to the south and east is that Tertiary sediments partially bury the Paleozoic and Precambrian rocks. The more extensive burial of the map area is related to differential structural movements, which may in part be downwarping,

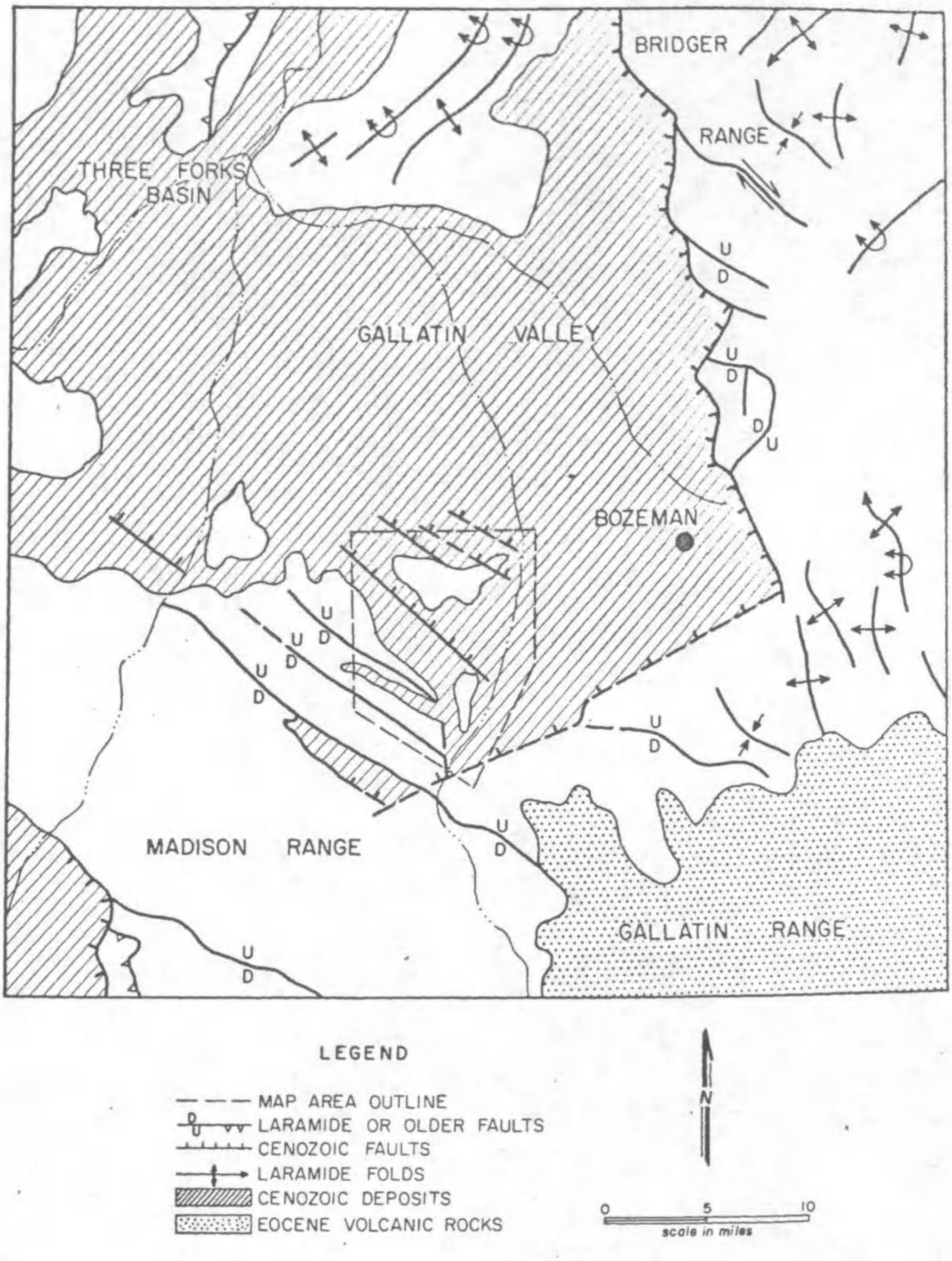

BRIDGER
RANGE
THREE FORKS
BASIN
GALLATIN VALLEY
BOZEMAN
MADISON RANGE
GALLATIN RANGE
U
D
LEGEND
MAP AREA OUTLINE
LARAMIDE OR OLDER FAULTS
CENOZOIC FAULTS
LARAMIDE FOLDS
CENOZOIC DEPOSITS
EOCENE VOLCANIC ROCKS
N
0
5
10
scale in miles

Figure 18

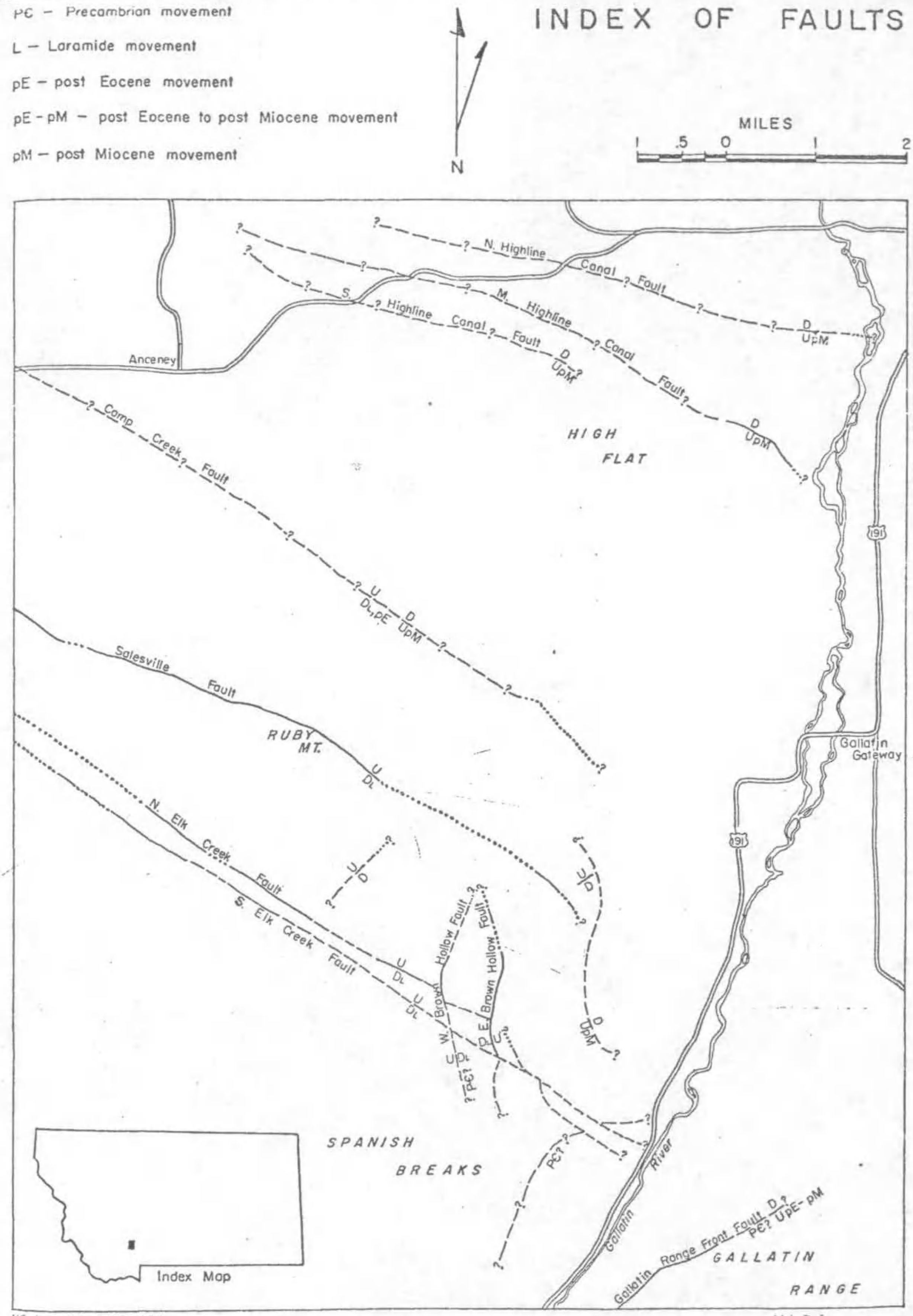
but is more likely related to downfaulting. Plate III shows the faults recognized in the map area and their relative movements at various times.

Faults

The dominant structural trend in the map area consists of west-northwest- to northwest-trending faults, that involve Paleozoic and Precambrian rocks for the most part; but some also involve the late Miocene-Pliocene strata. Similarly, the fold axes of the Paleozoic strata trend northwest.

The west-northwest and northwest regional structural grain in this part of Montana is believed by many investigators to be related to yield along weakness zones in the Precambrian basement rocks. Firm evidence of this is restricted to a few of the structures; others are inferred to be so related because of similar trends. In the map area there is no proof or even suggestive evidence for earlier Precambrian structures controlling the west-northwest and northwest faults of major displacement during Laramide deformation. Other faults are north and northeast striking, some of which appear to have been initiated before Middle Cambrian time. The evidence for the relationship is radical regional changes of foliation across the faults, changes which are geometrically incompatible with the displacement of Paleozoic strata. However, even the major movements during Laramide deformation along the northwesterly faults do not alter, except locally, the orientation of foliation on opposite sides of the faults. Therefore, it seems reasonable that large displacements could have occurred during pre-Middle Cambrian time. Acute angles

INDEX OF FAULTS



at which the Laramide faults cut the foliation are probably the reason for so little change in orientation of foliation, if it is assumed that major vertical displacements have occurred pre-Middle Cambrian and attitudes of foliation are constant for considerable depth, as well as laterally. Field evidence suggests reactivation of some faults of all three trends with reversal of movement in some cases.

The Salesville Fault: Peale (1896) and Hackett, and others (1960) recognized a fault of major displacement involving Paleozoic strata and Precambrian metamorphic rocks in the map area. Hackett traced the fault from the headwaters of Goose Creek (sec. 20, T. 3 S., R. 4 E.) to the northwest corner of Ruby Mountain. Northwest from the latter location the writer mapped the extension of the fault to Elk Creek Canyon at the margin of the map area and traced the fault by air photos and surface evidence farther northwest into Elk Creek Valley, where the trace is buried by Tertiary and Quaternary material. In the metamorphic exposures on the north side of Elk Creek Valley in sec. 26, T. 2 S., R. 3 E. (NW of the map area) sheared and solution-stained outcrops suggest possible projection of the Salesville fault even further northwest. The fault zone is covered by surficial deposits throughout the map area, but it is well defined on air photos and by stratigraphic relationships along the southeastern part of the fault. Elsewhere, it is delineated by physiographic evidence.

The trace of the Salesville fault on topography and relationships where Paleozoic strata are preserved indicates vertical to high angle reverse movement with a minimum of 1700 feet of stratigraphic displacement at the east end of Ruby Mountain. Farther southeast, where the fault is exposed again on the headwaters of Goose Creek, slightly overturned Paleozoic strata suggest high angle reverse movement along the fault, with perhaps greater minimum displacement than that at Ruby Mountain. Northwest of Ruby Mountain stratigraphic control is lost because the fault places metamorphic rocks adjacent to metamorphic rocks. This relationship and similarity in attitude of foliation on opposite sides of the fault could indicate a decrease in the amount of displacement along the northwest extension of the Salesville fault.

Attitudes of Paleozoic strata adjacent to the fault in the Ruby Mountain and Goose Creek areas suggest two possible interpretations of structural development. The Ruby Mountain Paleozoic strata dip toward the fault at an average of 35 degrees, with no exposed drag, whereas the Goose Creek Paleozoic strata are vertical and slightly overturned in reverse sequence for some distance away from the fault. The nearest exposures of Paleozoic strata south of Ruby Mountain and Goose Creek dip north-northeast at an average of 30 degrees. The gross relationships suggest either: (1) the fault diagonally cuts an asymmetrical syncline which has a northwest-trending axial plane and the overturned flank at Goose Creek may be related to drag along the fault. (2) The syncline developed with the faulting. In this situation the orientation of the syncline indicates left lateral as well as vertical displacement has occurred, and the overturned limb is a product of drag.

No conclusive evidence favoring either of the possibilities was noted in the study. However, the regional structural grain of west-northwest high angle faults found in areas south of the Belt shore line is not compatible with the gross north-south trending compressional features of the Laramide orogeny. Yet these disoriented faults and associated structures have major movement during the Laramide phase of deformation (Hall, 1961). Thom (1957) and others have called upon refracted forces to create such a divergence of the Laramide compressive movements. However, if the areas where Paleozoic and Mesozoic strata were directly underlain by the basement metamorphic complex acted as a relatively more rigid buttress than areas underlain by thick sequences of Belt strata, east-west compression of such a buttress should produce left lateral strike-slip components on northwest-trending fractures. Structures suggesting such strike-slip movement are found adjacent to the west-northwest high angle faults in the Gallatin Range (W. J. McMannis, personal communication, 1963; Hall, 1960); therefore, the writer favors an explanation including a significant component of strike-slip displacement on these northwest-trending faults.

Hackett, and others (1960, p. 53), suggest that a westward extension of the Salesville Fault displaces Tertiary sediments and a Pliocene(?) gravel-covered surface west of the Madison River in sec. 9, 10, and 14, T. 2 S., R. 1 E., and postulate renewed activity with reversal of displacement. Study of air photos of the country between the map area and the Madison River shows no features along the Salesville fault which might indicate movement involving mid-Tertiary sediments. Furthermore, the Salesville Fault would have to jog to the west from its consistant northwest trend to pass where Hackett proposes its extension. It appears more likely that other northwest-trending faults which occur south of the Salesville fault in the map area are related to the faulted erosion surface mentioned by Hackett and others (1960). These other faults will be discussed in the following paragraphs.

Elk Creek Faults: Two parallel faults which extend from near the mouth of the Gallatin Canyon to a few miles east of the Madison River are here termed the Elk Creek faults, for their proximity to Elk Creek. These faults are approximately one-fourth mile apart and trend N. 55 W. sub-parallel to the Salesville fault. Similarity of trend and displacement suggests a genetic relationship between the Elk Creek faults and the Salesville fault. Displacement on the Elk Creek faults is similar to that of the Salesville in that the southern blocks are downthrown but differ in that the apparent amount of displacement is much less.

Remnants of Flathead Quartzite are preserved on the downthrown side of the southern Elk Creek fault in sec. 32, T. 3 S., R. 4 E., and sec. 16, T. 3 S., R. 3 E. These are small isolated, poorly exposed outcrops in which attitudes of bedding are difficult to obtain. The apparent minimum displacement is approximately 700 feet, with the southern block downdropped relative to the northern block.

A minimum displacement of approximately 300 feet on the northern Elk Creek fault is suggested by the small patch of Flathead Quartzite in sec. 25, T. 3 S., R. 3 E. The southern block is downthrown relative to the northern block. It appears that the Elk Creek faults were developed by the same forces that produced the Salesville fault but perhaps were displaced less, having a gross combined minimum displacement on the two faults in the neighborhood of 1000 feet. This estimate of vertical

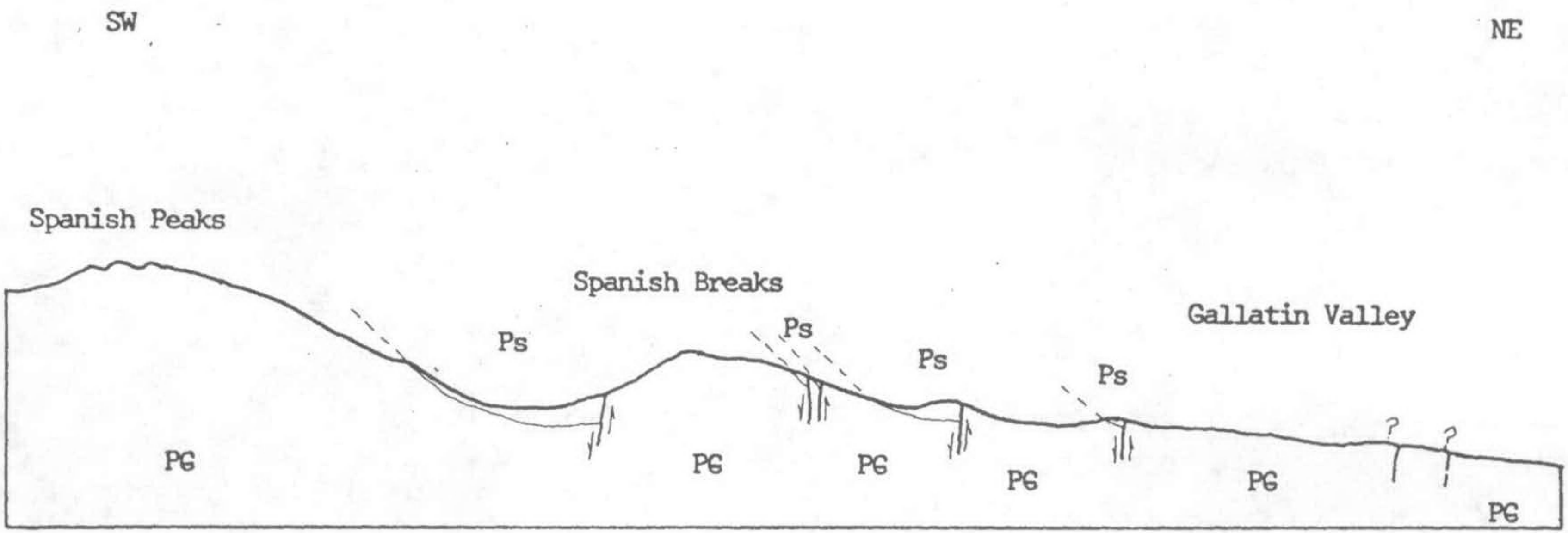


Figure 19 Diagramatic cross section illustrating tilted fault block relationships.

PG Precambrian metamorphic rocks

Ps Paleozoic sedimentary strata

displacement is similar to that arrived at when exposures of the Flathead Quartzite base on the northwest end of Ruby Mountain and in sec. 16, T. 3 S., R. 3 E., are used to determine stratigraphic displacement. The cross section , Plate II, Fig. 2 and Fig. 19, show the apparent tilted block relationships created by the Salesville and Elk Creek faults as well as similar fault blocks to the north and south.

Air photos vaguely suggest that Tertiary strata may be disturbed where the faults cross Elk Creek Valley in secs. 7 and 8, T. 3 S., R. 3 E. However, no conclusive field evidence was found to support this. Air photos show a zone of probable faults or fractures in secs. 34 and 35, T. 2 S., R. 2 E., northwest of the map area, all of which are more or less subparallel to the extended Elk Creek faults. Five miles further to the northwest along the same trend across the Madison River is the faulted Pliocene (?) surface mentioned by Hackett and others (1960). It therefore seems to the writer that the Elk Creek faults are more likely related to the fault that cuts the Pliocene (?) surface than is the Salesville fault, but this is not certain for detailed field study has not been done in the intervening zone. In the map area it is evident that the latest movements of magnitude along these faults must be pre-late Miocene and probably Laramide, because the recently exhumed metamorphic rocks show no physiographic expression of the fault. If movements had occurred along this fault after late Miocene strata buried the fault, some offset in the exhumed paleotopographic surfaces should be apparent near where the fault traces pass under the presently preserved late Miocene strata.

<u>Highline Canal Faults:</u> Across the northern margin of the map area there is evidence for three extensive faults of small displacement. The faults involve late Miocene - Pliocene (?) tuffaceous sediments and apparently also the underlying Precambrian metamorphic rocks. The name Highline Canal fault zone has been applied to the faults because the best exposure of their relationships is found along this canal.

The two southern faults are fairly well defined in the field and on aerial photos, but the northern fault is suggested only by faint air photo lineations similar to those of the two southern faults and by inconclusive field evidence. In some areas field evidence for the southern fault consists of a noticeable change in lithology in the Tertiary sediments as the fault is crossed, and the apparent control it imparts to the location of the northernmost Precambrian outcrops along the eastern extension of the fault. Field evidence for the northern fault consists of an abrupt change in foliation on the northern end of a low hill composed of gneiss in sec. 23, T. 3 S., R. 4 E., a thermal spring 50 yards to the north of the gneiss exposure in the Gallatin River alluvium, and some minor topographic irregularities developed on Tertiary strata along the suggested trace.

The middle Highline Canal fault is exposed in sec. 27, T. 2 S., R. 4 E., along a fresh ditch cut where Tertiary strata are in fault contact with Precambrian gneiss. South of the fault there is a zone 200 feet wide of solution-stained, sheared, and brecciated gneiss. There is a suggestion that the shear planes dip steeply southward, but this is not conclusively shown in the exposures. The contact of Tertiary strata with gneiss is limited to a vertical exposure approximately 5 feet high, and the fault plane dips approximately 85 degrees to the south in a reverse fault relationship. The normal depositional contact of the Tertiary sediment and gneiss is offset a maximum of about 80 feet and a minimum of about 40 feet depending on possible topographic relief on the pre-late Miocene erosional surface (Fig. 20). Along the westward extension of this fault in sec. 13, T. 3 S., R. 3 E., a late Miocene calcareous siltstone bed displays possible drag where exposed in a cut on the south side of the Anceney Road (Fig. 21). A short distance northwest of the latter location small displacement faults in a canal cut are probably part of the disturbed zone adjacent to the middle Highline Canal fault.

On the air photos faint lineations in fields of summer fallow can be seen where the Highline Canal faults are believed to lie. Of the three faults, the southern and the northern ones are the most clearly defined on the photos; the middle fault is obscure in comparison. In the fields the lineations show as discontinuous elongate light spots which occur on either side of the suggested fault trace. The light spots seem related to minor differences in lithology and apparently are not a reflection of recent movement. Other light spots not on the suggested fault traces occur where topography is favorable for erosion of the loess and soil blanket.

Projection of the three faults farther west was not possible because air photos of this area were not available to the writer. Locally there are suggestions of topographic control by the faults. Similar topographic expression suggests these faults or others of their kind may extend west

Fig. 20 View to northwest along the trace of Middle Highline Canal Fault from sec. 27, T. 2 S., R. 4 E. The break in the sky line marks the displacement of the bedrock surface. Canal cut exposure of the fault is behind the evergreen trees at the left side of the picture.

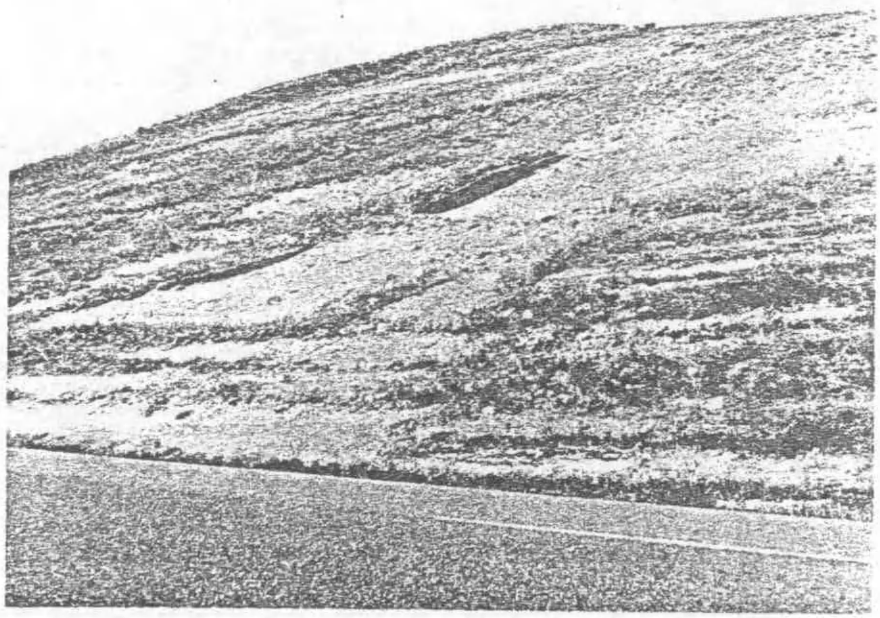


Fig. 21 Steep dip in late Miocene strata related to drag on the down-thrown side of Middle Highline Canal Fault in a highway cut in sec. 13, T. 2 S., R. 3 E. The Fault plane is not exposed in the cut, lies to the right of the picture.

of the map area. To the east the northern and middle Highline Canal faults are buried by Quaternary Gallatin River alluvium, and the southern fault dies out within poorly exposed gneiss.

At least part of the movement along these faults is pre-Gallatin River alluviation and post-late Miocene deposition; however, movements as young as latest Pleistocene do not seem probable because fault scarps are lacking and fault traces are obscure. Therefore, it is probable that the most recent movements on these faults may range from late Miocene to middle Pleistocene.

Gallatin Range Front Fault: The most important fault in the development of the Gallatin Valley is the northeast-trending Gallatin Range front fault, or fault zone named here on the basis of its geographic and physiographic position.

This fault extends from the south margin of Spanish Creek Valley (south of the map area) northeastward to the mouth of Bear Creek in sec. 26, T. 2 S., R. 6 E., a minimum distance of 23 miles, marking the common boundary of the Gallatin Valley and the Gallatin Range and the southern extent of the Tertiary strata in the Gallatin Valley. The fault cuts across the extreme southeast corner of the map area, and slightly more than a mile of it was mapped (Fig. 22). This segment of the fault and the surrounding relationships provide evidence for movement of considerable magnitude since middle Eocene time.

In the map area the Gallatin Range front fault displaces gneiss of the Spanish Breaks structural block against gneiss of the Gallatin Range

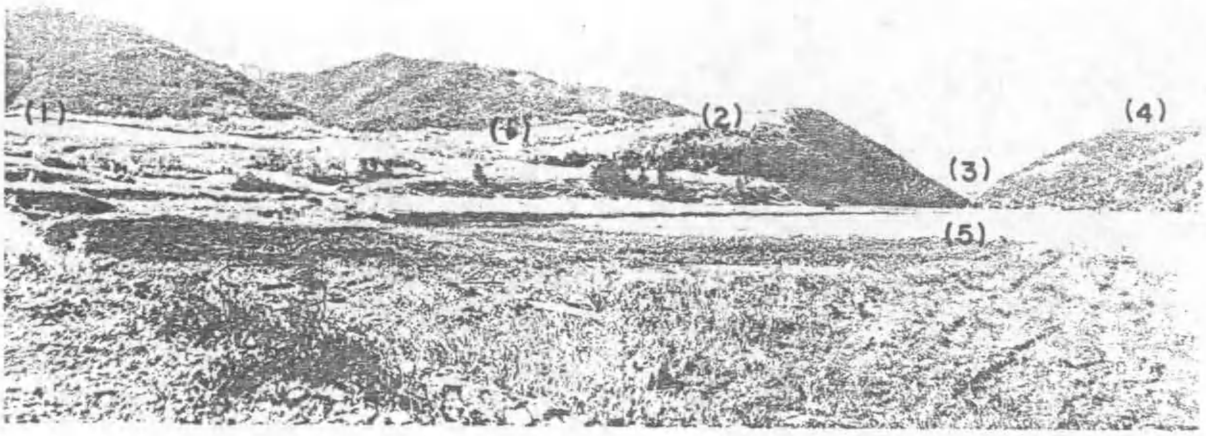


Fig. 22 View to the south from sec. 27, T. 3 S., R. 4 E., showing: (1) Gallatin Range Front Fault, (2) remnant of Eocene paleotopographic surface isolated from the Spanish Breaks portion by (3) the West Gallatin River canyon, (4) one of the high bedrock cut benches, (5) river terrace Qtg_2, and (6) basic volcanic rock remnant.

block. However, on the downthrown side the gneiss is unconformably overlain by a small remnant of basic volcanic rocks, similar to those which cap and form a major portion of the Gallatin Range (see regional tectonic map, Fig. 18). The geographic distributuion of these poorly exposed remnants strongly suggests they rest on the tilted and gently undulating surface found along the crestal portion of the Spanish Breaks. Much of this crestal surface was apparently mantled by volcanic rocks and provided the source for basic volcanic rock clasts found in the pediment gravels and other high gravel deposits of the map area. On the southeast (upthrown) side of the fault in the Gallatin Range a similar surface of low relief forms crestal portions of adjacent ridges at an average elevation 2000 feet higher than the corresponding surface on the downthrown side. Also, on the upthrown side there are more extensive remnants of the basic volcanic rock. The similarity of the two surfaces bearing basic volcanic remnants indicates that the two are parts of a once coincident gently undulating sub-volcanic surface. Plate II, Fig. 3 illustrates these relationships and indicates that the displacement of the surface by the fault in the southeast part of the map area may be between 1,500 feet and 2,500 feet, depending on the relief of the subvolcanic surface. The cross-section has been drawn so as to avoid dissected parts of the remnant surface, and therefore is not a straight line section.

The Spanish Breaks block (downthrown) and the Gallatin Range block (upthrown) of the Gallatin Range front fault have foliation trends which are essentially opposite in dip except for local reversals in the Gallatin Range block. The gross reversal of foliation is a geometric impossibility if attributed to drag along the fault which offsets the sub-volcanic surface. Elsewhere in the map area, similar change of orientation of foliation along linear trends were noted. Evidence supporting pre-Flathead development of these structures occurs on the western north-trending fault in the south-central part of the map area, where the 100 feet of stratigraphic displacement of Flathead Quartzite does not appear sufficient to produce a consistant 50 degree change of foliation across the fault. Reid (1963) reports evidence of similar Precambrian (or early Cambrian?) faulting and folding in the northern Tobacco Root Mountains to the west, which indicates this phase of structural movement may be common in many areas of the Precambrian basement rocks.

The regional relationships of the Gallatin Range front fault are important in understanding the origin of the Gallatin Valley. In the map area the pre-Cenozoic bedrock exposures disappear along a sinuous northerly trend, east of which no bedrock is known to be exposed except at the southeastern and eastern margins of the Gallatin Valley. The apparent dip of the Cenozoic basin sediments east of the older bedrock exposures is either gently eastward (Hackett, and others, 1960), as in the more central part of the valley, or southeastward nearer to the Gallatin Range front fault (W. J. McMannis and J. de la Montagne, personal communication, 1963). This suggests the probability that displacement is greater on the northeast part of the fault and that the bedrock "floor" of the basin is partially exposed in the writer's map area. The Tertiary strata that are in Spanish Creek valley are perhaps still preserved due to the fact that this area as well as the Spanish Breaks (the northern drainage divide of Spanish Creek) is structurally part of the downthrown Gallatin Valley block. The strata are not continuous with late Miocene strata of the Gallatin Valley due to the old late Miocene Spanish Breaks topographic high. Near the confluence of Spanish Creek and the West Gallatin River the west-northwest Cherry Creek-Squaw Creek fault (W. J. McMannis, unpublished map, 1962), a major fault similar to the Salesville fault both in origin and configuration, is crossed and slightly offset by the Gallatin Range front fault, suggesting the latter fault is younger and essentially uninfluenced by the earlier west-northwest trends.

Basin fill deposits of probable late Miocene age just north of the West Gallatin River Canyon are important in determining the age of movement on the range front fault. Though no direct fossil evidence was found in these beds, similarity in lithology as well as apparent contiguity with fossiliferous deposits elsewhere in the map area indicates the strata are late Miocene. These beds crop out in attitudes generally sub-parallel to the tilt of the Spanish Breaks surface (Fig. 23). Also, to the northeast, just beyond the map area, probable late Miocene beds dip gently southeastward toward the fault. The tilt of these beds indicates at least some of

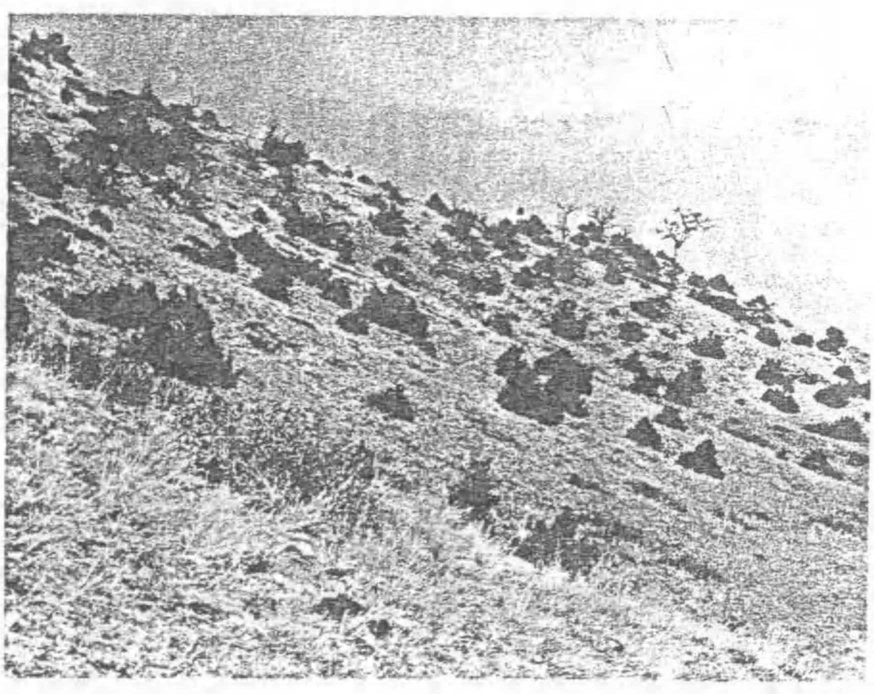


Fig. 23 Twenty-five degree dip of late Miocene strata which may be related to valley forming tectonics. In sec. 29, T. 3 S., R. 4 E.

the movement along the fault was post-late Miocene. Depending on the amount of relief developed on the sub-volcanic erosion surface, it is conservatively estimated that there has been between 1500 and 2500 feet of displacement on the Gallatin Range front fault since middle Eocene time. The amount of post-late Miocene movement is unknown in the map area but may exceed 1500 feet southeast of Bozeman (W. J. McMannis, personal communication, 1963).

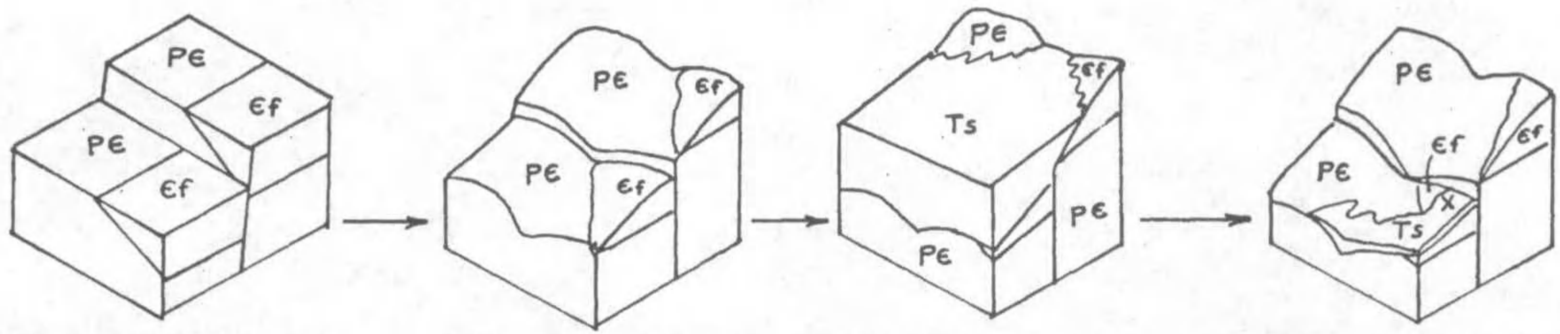
Brown Hollow Faults: The eastern of two north-trending faults that cross Brown Hollow has a well-defined fault line scarp (Fig. 24). The

Fig. 24 East Brown Hollow Fault scarp in sec. 30, T. 3 S., R. 4 E. Note the similarity of the form of the erosion surfaces above (a) and below (b) the scarp.

fault vertically displaces Flathead Quartzite 500 feet, with no evidence of strike-slip movement. Siliceous arkosic conglomerate beds of late

Miocene age are exposed on the downthrown block 40 to 50 feet from the 0 to 30 foot scarp. No debris from Flathead Quartzite occurs in the Tertiary beds, yet the beds rest nearly on or slightly above the downthrown Flathead Quartzite. On the upthrown block immediately adjacent to the arkosic exposures, no Flathead strata is preserved, and the absence of Flathead debris suggests that at the time of deposition of the arkosic beds an upslope environment similar to that of the present existed. Two possibilities for the preservation of the fault line scarp seem apparent to the writer: (1) the latest fault movements took place during late Miocene time, and the scarp reflects the fault line scarp at the time the late Miocene strata onlapped and buried the fault, or (2) the major displacement occurred before the late Miocene, when the Flathead Quartzite had been displaced and eroded from the upthrown block before the fault was buried by the late Miocene sediments; then renewed movement along the fault created the scarp. The renewed movement could have occurred after the late Miocene strata were stripped from the fault zone or before stripping. In either case the height of the scarp would reflect the amount of post-late Miocene movement because recent exhumation has not permitted much erosional modification of the resistant bedrock. Figure 25 diagramatically shows two possible sequences of events.

No definitive field evidence favoring either possibility was noted. However, similar topographic configuration on both sides of the scarp might suggest a faulted paleotopographic surface, which favors possibility (2). The importance of these relationships lies in the suggestion of possible faulting very near the time of late Miocene sediment onlap. This interpretation would indicate a close association between tectonic activity

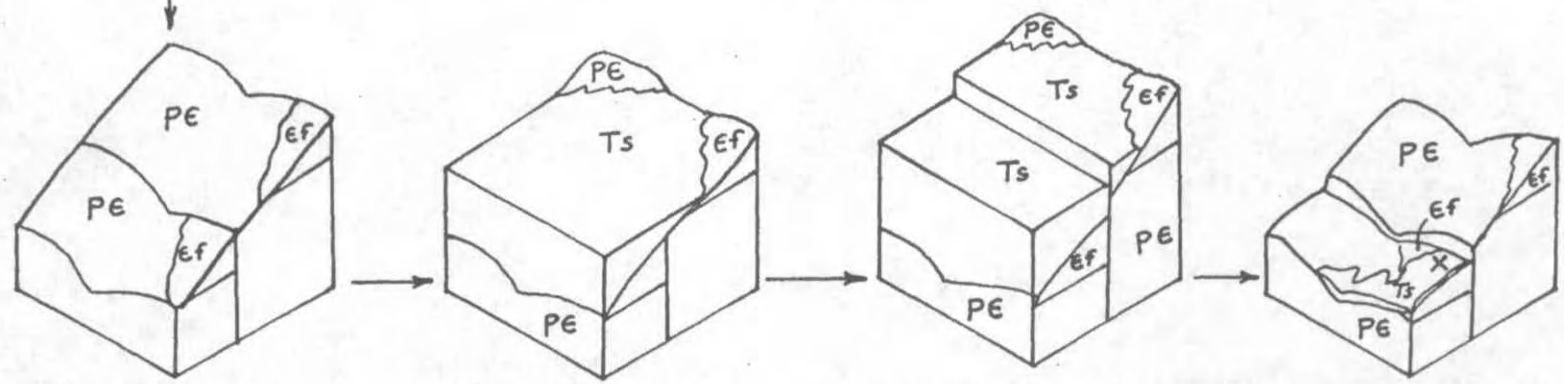


Original 500 feet of vertical displacement.

Development of fault line scarp and topography before burial in late Miocene, suggesting movement before burial.

Burial of fault line scarp in late Miocene.

Present relationships after exhumation in Pleistocene. X marks location of arkose.



Erosion has obliterated fault scarp before late Miocene.

Burial of fault in late Miocene.

Renewed fault movement post late Miocene.

Exhumation, with fault line scarp preserved due to the resistant bedrock lithologies. X marks location of arkose.

Figure 25 Two interpretations of E. Brown Hollow Fault scarp development.

and sedimentation, hence supporting the hypothesis of tectonic damming of the basins, as suggested by Robinson (1961 b).

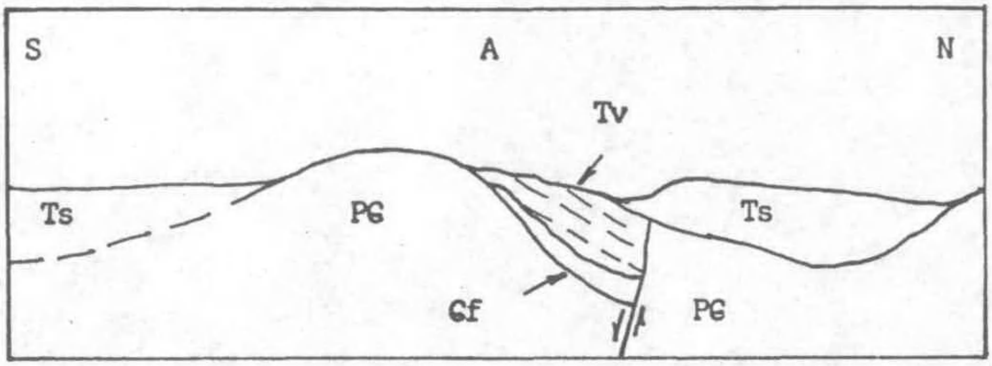
The present attitude of the late Miocene strata may also be important in the interpretation of time of faulting. Tertiary strata between the fault and the Gallatin River dip eastward at 15 to 25 degrees. No strata are exposed west of the fault in the vicinity of the fault scarp. The attitude of the strata could be related to post-late Miocene movement of the fault, but could also be related to movement on the Gallatin Range front fault, because the sub-volcanic surface of the Spanish Breaks appears to be tilted eastward into the latter fault; hence the eastward dip of the late Miocene strata may be related to the same warp or sag phenomena.

<u>Camp Creek Valley Faults:</u> Flathead Quartzite dips steeply northeast off the flank of a gneissic bedrock high three and one-half miles west of Gallatin Gateway and suggests a major high angle fault similar to the Salesville fault. This fault is now completely obscured by late Miocene and younger surficial deposits. All pre-Cenozoic bedrock exposed north of this locality consists of Precambrian metamorphic rock. On strike, one-fourth mile northwest of the Flathead Quartzite exposure, a small outcrop of gneiss underlies Cenozoic basic volcanic rock. However, the fact that the Flathead Quartzite is absent at the northern bedrock exposures does not prove the fault exists, as erosion or an eastward plunging syncline could also account for the absence. Strike of foliation in the gneiss bedrock high underlying the Flathead Quartzite gradually changes from N. 45 E., and 60° SE dip in the southern exposure, to N. 30 W.,

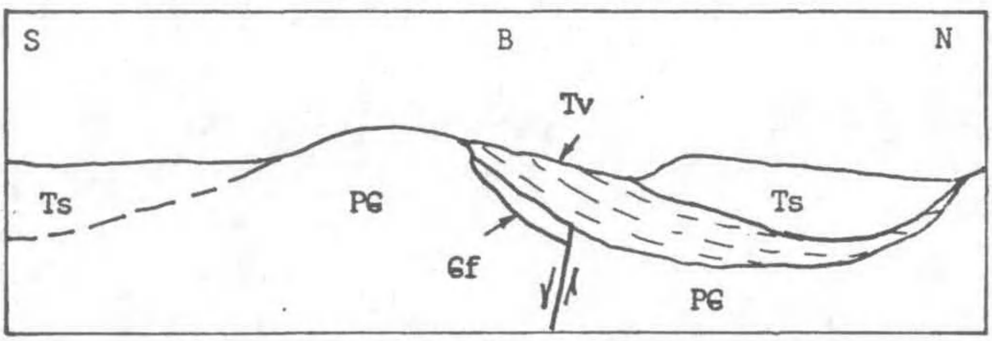
75^{o} NE on the northern exposures of the bedrock high, suggesting the change could be related to an eastward plunging anticlinal structure. This is also suggested by the distributuion and attitudes of flanking Flathead Quartzite exposures.

Adjacent to the Flathead strata, resting on the gneiss, is a remnant of basic volcanic rocks of composition similar to those of the Gallatin Range. Exposures are poor but flow layers appear to dip about 30^{o} to the north. The volcanic remnant also lends support to the hypothesis of a fault or fault zone immediately to the north of the bedrock high as no exposures of this rock type occur in the nearest bedrock exposures to the north. It is possible that the volcanic strata are also in an eastward-plunging syncline, but this is not consistant with the apparent absence of compressional tectonic features in the basic volcanic sequence where widely exposed in the Gallatin Range. Also, the attitude of the volcanic remnant is different (with less dip, different strike) than the adjacent Flathead exposure. The most probable relationship suggested by the attitudes and distribution is that renewed movement along a high angle Laramide fault trapped the basic volcanics adjacent to the fault on the downthrown side, as shown in Figure 26A. If this is so, and surrounding tectonic features of the map tend to support this hypothesis, flat-lying late Miocene strata which rest directly on the volcanic rocks indicate this renewed movement was post-middle Eocene and pre-late Miocene. An alternative is to explain the attitude of the volcanics as initial dip, which eliminates the necessity for recurrent movement.

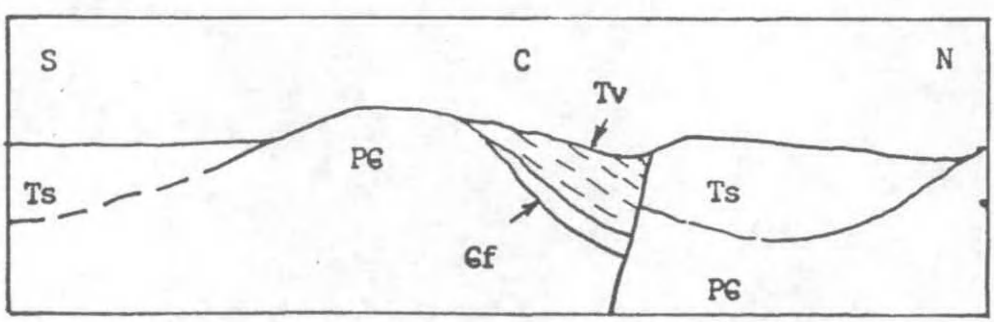
Figure 26 Possible relationships of the SE portion of Camp Creek Fault.



Laramide faulting of Paleozoic and Precambrian rocks; renewed movements post Tertiary volcanic rock accumulation; then burial by late Miocene strata.



Laramide faulting of Paleozoic and Precambrian rocks; the attitude of the Tertiary volcanic rock due to initial dip and pre-late Miocene erosional removal to the north.



The same early fault history as in situation A, with reverse fault movement post late Miocene. This interpretation is favored by the writer.

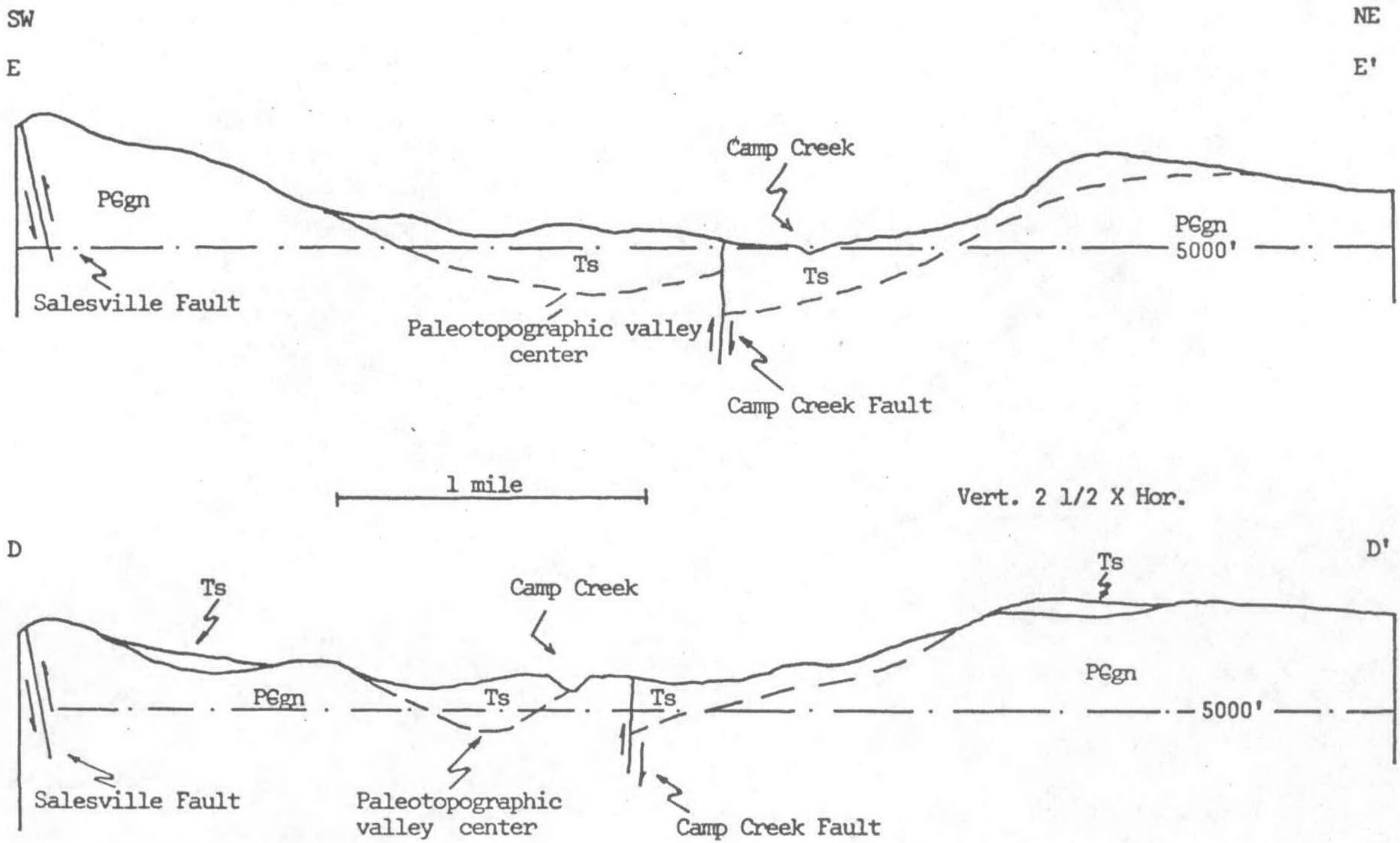


Figure 27 Relationships of Camp Creek Fault post late Miocene movement.

Northwestward on the same trend scattered outcrops of late Miocene strata display high dips and bedrock outcrops form patterns indicative of displacement of the Miocene paleotopography. The dendritic patterns of partially exhumed bedrock tributary valleys on the southwest margin of Camp Creek Valley indicate Miocene drainage was into a master stream valley that is still not exhumed. On the northeast margin of the present valley none of the tributary paleotopographic valleys are exposed due to post late Miocene displacement along Camp Creek Fault which has not permitted exhumation to a deep enough level. Two isolated knobs of bedrock in the central part of Camp Creek Valley are manifestations of the displacement of the Miocene paleotopographic surface as illustrated in Figure 29.

In summary, evidence favors recurrent movement along the Camp Creek Fault. Displacements with the southern side relatively down occurred in Laramide and perhaps post Middle Eocene time. In post late Miocene time, the sense of displacement was reversed.

Other Faults: A number of other faults and probable faults occur within the map area. In addition to the prevailing west-northwest fault trend, there are north and northeast trends. On some of these faults and probable faults changes occur in regional trends of foliation as the faults are crossed, suggesting preexisting structures in the metamorphic rocks controlled more recent movements. The basis for this suggestion is the fact that some of the faults displace younger rocks which provide stratigraphic control for the latest movements; yet movements typically are not of the magnitude to create radical alteration of the regional foliation trends.

These relationships are not true of the west-northwest-trending faults, where foliation trends commonly change slightly from one block to the next but never to reversed or essentially reversed attitudes.

Other Structural Features

In most of the map area exposures of the late Miocene deposits are widely scattered, and therefore the interrelationships of the various attitudes are difficult to interpret. North of High Flat dips are generally north or northeast at low angles. Exposures flanking High Flat on the west are varied in attitude, with dips as high as 40^{o}. Here possible explanations of the varied attitudes are compaction or tectonic movements, initial dip, and slumping, listed in the most probable order of importance. In the exposures along the topographic break between the pediments and the West Gallatin River alluvial plain dips are generally northeast to east at low angles. Therefore, grossly speaking, in the northern half of the map area a northeast tilt in the strata is apparent. Along the irrigation canals where fresh and in places fairly continuous lateral exposures of the late Miocene deposits occur, faults of small vertical displacement and gentle broad flexures are exposed. These features may be related to major faults which cut the Tertiary strata, or possibly are manifestations of differential compaction.

The structural grain of the Precambrian metamorphic rocks is generally constant throughout the map area, but deviations from the usual east-northeast strike and northwest dip occur locally along faults. Also, local variations in the foliation occur where contorted isoclinal fold crests are present, but these, where well exposed, do not greatly disturb the general trend of the foliation . It is because of the general constant trend over large areas that linear abrupt changes from area to area are believed to represent faults. However, in some cases no other supporting field evidence for such suspected faults was recognized, and therefore no fault was mapped.

In sec. 21, 22, 27, and 28, T. 2 S., R. 4 E., the metamorphic rock is anticlinally folded and slightly offset by the middle Highline fault. The axial plane of the anticline trends nearly perpendicular to the middle Highline Canal fault zone, and a well-exposed joint zone associated with rotation of foliation occurs in the saddle of Pine Butte. These features suggest the possibility of a small amount of left lateral strike slip movement on the latter fault.

Conclusions Regarding Structure

The predominating structural pattern in the map area is that of west-northwest faults of large Laramide displacement. This pattern is evident because: (1) the late Miocene strata have been eroded so that bedrock structures are partly exposed; (2) some west-northwest fractures have yielded during formation of the Gallatin Valley; and (3) relic topography which may be residual from that developed on Laramide structures is now partly exhumed.

However, the most important structural element in the configuration of the southern margin of the Gallatin Valley is the Gallatin Range front fault. The gross relationship of the fault suggests greatest displacement to the east of the map area near Bozeman, with less displacement to the southwest in Spanish Creek Valley. In the south end of the map area, the displacement is in the neighborhood of 2,000 feet, suggesting the fault movement may have downdropped and preserved a thick sequence of middle Tertiary strata in the Gallatin Valley, or created enough subsidence to permit the concurrent accumulation of the sediments. At least some movement is probably post-late Miocene, as indicated by the gentle dip of late Miocene strata toward the fault to the east of the map area, and eastward dip of similar strata just north of the sag in the pre-Middle Eocene Spanish Breaks erosion surface. Why the major valley forming faults followed a northeast trend on the south margin of the valley and northerly trends on the east margin of the valley along Bridger Range is not apparent from the study. Middle or late Cenozoic displacement along pre-existing structures is common in the map area, yet major displacement of this age along the northwesterly trends seems to be absent. This relationship may be a clue to the cause of the Cenozoic tectonic activity; however, any further statement would be presumptive.

GEOLOGIC HISTORY

In such small areas as the one studied, only a part of the total geologic history can be expected to be recorded. As the principal emphasis of this study pertains to the Cenozoic history of the Gallatin Valley, pertinent evidence from nearby areas will be called upon to aid in the interpretations of relationships within the map area.

Precambrian

Early Precambrian history in much of Montana and adjacent areas is little understood; however, it seems clear that an ancient sedimentary and igneous (?) sequence was severely deformed and metamorphosed in a number of stages. Subsequent uplift and erosion placed these rocks near the surface of the earth by the beginning of Belt deposition. Faulting along the south side of the central Montana Belt embayment at the beginning and during deposition of the coarse-grained La Hood Formation (Belt) is probably related to the uplift. The north, northeast, and perhaps northwest fault trends in the map area may also be related to this uplift, however, earlier or later phases of deformation may have initiated these structural features. All that can be said with certainty is that in the map area some deformation of the metamorphic rocks occurred after high grade metamorphism and before Middle Cambrian deposition.

Paleozoic to Early Late Cretaceous

Presence of about 7,500 feet of mainly shallow marine Paleozoic and Mesozoic deposits and absence or near absence of Ordovician, Silurian, Lower and Middle Devonian, Upper Mississippian, Permian, Triassic, and

Lower Jurassic sedimentary rock in this and adjacent areas indicates that this long interval was characterized by alternating shelf sedimentation and erosion or non-deposition. Minor non-marine deposition occurred in latest Jurassic and Early Cretaceous time. In sections near the map area representatives of this long time interval are thinner than in areas to the north and southwest, a fact consistent with a location on the northwest extremity of the Wyoming shelf (Sloss, 1950). The map area includes only the lower half of the Paleozoic sequence and none of the Mesozoic, because of pre-late Miocene erosion.

Late Cretaceous and Paleocene

Based on relationships in adjacent areas following deposition of the dominantly marine early Late Cretaceous part of the Colorado Group, the environment changed quickly and radically to one in which thick non-marine, volcanic-rich sediments were being laid down, immediately preceded by deposition of about 650 feet of non-marine, largely non-volcanic Eagle Formation. Overlying the Eagle is the Livingston Formation, which attains thicknesses as great as 14,000 feet and is characterized by coarse grain size and volcanic debris. The nearest deposits of this interval are found north and east of Bozeman along the west side of the Crazy Mountains Basin, and there is evidence suggesting that the map area may have been part of a source area supplying some of the detritus to that rapidly subsiding basin. This evidence may be summarized as follows:

(1) Cursory observation of cross-bedding in the Livingston Formation

between Bozeman and Livingston indicates a persistent northeast direction of transport, away from the vicinity of the map area (W. J. McMannis, personal communication, 1962).

(2) The Livingston Formation includes clasts of andesite and dacite porphyries,welded tuff, and sedimentary formations as old as Pennsylvanian Quadrant Quartzite in conglomerate beds equivalent to the Bearpaw Shale (Late Cretaceous) and clasts of the previous types, as well as those of older formations down to and including Precambrian metamorphic rock, occur in the Paleocene conglomerate. Notably absent from the latter conglomerate is detritus from the Late Precambrian Belt Series. The lithologies present establish a source area that was in part denuded down to Quadrant Quartzite by Bearpaw time (Late Cretaceous) and further denuded to Precambrian metamorphic rock by Paleocene time. Furthermore, this source area lacked Belt sedimentary rock. Such a source of Precambrian metamorphic debris without Belt debris could only have lay to the south of the old Belt shoreline; that is, in the region including the map area. Absence of welded tuff in potential sources of volcanic debris, except in the Elkhorn Mountains Volcanics adjacent to Boulder batholith (Klepper, and others, 1957), suggests the latter area was one major source of the volcanic detritus in the Livingston Formation.

(3) Presence of another major area of accumulation of volcanic detritus equivalent to part of the Livingston Formation lies in the upper Gallatin Basin (Hall, 1961), south of the map area, suggests that the source for the Crazy Mountains Basin Livingston Formation must have been between the two depositional centers.

In all respects, it seems very likely that the map area was part of a major uplift that contributed debris to the Livingston Formation. Such an uplift was probably related to Late Cretaceous and Paleocene movements on northwest-trending faults.

Eocene

The Late Cretaceous and Paleocene Livingston Formation was subjected to severe compressional deformation very late in Paleocene and/or early in Eocene time (McMannis, 1955). Structures produced in western Montana during this major episode of Laramide orogeny trend generally north-south, indicating an over-all east-west compressional stress. Northwest-trending fractures, such as the Salesville Fault, should, therefore, show some evidence of left-lateral, strike-slip movement. Evidence suggestive of this type of movement has been presented in the section regarding structure of the map area.

The active compressional phase of deformation had ceased by Middle Eocene time, as evidenced by lack of compressional deformation in the middle Eocene volcanic rocks of the Gallatin Range. By this time, uplift probably related to the compressional deformation, had permitted erosion to proceed to near the present structural level in the map area, as well as in areas to the west, south, and east.

In the map area, evidence strongly suggests that the Gallatin Range volcanic material accumulated to at least 1,000 feet in thickness and covered the present Spanish Breaks topographic high, probably spreading at least as far north as High Flat. In the map area no tuffaceous breccias were noted; however, in the Gallatin Range such breccias are interbedded with extrusive volcanic rock, and they thin to the northwest toward the map area (R. A. Chadwick, personal communication, 1963). This relationship suggests a source area in the northwest part of the Gallatin Range or beyond and that during volcanism such a source probably was topographically higher than the central parts of the Gallatin Range. The map area may have been part of that source, but is only one possibility and no clearly related feeder bodies are exposed. Evidence regarding termination of this phase of basic volcanic activity has not been founded in the northern Gallatin Range; however, it may correspond to termination of Early Basic Breccia extrusion in Yellowstone Park, which dates as "well into middle Eocene" (Dorf, 1960, p. 259).

Oligocene to Late Miocene

In the northwest part of Three Forks basin deposition of basin fill began at least by late Eocene time and continued on into Oligocene time (Robinson, 1961). However, in the map area paleotopographic and late Miocene strata relationships suggest there was either no deposition of Oligocene or earlier strata, or entire removal of such strata occurred before late Miocene time. Thus it appears that the "floor" of the basin in the map area was higher than other parts of the basin floor. Previous investigators have found evidence for ceasation of deposition and perhaps deformation of older fill in the Three Forks basin during the interval between Oligocene and late Miocene time. This is based on local angular unconformity between Oligocene strata and late Miocene strata and on a greater amount of deformation in the Oligocene strata (Robinson, 1960).

It has already been suggested that the map area was topographically higher than areas to the southeast during extrusion of the Gallatin Range volcanic materials. It appears quite probable that the area remained high and was subjected to erosion until accumulating fill of the Three Forks Basin buried it in late Miocene time. At that stage of development, remnants of Gallatin Range volcanic rocks still capped the Spanish Breaks surface and perhaps capped at least part of the High Flat surface.

Paleotopographic barbed valley patterns of the map area suggest that drainage, developed during the erosional period prior to late Miocene onlap, was at least in part to the southeast toward what is now the Gallatin Range and range front fault. Unless the old valleys swing northward where unexposed, the Gallatin Range at this time must have been topographically lower than the map area and thus would have been at least 5,000 feet structurally lower to permit drainage over the presently preserved rocks. This radical hypothesis is supported by only one other suggestive relationship. The volcanic rocks are extensively preserved in the Gallatin Range, whereas only ridge-capping remnants were left in the map area by late Miocene time, which, depending on the initial thickness of the volcanic rock, may indicate erosion was much more extensive in the map area. In complete discord with the latter hypothesis is the strong evidence for about 2,000 feet of displacement of the subvolcanic surface by the range front fault. It is difficult to envision a mechanism that permits southeastward drainage across the range that is still compatible with the 2,000 feet of displacement on the range front fault. Only further work can solve this problem, and the best evidence at hand favors drainage elsewhere than across what is now the Gallatin Range.

Late Miocene to Pliocene

Depending on choice of fault interpretation, the Brown Hollow and Gallatin Range front faults could have been active during this period of time. Widespread sediment accumulation occurred in the map area as well as throughout most of the Three Forks basin. Some investigators attribute this widespread deposition to eolian ash falls clogging and overloading the drainages, thus promoting deposition, whereas others attribute accumulation to tectonic damming of the basins. The characteristics of the sediments could favor either or a combination of both hypotheses. From this study it can be said that ash accumulation was not so rapid as to completely blanket paleotopographic highs, because the sediments typically are composed in part of pre-late Miocene bedrock detritus. Evidence in the area suggests that tectonic damming could have been a causal factor in that the two faults could have been active at this time and that erosion was sufficiently active to keep the higher or steeper slopes well cleaned of ash. However, no definite choice between the hypotheses can be made from the evidence at hand.

Pliocene to Early Pleistocene

Fossil vertebrate remains from Tertiary strata in the Beacon Hill area just east of Bozeman suggest the younger beds may be of Pliocene age. Bones from the map area, with one exception (the high pediment gravel equivalent), were all identified as of late Miocene age. Therefore, there is no conclusive evidence of Pliocene deposition in the map area. It is probable that most of the deformation of late Miocene strata occurred during the Pliocene to early Pleistocene interval. Movement on numerous small faults and the Highline Canal faults, as well as warping and tilting of the strata, and major fault movement on the Gallatin Range front fault all appear to have taken place during this interval.

Pardee (1950) cited the flat crestal part of the Spanish Breaks as partial evidence of a Tertiary peneplain developed during the Pliocene to early Pleistocene interval. Evidence presented here indicates it is a subvolcanic surface stripped of the last remnants of volcanic rocks during Buffalo glaciation. Contrary to Pardee's opinion, this period appears to have been one of considerable tectonic activity and erosional downcutting rather than stability.

This is also the time interval during which the first evidence of the presence of an ancestral West Gallatin River is recorded in the form of the high bedrock cut benches in the canyon area. Either by antecedance or by superposition, the West Gallatin River was able to cut the deep canyon. Antecedance is the best choice on the basis of apparent relative displacement between the Gallatin Valley block and the Gallatin Range block. However, in reality, antecedance is not satisfactory in itself because the river had to be at the Spanish Breaks summit level on Cenozoic rock to initially cut into the bedrock, as Cenozoic deposits are still preserved on either side of the Spanish Breaks bedrock at much lower elevations and are in normal depositional contact. Therefore, superposition must have played an important part in the formation of the present river canyon.

Middle and Late Pleistocene

During this interval of time, erosion must have been the dominant process. The various terrace surfaces tentatively assigned to ages as old as pre-Wisconsin indicate that erosion and exhumation of the basin sediments were proceeding at a fairly rapid rate. The possibility of continued tectonic activity is suggested by the terrace gradients and their apparent warping, as well as by the abnormally thick Quaternary alluvial deposits near Belgrade.

Summary

The present configuration of the southern margin of the Gallatin Valley is the result of fault movement, which has essentially controlled the gross distribution of late Miocene and younger basin deposits. Dating of movement on the Gallatin Range front fault and stratigraphic data are not clear enough to establish whether the fault movement was a causal factor in accumulation of late Miocene sediments. Major post-Miocene movement is suggested by the evidence at hand, and erosion of possible Miocene strata from the upthrown block could account for the present distribution. Evidence also suggests valley-forming movements have been active up to the present but were gradual and are not directly observable as they are elsewhere in adjacent intermontane valleys.

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