THE GEOLOGICAL SURVEY OF WYOMING

HORACE D. THOMAS, State Geologist

BULLETIN No. 48

STRUCTURE AND PETROLOGY OF THE NORTHERN BIG HORN MOUNTAINS, WYOMING

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STRUCTURE AND PETROLOGY OF THE NORTHERN BIG HORN MOUNTAINS, WYOMING

by

FRANK W. OSTERWALD*

INTRODUCTION

The Big Horn Mountains trend north to northwest in north-central Wyoming. The range forms a pronounced mountain barrier between the Powder River and Big Horn basins. The north end of the range rises in southern Montana, southeast of the Pryor Range, from which it is structurally

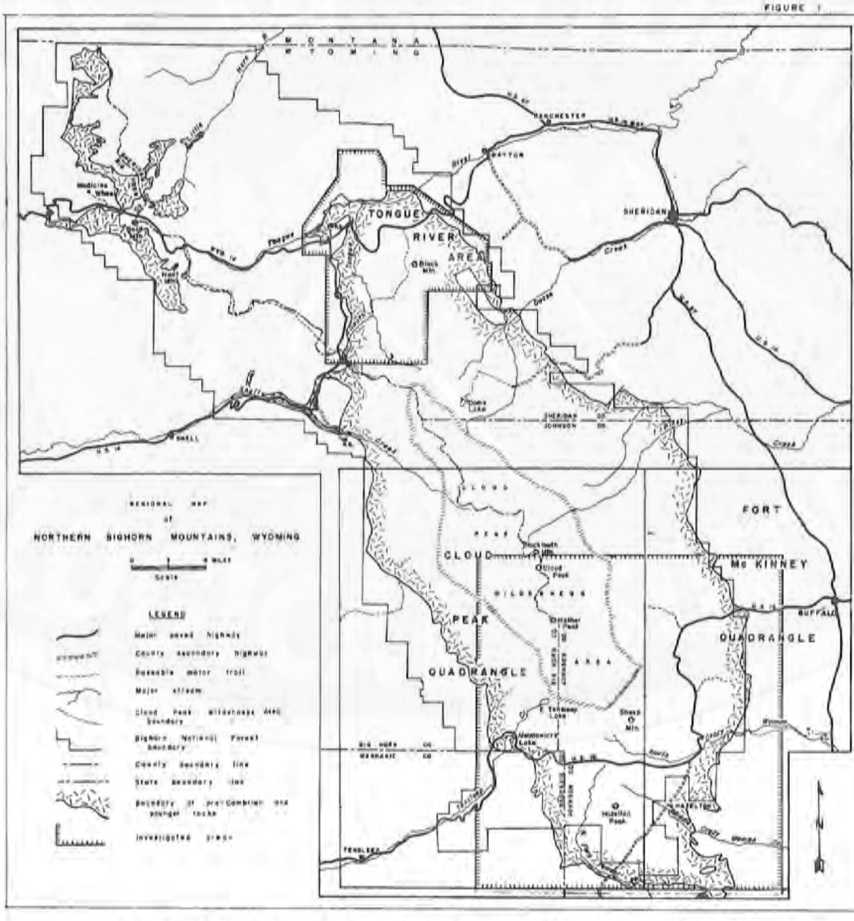

Fig. 1. Regional map of northern Big Horn Mountains, Wyoming.

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separate (Blackstone, 1940). The range is more subdued in central Wyoming, and trends west until it joins the Absaroka Mountains. The southern and western extensions are called the Bridger and Owl Creek Mountains, respectively.

The highest part of the Big Horn Range, an area about 40 miles long in a north-south direction, is west of Buffalo, Wyoming (Fig. 1), where several peaks rise above 10,000 feet elevation, and culminate in Cloud Peak (elevation 13,175 feet) and Black Tooth (elevation 13,024 feet.) The high part of the range was extensively glaciated, and has many spectacular cirques and steep-walled canyons (Mathes, 1900, pp. 173-190; Salisbury, 1906, pp. 71-91). North and south from the strongly elevated region, many gently sloping grass- and timber- covered slopes are cut by sharp canyons which extend into the range from east and west, and which are more precipitous near the margins of the range. Paleozoic and Mesozoic sediments dip away from the pre-Cambrian crystalline core on the east and west flanks (Pl. 4, A). Tertiary sediments crop out in the central portions of the Big Horn and Powder River basins, and also in small scattered areas on mountain divides.

The geology of the Big Horn Mountains was first described and mapped in detail by N. H. Darton (1906a, b, c); little detailed work has been published since. C. W. Wilson, Jr. has published excellent descriptions of the structures in the Five Springs area (1934) and of the Tensleep-Horn fault zone (1938). Structural features of the pre-Cambrian core were first studied by Ernst and Hans Cloos, who noted the parallelism of foliation planes in granitic rocks with the trend of the mountain range (Cloos and Cloos, 1934). Demorest (1941) investigated the general tectonic features of the range in his study of structural trends. Chamberlin (1940) summarized the work of many members of the Yellowstone-Bighorn Research Association who studied the region, and on the basis of their work, and that of Demorest, Thom (1947) distinguished four segments of the range:

1. Northern segment, which is bounded on the south by the Wyoming-Montana state line, and characterized by west-dipping monoclines and by east-dipping thrust planes along the western margin of the range.
2. North-central segment, which extends from the Wyoming-Montana boundary south to the latitude of Dome Lake and Shell Creek Canyon. The segment is characterized by west-dipping monoclines.
3. South-central segment, which is bounded on the south by the Tensleep Canyon-Crazy Woman cross-fault, and characterized by a gentle slope on the west side of the range and a steep slope on the eastern margin. The east margin is overthrust to the east.
4. Southern segment, which is characterized by folds with a steep western limb. The boundaries of this segment are ill defined.

The present paper is part of a study undertaken for two purposes: (1) to study the genesis of the pre-Cambrian crystalline rocks and their structural feature, and (2) to see if any relations could be found between pre-Cambrian structure and Laramide deformation. While much work remains to be done before a complete discussion of Big Horn geology is possible, it is believed the following limited objectives were attained:

1. The mapping of the crystalline core as a homogeneous granite mass (Darton, 1906a) is wholly inadequate. The pre-Cambrian rocks are now classified into at least two series.

2. Some relations between pre-Cambrian and Laramide structures are demonstrated, and possible control of later structures by pre-Cambrian structural and petrologic features is suggested.

3. Oligocene rocks crop out on mountain divides in the northern Bighorns, and the pre-Oligocene age of some marginal faults can be determined.

4. The division of the Big Horn Range into segments based on structural cross-lineaments (Chamberlin, 1940, pp. 681-684) should be modified.

5. Investigations of the structural geology of crystalline mountain cores should, wherever feasible, precede the construction of broad, regional patterns.

FIELD WORK

Field work in the region was done during the summer of 1948 and the late summer and early fall of 1949. The Tongue River granite area was studied first, as it was believed that the contact of "red granite" and "gray granite" shown on Darton's maps (1906b) would yield significant data. The granite area was mapped in detail (Osterwald, 1949, 1955) but no definite contact found; the two types of rock are intimately intermingled. As a result the study was extended to include a regional investigation of the southern part of the crystalline area (Pl. 1), where an entirely different series of gneissic rocks crops out.

Clean, unweathered outcrops are not common in visited localities. Beds of Eocene gravel abut the eastern flanks of the range, and remnants are found at elevations up to 8,000 feet in the Cloud Peak and Fort McKinney Quadrangles (Brown, 1948, p. 1165; Sharp, 1948, pp. 1-15; Darton, 1906c). Above 8,000 feet glacial till is widespread, and few bedrock outcrops are found between 8,000 feet and 10,000 feet. Below 8,000 feet the outcrops are weathered deeply and are heavily lichen covered, and above 9,500 feet large areas are covered with *felsenmeer*. Travel at higher elevations is well described by Mathes (1900, p. 182).

> All the high peaks and slopes, which on the map appear so strangely smooth of outline, possess an extremely rough surface in detail, toilsome to climb. Rapid weathering at the joint cracks has loosened vast numbers of angular blocks of granite of all sizes. Wherever the slope is at all steep fine material is not retained, and as the climber lifts himself from one block to another the sound of trickling water reaches him from the depth of the gaping holes under his feet.

Structural data may be obtained in such an area only with great difficulty; it is fit only for a cursory observation of the rock types. Excellent exposures are probably present in many cirque walls, but these could be examined only by trained and well-equipped mountain climbers.

ROCKS

PRE-CAMBRIAN COMPLEX

The pre-Cambrian rocks of the Big Horn Mountains are divided into two series. The older, an interbedded sequence of black to gray gneisses, amphibolites, chlorite-biotite-hornblende schist, talc-chlorite-actinolite schist, and other rock types, suggests metamorphosed crystal tuffs and metagglomerates. The younger series varies from granite to quartz diorite, and consists of interlayered red and gray rocks. The granitic rocks have been discussed in other papers (Osterwald, 1949, 1955) and will be mentioned briefly here.

Gneisses predominate in the higher central portion of the range, while granitic rocks are more common in the northern and southern ends of the pre-Cambrian area. Darton (1906a, pp. 22-23) first described the amphibolites and various other rocks in the central Bighorns, and some of them are shown on his maps.

The northern contact of gneiss and granite is gradational, and forms an east-west zone about 10 miles wide near the latitude of Dome Lake (Fig. 1). The southern contact (Pl. 1) is generally sharp, though interlayered and gradational at certain places.

Rocks of both series are cut by numerous diabase dikes of many lengths, and at least some of them are older than the granitic rocks. Nearly all diabase dikes dip steeply to vertical, and many strike nearly east-west (Pl. 2) (Darton, 1906a, p. 19).

Granitic rocks are sheared, silicified, and epidotized at many places.

Gneisses

The predominant gneisses are gray fine- to coarse-grained rocks with lenses and layers of amphibolite and schistose rocks. They present a wide variety of mineralogical, structural, and textural types, and each outcrop exhibits a number of unique features. A single outcrop may show dark, almost black lenses or fragments of amphibolite surrounded by dark gray irregularly shaped biotite-rich zones which are in turn enclosed by more siliceous gneissic material; a few hundred feet away from this outcrop there may be only uniform gneissic rock. The variable textures are frequently associated with gneiss containing abundant feldspar, and unusual rocks are the rule rather than the exception. These extremely variable rocks are well exposed in many places along U. S. Highway No. 16. The rocks, however, are remarkably uniform in thin section; fragmental fabric is widely distributed and mineral composition is rather uniform, though the amounts and grain size of individual minerals vary.

Gneiss generally consists of approximately 20 percent biotite and hornblende; the remainder is white feldspar and quartz. Many rocks contain augen of pink or white feldspar up to two inches long, which are most common in localities showing particularly strong deformation. Closely adjacent augen commonly form veinlets or pegmatite-like masses (herein called *augen-pegmatites*), and may make up a large percentage of the rock. The gneiss contains much dark, well-foliated hornblende rock, with interlayered coarse white feldspathic material at locality 1-1.* The gneiss at this locality contains log-like masses of coarse gray and pink pegmatitic material, which range up to 1½ feet in diameter and a few feet long. These masses are elongated parallel to the lineation of the surrounding rock, and contain streamers of coarse biotite crystals parallel to their length. Many such masses were observed in the crests of folds.

The gneissic rocks contain more pink feldspar augen, pegmatitic material, and narrow dikes of pink feldspar near the northern and southern contacts with granitic rocks. Pink feldspar also becomes more prominent along the western margin of the pre-Cambrian area, between Meadowlark Lake and West Tensleep Lake (Fig. 1).

*References to localities are keyed by number, thus 1-6 refers to Plate 1, locality 6 indicated on that plate. Descriptions of localities are given in Table 2, Appendix.

Interlayered medium-grained gray gneiss and coarse-grained black amphibolite at locality 1-1 contain many pink and white feldspar augen. Thinly laminated even-textured biotite gneiss contains feldspar augen up to one inch in diameter, lenses of amphibolite, and irregularly shaped masses of granite up to two feet in diameter. Crenulated biotite streaks in granite, lineation in gneiss, and lineation in amphibolite are parallel. The gray gneiss is interlayered with a fine-grained pink to gray siliceous gneiss containing dark hornblende-rich lenses. The layers of siliceous gneiss contain much epidote and small crystals of hornblende.

The gneisses at locality 1-2 contain quartz-feldspar rich masses several feet in diameter, which resemble white aplites. Within the masses, quartz crystals approximately 3/8 inches long are elongated parallel to the regional lineation. A hornblende-plagioclase rock, interlayered with the aplitic variety, has hornblende needles one inch long elongated parallel to the lineation in gneiss.

Mixtures of coarse- and fine-grained material are common in gneisses. Compositional layering is common; some layers consist entirely of hornblende, others are largely feldspar. Small insets of white feldspar, as well as smaller ones of biotite and hornblende are found at many localities. Locally large crystals of feldspar and ferromagnesian minerals are found in the same rock, as at localities 1-3, 1-4, and 1-5. These rocks contain odd-shaped sub-angular crystals of plagioclase up to one-half of an inch in diameter, and crystals of biotite and hornblende up to three-eighths of an inch long, embedded in a dark fine-grained groundmass. The outcrops at locality 1-3 also contain many odd, ellipsoidal masses of light-colored rock up to one foot in diameter (Pl. 4, B; Pl. 5, A,B). The masses commonly contain thin layers of green amphibolite less than one-eighth of an inch thick arranged in shells parallel to the ellipsoidal surface. These rocks strongly resemble metamorphosed pyroclastics.

The gneissic rocks show widespread linear mineral parallelism (lineation), which is usually marked by elongated aggregates of dark minerals (biotite and hornblende). Feldspar augen are also aligned parallel to lineation. The lineation is commonly very strong, and causes the gneisses to weather into characteristic shapes.

Locally gneisses show good foliation (s_1), though less pronounced than lineation. Foliated rocks are particularly abundant near Powder River Pass, and at this locality amphibolite layers are more common than elsewhere. Isoclinally folded gneiss at locality 1-6 contains individual folds with an amplitude up to three feet from limb to limb. Foliation has a variable strike, and is not consistent even in closely adjacent outcrops (Pl. 1).

Axial plane cleavage (s_2) in gneiss is best developed along a zone which extends from 1-5 to 1-7, and at 1-1 an isoclinal fold, approximately six feet from limb to limb, is sheared parallel to the axial plane. Oriented thin-sections from this fold show biotite plates which have grown on the shear planes; cleavage of the plates is aligned parallel to the megascopically visible fracture planes and to microscopic zones of cataclastic debris within the slides. A less pronounced preferred orientation of biotite cleavage makes an angle of about 35° with the megascopically visible fractures. In some oriented sections, compositional layering of the rock is crossed at a high angle by the biotite plates.

All thin-sections of pre-Cambrian rocks from the central Big Horn area show strong fragmental texture. Large elliptical anhedra of plagioclase

(An_{12-25}) and of microcline (<10 mm) are set in a matrix of quartz and feldspar fragments (<0.1 mm). In a few slides, many small fragments of quartz and feldspar surround an elliptical microcline augen, and the outlines of many augen are indented by the fragments. Some of the feldspars contain a few pieces of quartz. S-planes in most rocks are plainly marked by layers of fine-grained fragments of quartz and feldspar. Streamers of elliptical recrystallized quartz grains are parallel to the s-planes. Small plates and shreds of biotite (pleochroism X = yellow, $Z = Y$ = dark brown, $\beta = 1.619$ to 1.636) and chlorite are scattered through the fragmental portions of the rock. Biotite is associated, in some slides, with minute rounded to subhedral garnets and apatite crystals. Occasionally, these minerals are associated with small (<1 mm) anhedral grains of pistacite or of clinozoisite.

Anhedral crystals of hornblende (<1.5 mm in diameter) display very irregular outlines, are embayed by quartz and feldspar, and enclose numerous small, rounded inclusions of quartz. Some fragments of hornblende are drawn out from a crystal, into a crude row parallel to s_1 of the rock (Pl. 6, A). Hornblende has the following optical properties: X=pale brown, Y = greenish, Z = dark green: α 1.656 — 1.671, γ = 1.672 — 1.683; $Z \wedge c = 6°$.

Rounded or elliptical anhedra of cordierite are most abundant in rocks from locality 1-2. The mineral attains about one millimeter maximum diameter, and exhibits numerous spindle-shaped twin units, set off from each other by irregular borders. Many cordierite grains show incipient alteration to sericite and talc, and most crystals are rimmed by a thin film of limonite.

Elongated, anhedral, irregular perovskite (?) crystals were noted in several thin sections. The mineral is pale brown, has high relief, and is strongly birefringent. Cleavage is not apparent, but cross-fractures are common.

Quite commonly large aggregate masses of quartz crystals with an extremely irregular shape follow strongly fragmental zones, and most are elongated parallel to s_1. Individual crystals of quartz embay plagioclase, hornblende and biotite. Undulatory extinction is not pronounced, and adjacent crystals within an aggregate have interlocking boundaries. The quartz appears to have recrystallized at a relatively late stage.

Microcline crystals (<4 mm) pervade many fragmental zones; most are molded against other minerals, and many contain small inclusions of plagioclase and quartz debris. Many plagioclase inclusions within microcline enclose myrmekitic quartz at their margins; other plagioclase inclusions are zoned and have a rim more calcic than the core. A few plagioclase crystals within strongly fragmental zones are rimmed by narrow, interrupted shells of microcline.

Irregular anhedra of muscovite are common within fragmental areas. Occasional muscovite plates are rare within plagioclase, and are more rare in potash feldspar; sometimes several fragments of muscovite are optically continuous.

Anhedra of chondrodite (<1.5 mm) in a slide from locality 1-3 are embayed by quartz, and contain inclusions of that mineral. Many chondrolite crystals are fractured; the fractures are filled by crudely vermicular quartz.

A few slides show narrow veinlets of subhedral pistacite or clinozoisite crystals about 0.5 millimeters wide or less. Crystals are elongated at an angle to the trend of the veinlets, and a few sheaf-like aggregates of epidote minerals are within the veinlets. A thin-section of coarse-grained red gneiss from locality 1-9 shows several minute fractures filled with calcite.

A thin-section of dark-colored metamorphosed crystal tuff from locality 1-3 (page 9), exhibits good fragmental fabric and shows many subrounded fragments of quartz and plagioclase up to several millimeters in diamenter. Most of the quartz fragments are rimmed by biotite plates (pleochroism: X = yellow, Y = Z = black) and small crystals of quartz; plates of biotite are bent around the quartz fragments. Hornblende crystals (X = yellow, Y = green, Z = blue-green) have very irregular outlines, and the mineral shows sieve structure with inclusions of quartz and more rarely of plagioclase. Subhedral prisms of faintly pleochroic actinolite (colorless to pale green) up to 1 millimeter long form parallel overgrowths on hornblende. The matrix of the rock contains a mosaic of quartz, plagioclase, and microcline, with a lesser amount of biotite and an occasional small rounded crystal of garnet or of perovskite (?). A few muscovite plates (pleochroism X = pale yellow, Y = Z = colorless) are scattered throughout the matrix.

A thin-section of the light-colored, fine-grained rock from locality 1-3 consists mostly of odd-shaped crystals of unstrained quartz. Adjacent quartz grains usually have interlocking boundaries, and form a mosaic. Irregular-shaped plagioclase crystals about 1 x 2 millimeters in size are scattered through the fine-grained matrix, together with minute anhedral crystals of microcline. One large plagioclase crystal is rimmed by minute crystals of microcline; it also contains many minute anhedral inclusions of potash feldspar in parallel optical orientation. Quartz commonly is associated with feldspars in a crude coarse- to fine-textured micrographic intergrowth. Many grains of plagioclase show no twinning and enclose felted masses of fine-grained sericite, zeolites, and other minerals, which entirely cover some grains. Biotite laths, some of which are subhedral, are relatively long (<0.2 mm) compared to their width, and have a strong parallel alignment. A few grains of garnet and perovskite (?) are associated with biotite.

ORIGINAL ROCKS

The pre-Cambrian rocks of the Big Horn area were strongly deformed and metamorphosed to varying degrees so that their original character cannot be ascertained. The outcrops of gneiss with rounded masses of biotite and feldspathic material that resemble volcanic agglomerates when seen in the field (Pl. 4, B; Pl. 5 B), and the strong fragmental fabric of many thin-sections, suggests that many of the original rocks were pyroclastics. At least a part of the fragmental fabric is due to later cataclasis, but it is difficult to distinguish accurately between pyroclastic and cataclastic textures (Marmo, 1949, pp. 48-53).

Description of the Ammonoosuc series in New Hampshire and Vermont closely resembles that of many Big Horn rocks (Billings, 1937, pp. 863-936; Chapman, 1948, pp. 1070-1072; Kruger, 1946, pp. 169-172 *et al.*); these rocks are believed to be pyroclastic and volcanic. Reiche (1949, pp. 1186-1192) described pre-Cambrian tuffs and rhyolites in New Mexico which resemble some of the Big Horn rocks, though of a more acidic composition. Descriptions of the Swedish leptites, which are volcanic and pyroclastic,

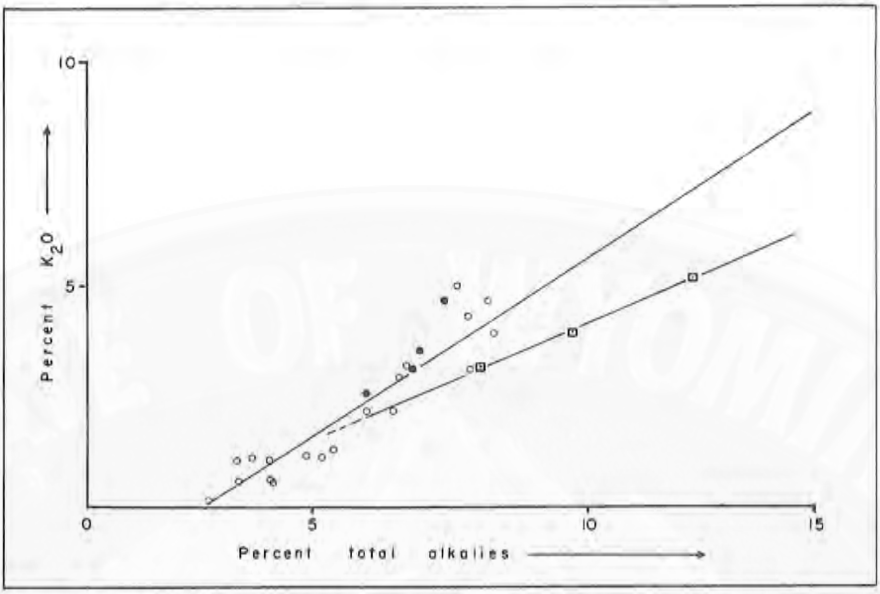


Fig. 2. Total alkali content of Big Horn Mountain rocks, plotted against potash content, and compared with analyzed rocks from orogenic regions (open circles) and with rocks from Otago, New Zealand, a strongly deformed subprovince (squares) (Turner and Verhoogen, 1951, p. 150).

closely resemble those of the Big Horn rocks (Hjelmquist, 1938, pp. 25-28; Magnusson, 1938, pp. 7-13).

The rocks near Powder River Pass probably contain more sedimentary material than those elsewhere, as they are more evenly layered and are more uniform in composition. However, the basic mineralogy as well as the fragmental fabric of these rocks suggest that they are probably pyroclastic. In general, the gneisses show characteristics similar to both the eruptive gneisses and sedimentary gneisses, according to a table by Schwenkel (1912, pp. 16-17), which is easily explained if they had an original pyroclastic origin.

Figure 2 represents the total alkali content of the Big Horn rocks plotted against potash alone. Three of the rocks fall within a group of points representing published analyses of volcanic rocks of orogenic regions (Turner and Verhoogen, 1951, p. 211); the fourth is well above the curve, probably because it was subjected to potash metasomatism. The graph strongly suggests that the original rocks were volcanics formed in an orogenic belt.

The plot of total alkalies against potash was suggested to the writer by B. D. Carey (personal communication). According to Carey the plot has many advantages over a typical silica variation diagram, because rocks belonging to petrographic suites plot as nearly smooth curves which can be represented by a mathematical equation.

DIKES IN GNEISSES

White feldspar-rich dikes: An intricate network of white feldspathic dikes, which vary from thin films up to four inches thick, commonly traverse the gneiss. The white dikes stand at every conceivable attitude, and are either ptygmatically folded, slightly bent, or straight. When ptygmatic and straight dikes of this type are found in the same outcrop, ptygmatic dikes are cut and sometimes offset by straight ones. Contacts of dikes with gneiss are sharp, but not smooth, as they are deflected around the boundaries of individual crystals. Fine-grained gray gneiss at locality 1-10 is cut by numerous narrow white ptymatically folded dikes, and the axes of ptygmatic folds are parallel to lineation in the gneiss.

Thin-sections of dike rocks closely resemble those from gneiss, although the grain sizes and amounts of individual minerals differ in the two rock types. Less biotite, chlorite, and garnet, and a proportionately greater amount of feldspars and quartz are found in the dike rocks. Mortar structure and recrystallized quartz are common, and many large crystals of feldspar are surrounded by a matrix of dust-like debris. These dikes were probably formed by repeated recrystallization during the metamorphism of the gneissic series.

Dark-gray coarse-grained dikes: A few gray coarse-grained, feldspar-rich dikes that resemble anorthosite cut the gneiss. Rounded or elliptical plagioclase crystals (An_{21}), collected from some of these dikes, have a dark bluish-gray core and lighter colored rim, and average approximately $\frac{3}{4}$ by $\frac{1}{2}$ inch in size; they are surrounded by a small amount of fine-grained matrix of quartz and plagioclase. Occasionally the plagioclase grains have bent twin lamellae, and the margins are embayed by quartz. Some areas of the matrix show masses of recrystallized quartz. The walls of many dikes are unmatched. The dikes appear to have been granulated. This process was accompanied by recrystallization of plagioclase and, to a lesser degree, of quartz.

Gray fine-grained dikes: Gray fine-grained dikes cut the gneiss at many outcrops; a typical outcrop is at locality 1-11. Here the dikes contain about 15 percent fine-grained biotite arranged in streaks parallel to the lineation in the gneiss, though most of the dikes are foliated parallel to their walls. Many have smoothly curved outlines. A thin-section of a dike at locality 1-11 consists of 45 percent biotite (pleochroism X = yellow, Y = Z = dark green). The rock has an average grain size of 0.4 millimeter, and shows well-developed s-planes, fragmental texture, and recrystallized quartz.

Red aplitic dikes: Lineated dikes of red aplitic material are also present; a good outcrop is at locality 1-12. They contain minute biotite plates arranged in rows parallel to the lineation of the enclosing gneiss. Approximately 50 percent of the rock consists of irregular anhedral grains of quartz, plagioclase, and potash feldspar. Most crystals are less than 0.5 millimeter in diameter, and have very intricately curved boundaries. Large elongated aggregates of unstrained quartz crystals up to 1.0 millimeter wide and 3.0 millimeters long are arranged parallel to the s-planes of the dike, are molded against, and contain inclusions of other minerals.

AUGEN AND AUGEN PEGMATITES

The term "augen pegmatites" as here used refers to coarse feldspathic masses with numerous lenticular augen of potash feldspar microperthite and, more rarely, plagioclase (An_{5-12}), with or without biotite sheaths. Irregular masses, layers, and dikes of coarse pink and white augen pegmatite with up to 25 percent biotite are common in gneiss, and are typically shown at locality 1-13. At this point gray coarse gneiss encloses ellipsoidal white feldspar crystals up to 2 by 3 inches, measured on cross-joint surfaces. Large feldspar crystals and masses of augen pegmatite are enlongated in the direction of lineation of the enclosing rock.

Occasionally, as at 1-14, coarse feldspathic gneiss is interlayered with masses of augen pegmatite, and locally contains as much as 70 percent feldspar. Closely adjacent augen in these rocks form narrow feldspathic veinlets less than ½ inch thick, with extremely undulatory boundaries, parallel to the foliation of the rock. Individual augen in such a veinlet are commonly elongated parallel to gneissic lineation, and may lie at an angle to the trend of the veinlet. Locally elliptical pegmatitic masses up to three feet long within the vein are elongated parallel to the gneissic lineation.

Gray fine- to medium-grained biotite gneiss at location 1-14 contains interlayered red feldspathic material with individual augen of feldspar up to six inches long. Layers of gneiss fray out into pegmatitic material, both up and down the dip of s_1, though many contacts between gneiss and pegmatite are serrate and appear to be at a very slight angle to s_1. Biotite crystals are bent around clots of feldspar in small veinlets. These relationships are sketched in figure 3.

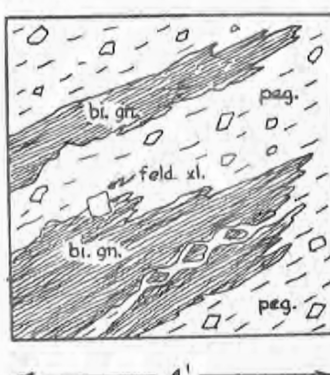


Fig. 3. Feldspathic material (augen pegmatite) replacing biotite gneiss at locality 1-14.

Augen pegmatites commonly fill the crests of small folds, and are elongated parallel to the fold axes (Pl. 6, B) or in elongated zones up to three feet wide. Some pegmatites are fractured parallel to their walls. Most contacts of pegmatite with other rocks are very irregular, with embayed and unmatched walls.

A three-foot dike at locality 1-11 exhibits numerous coarse feldspars at the margin which are similar to those in augen pegmatites. Lineation and s-planes of the gneiss adjacent to the dike are also in the coarse material. These features suggest that the dike is an early one which is partially recrystallized.

Thin-sections of augen pegmatites are similar to those of the surrounding gneiss, but have proportionately greater amounts of feldspars. Pegmatites and individual feldspar augen are composed dominantly of potash feldspar microperthite in elliptical anhedra up to several inches in size. Large crystals of potash feldspar show good grid twinning and fine-textured string and film perthites of albite. Rarely potash peldspar shows strain shadows, and some contains anastomosing networks of muscovite.

The augen, augen veinlets, and augen pegmatites which cut the southern gneisses are localized along fractures, shear zones, and within the crests of small folds. This suggests that minute fractures and openings within the gneiss aided in localizing the growing feldspars which make up the augen masses. The localization may have taken place in three ways: (1) by diffusion of cations in the solid state to places of low mechanical pressures, as outlined by Bugge (1946, p. 24) and Ramberg (1944), (2) by circulating solutions bearing granitic material along fractures, and (3) by injection of fluid magmatic material.

The injection of fluid magma is unlikely, because the outline of the masses is highly suggestive of a replacement origin (Osterwald, 1955). For this reason the augen probably were emplaced by one or both of the other two methods, though it is unlikely that an actual mechanical vacuum (as postulated by Bugge and Ramberg) could exist at the depths in the earth at which the Big Horn rocks formed. The minute fractures probably served as pathways for circulating fluids or gasses, through which the soda, silica, and alumina necessary to form the augen could move. Diffusion through gasses or fluids is much easier to visualize than purely solid diffusion. Diffusion in the solid state probably aided the process by moving material laterally, away from the fractures.

The granitic dikes and veins, mentioned above, probably had a similar origin, as many of them appear to consist of recrystallized closely adjacent augen.

LESS ABUNDANT ROCK TYPES

Interlayered with gneiss are a variety of other rock types that form lenses up to a few tens of feet thick; they are most common near Powder River Pass (Fig. 1; Pl. 1).

Amphibolite: Hornblende amphibolite is the most abundant of the lesser rock types. A belt of this rock 3,000 feet wide extends for about five miles from the Middle Fork of Crazy Woman Creek to Canyon Creek (Pl. 1) (Darton, 1906a, pp. 22-23). The rock is greenish black, and coarse grained, with strongly aligned hornblende crystals (<10 mm), and varying amounts of plagioclase (An_{17}). In thin-section the rock shows good to excellent mineral parallelism. Anhedral hornblende crystals (pleochroism $X =$ yellow-green, $Y =$ green, $Z =$ blue-green; $\alpha = 1.625$, $\gamma = 1.668$) have pronounced embayments of quartz and show good twinning. About 10 percent of the rock is made up of minute opaque minerals. Grains of quartz and plagioclase are usually small, and form a mosaic between crystals of hornblende, but mortar structure is not evident.

Actinolite-plagioclase rock: A striped actinolite-plagioclase rock with approximately 50 percent plagioclase (An_{16}) crops out locally along the northern boundary of the amphibolite mass, at locality 1-15. Needle-like actinolite crystals (pleochroism X=colorless, Y=yellow-green, Z=pale green; $\alpha = 1.671$, $\gamma = 1.683$; $Z \wedge c = 3°$) up to three millimeters in cross-sectional diameter make up most of the rock. The remainder is a fine-grained mosaic of quartz and plagioclase, free from mortar structure, which completely surrounds clots of actinolite needles.

Chlorite-talc-actinolite schist: The largest mass of chlorite-talc-actinolite schist in the region is located a few hundred feet south of locality 1-15. Beckwith (1939, p. 837) has noted similar rocks on Canyon Creek, about five miles west of that locality. Megascopically the chlorite-talc-actinolite schists are green to gray, medium- grained, highly foliated rocks with many crenulated folds, and with lineation parallel to that of the adjacent rocks. The schists are enclosed in amphibolite, which probably has been converted to chlorite-talc-actinolite rock along a shear zone. Where most highly sheared the schist is gray, contains mostly talc and chlorite, and has many crenulations. Zones of less sheared rock consist of actinolite-chlorite schist and are more greenish. In thin-section the rock consists of an intimate intergrowth of quartz, actinolite, ($\alpha = 1.633$, $\gamma = 1.645$) and chlorite, with pronounced mineral parallelism. Chlorite is commonly in larger crystals than actinolite, and is restricted to definite layers parallel to s-planes. Recrystallization has been in part later than the principal folding, as many folds are outlined by rows of unbent chlorite plates. Later shearing parallel to fold axes has broken the straight crystals in some folds, and part of the rock was recrystallized.

Chlorite-hornblende schist: Chlorite-hornblende schist lenses are associated with masses of medium-grained amphibolite at locality 1-7, and have gradational contacts with that rock type. The schist is green, medium-grained, usually very contorted, and contains augen of pink feldspar up to two inches in diameter. Irregular masses of quartz in schist are common. Occasionally chlorite rocks resemble amphibolite, but are green rather than black. Under the microscope the rock displays anhedral hornblendes (pleochroism X = yellow-green, Y = olive-green, Z = green; <1.5 mm) with very irregular outlines. Minute biotite plates (<1.5 mm) are associated with most grains of hornblende, and are intergrown with chlorite in parallel optical orientation. Some tiny plates of biotite are oriented at a slight angle to s_1.

Quartz-epidote-plagioclase rock: Light-colored lenses of quartz-epidote-plagioclase (An_{22}) rock within the gneiss crop out near localities 1-1 and 1-8. These are typically white, faint pink, or green rocks, and are rather resistant to hammer blows. Clinozoisite-bearing amphibolite is interlayered with gneiss just north of the parking area at Powder River Pass (Pl. 1).

GABBRO-PERIDOTITE

A mass of medium- to coarse-grained dark rock crops out at locality 1-9 (Pl. 1). Most of the mass consists of large interlocking crystals of pyroxene, which at places are arranged to give the rock a nearly horizontal plane structure. Locally the rock contains large plagioclase laths. Most pyroxenes are poikilitic, with small rounded inclusions of enstatite, commonly serpentinized, enclosed in augitic pyroxene. Local swarms of elliptical serpentinized enstatite masses are concentrated at places where pyroxenes are strongly poikilitic. The contacts of the mass are not exposed, though locally small red coarse-grained granite apophyses project into the basic rock. In thin sections a few rounded crystals of olivine, badly fractured and serpentinized, are associated with pyroxenes. Large microcline anhedra appear in several slides.

The mass of dark rock is fractured and traversed by numerous veins of mylonitic material. It is the only body of such rock known from the Big Horn region.

DIABASE DIKES

Diabase dikes are common throughout the Big Horn Range; their general distribution and character were described by Darton (1906a, pp. 19-22). The rocks are commonly black to dark green, blocky, fine-grained, and very hard; they weather chocolate brown and impart a characteristic brown color to the soil. In thin-section, diopside crystals ($\alpha = 1.683$, $\gamma = 1.694$) are cut by laths of plagioclase (An_{53}) and most grains are surrounded by laths of plagioclase though occasionally the reverse is true. Some of the pyroxene is polysynthetically twinned. Skeleton crystals of opaque minerals, probably magnetite, are associated with pyroxene, and are indented by and molded against it. Many pyroxene crystals are chloritized and biotitized.

Occasional coarse-grained dikes which approach gabbro in composition and texture are found along the upper reaches of South Tongue River, at localities 2-14, 2-15, and 2-16. At locality 2-14 a gabbroic dike 20 feet thick is separated from the gneissic granite by a fine-grained tough dark-colored rock with feldspar phenocrysts up to ½ inch by 1¼ inch which are occasionally grouped to form a pegmatitic layer. In thin-section the contact rock shows a medium- to fine-grained intergrowth of quartz, plagioclase (An_{45}), and biotite, with lesser amounts of chlorite, sericite and carbonate. Most plagioclases are small twinned lath-like subhedra, but some larger irregular masses are as much as two millimeters across; nearly all are altered to an ill-defined mass of sericite, carbonate, and dusty clay minerals. It is intergrown with, and embayed by, anhedral crystals of clear quartz, is cut by fine-grained microscopic veinlets of quartz, and contains many irregular-shaped inclusions of that mineral. In places, the two minerals look like a crude micrographic intergrowth. Clear anhedra of quartz (<2 mm) are embedded in the groundmass of the rock, and one quartz crystal encloses a grain of zoned plagioclase with a wide rim of myrmekitic quartz. Very irregular biotite laths are embayed by quartz. Chloritization of biotite is very extensive, and occasionally biotite is closely associated with odd-shaped pieces of muscovite in parallel optical orientation. These features suggest that the contact rock is the result of alteration of the basic rock (diabase?) by granite, rather than simple contact metamorphism of the granite by an intruded dike.

Amphibolitized and chloritized diabase: Many dikes which cut both the granitic and the gneissic rock series consist of varying amounts of chlorite or amphibole. Some of these have curving contacts and are of varying width, and many are foliated parallel to their walls. Some dikes pass laterally into unaltered diabase (Darton, 1906a, p. 22) and many have amphibolized or chloritized margins, with unaltered central parts. Moderately strong lineation of amphibole crystals is common. Typical examples of altered dikes crop out at the following localities: (1) 2-2, (2) in the bottom of Shell Canyon, near the confluence of Shell and Cedar Creeks (Fig. 1), (3) 1-17, (4) in a prominent outcrop at 1-18, and (5) near 1-6.

A thin-section of an amphibolitized and chloritized dike from locality 2-5, shows many small anhedral biotite, hornblende, and chlorite crystals in a subparallel alignment. The biotite and chlorite are commonly intergrown in parallel crystallographic orientation, and some intergrown biotite-chlorite plates are crinkled into minute folds with the axial planes at almost right angles to the s-planes of the rock. Irregular crystals of hornblende ($\alpha = 1.634$, $\gamma = 1.657$; X = yellow, Y = yellow-green, Z = blue-green) are embayed by quartz and plagioclase, and contain many small inclusions of those minerals. Irregular-shaped areas of plagioclase crystals with good polysynthetic twinning parallel to s-planes are scattered throughout the slide; they appear to have crowded aside the hornblende and biotite. Other areas in the slide show a granitoid intergrowth of quartz and feldspar; in these areas plagioclase twin planes show random orientation. The rock contains a few scattered anhedral grains of monoclinic pyroxene (diopside?), largely converted to hornblende and biotite. Garnet, epidote, apatite, muscovite, and magnetite are present as accessory minerals.

"Leopard Rock": "Leopard Rock" is a local term applied to diabase with large, white, subhedral crystals and aggregates of plagioclase (An_{80}) up to eight inches in length (Pl. 7, A), which are best shown in two long dikes near North Tongue River (Pl. 2). Locally diabase dikes located at 1-6, 1-8, 1-1, and 2-17 show large plagioclases and the two large dikes near Tongue River are locally free from white crystals.

The longer dike extends for approximately six miles from locality 2-18 to 2-19. The dike strikes N. 80° W., and dips 80° to 85° south. A small amount of material taken from the dike near Twin Buttes was used locally for ornamental building stone, and the U. S. Civilian Conservation Corps quarried rock from the dike at 2-20 which was used to decorate highway culverts (Haff, 1944a). The thickness of the dike varies from 30 to 60 feet except where highly sheared. A red fine-grained rock crops out along the contact between dike and granite at locality 2-21; the red rock contains a few disseminated minute specks of pyrrhotite. Feldspar-free layers within the dike itself also bear pyrrhotite. Marked layering of large white feldspar crystals parallels the walls of the dike in a small open cut at 2-20.

A thin-section from 2-20 consists mainly of interlocking plagioclase laths (<1.5 mm long) and diopside; plagioclase is slightly more abundant. Scattered aggregates (<4 mm) of plagioclase are analogous to megascopically visible white crystals, and contain fine-grained sericite, zeolites, and clay around grain boundaries and along fractures. Crystals near the edge of an aggregate are sometimes zoned, with a rim slightly more calcic than the core. Neutral to pale-greenish crystals of diopside ($<1\frac{1}{2}$ mm) are anhedral to subhedral and sometimes are twinned parallel to (100).

A few discrete grains and skeleton crystals of magnetite-ilmenite are intergrown with quartz and feldspar, and a few crystals of chlorite, most of which are associated with pyroxene, are scattered through the thin-section. A thin-section from the eastern end of the southernmost long dike (Pl. 2) contains diopside, hornblende (X = yellow-green, Y = green, Z = dark green; $2V = 54°$), and chlorite arranged in small clots which have a radiating to fibrous structure because of the arrangement of hornblende needles. The groundmass is an intricate intergrowth of quartz, plagioclase laths (An_{37}), and anhedral grains of untwinned potash feldspar. Some plagioclase is zoned, and all feldspars are extensively altered as previously described.

Several large anhedral insets of clear quartz ($>2\frac{1}{2}$mm) are scattered through the section. A red granitic rock beside the dike at this locality contains chlorite, plagioclase (An_{34}), epidote, potash feldspar, and actinolite (X = faint yellow-green, Y = very pale green, Z = pale green). Most of the small equant anhedra of potash feldspar are Carlsbad-twinned, and plagioclase laths interlock and show a rude but fine-textured myrmekitic intergrowth with quartz. Thin-sections from dikes that contain only sporadic insets show relationships similar to those described.

Similar dikes were described from the New World Mining District, Montana (Lovering, 1929, p. 17), and in pre-Cambrian volcanic flows in Ontario (Graham, 1932, pp. 35-36; Hurst, 1932, pp. 9-11, 33; Laird, 1932, pp. 10-11; Rickaby, 1932, pp. 5-6; Thompson, 1932, p. 34). The origin of such rocks is obscure but the composition of the insets probably demands an addition of calcic material to the magma, perhaps by assimilation.

CRUSHED ROCKS

Narrow elongate belts of crushed and recrystallized rocks, of various kinds, cross the northern granite area. Most belts are oriented nearly east-west, but a few strike north-northwest. Most of the rocks, which resemble quartzites, are pale pink, gray or green, very fine-grained quartz-rich types. Others are green vitreous-looking varieties with ellipsoidal quartz crystals up to five millimeters long. Some dark colored crushed rocks are probably altered and sheared dikes. With the exception of the basic types, most crushed rocks have a relatively high quartz content, and for this reason most are very resistant to weathering and form low, elongated ridges up to 15 feet high.

Quartz Rock: Numerous white, gray, green, or violet quartz-rich rocks crop out in the extreme northern part of the granite area (Pl. 2) near North Tongue River. Masses of quartz rock are usually tabular or lenticular, are strongly fractured and are commonly seamed by close-spaced epidote veins ($<\frac{3}{8}$ inch wide) in diverse orientations. Many small vugs are filled with small subhedral quartz crystals, and locally manganese oxide dendrites appear on fracture surfaces. Margins of most quartz rock lenses are obscure; the rocks merge into fine-grained quartz-epidote granite. Locally the quartz rock contains swarms of dark rock fragments, commonly elongated parallel to the direction of the swarm.

A thin-section from location 2-32 consists of about 65 percent anhedral quartz (<2 mm), but with a wide variation in grain size; mortar structure is common. In some areas of the slide, small quartz crystals are arranged along closely spaced parallel planes, separated by larger crystals. Areas which consist mostly of granulated quartz are sprinkled with anhedral epidote (<0.2 mm). Crystals of plagioclase (An_{17-31}) (<0.4 mm) show good albite twinning, are angular in outline, and are set in a matrix of fine-grained quartz and feldspar. Twin lamellae of plagioclase are commonly bent or offset.

A thin-section of fine-grained quartz-epidote granite from the contact with quartz rock at 2-22 shows many closely-spaced parallel fractures about one-third of a millimeter apart. Some fractures are filled with vein quartz, more rarely with epidote. Most of the rock consists of quartz, plagioclase, and potash feldspar, listed in order of decreasing relative abundance. A few areas in the slide consist of broken pieces of quartz and feldspar, with interstitial recrystallized quartz and scattered aggregates of fine-grained epidote. Some potash feldspar anhedra (<1.5 mm) are cut by a vein of albite parallel to the closely spaced fractures. Apparent displacement along fractures is about one-fifth of a millimeter along planes with no granulation debris, but more displacement may have taken place along zones of fragmentation.

Green crushed rock: Several lenses of green rock with ellipsoidal quartz crystals crop out near North Tongue River, north of U. S. Highway No. 14. Sometimes several masses of green, siliceous, crushed rock are within a single outcrop of granite, as at localities 2-23, and 2-24. A green rock lens at 2-23 is slickensided parallel to the trend of the lens, and red massive granite next to the green rock is cut by many narrow epidote seams. Nearby a partially sheared mass of red granite contains many scattered rounded pink potash feldspar crystals.

A thin-section of the green rock from the Hope No. 2 Claim (2-25) shows strong cataclastic structure, and subsequent recrystallation. Anhedral quartz (<1 x 2 mm) makes up approximately 50 percent of the rock and only rarely exhibits strain shadows. A few quartz grains show secondary growth, the original grain boundaries marked by a thin line of sericite and chlorite. Anhedral chlorite laths ($X =$ faint green, $Y =$ colorless; <1 mm long), which sometimes enclose small garnets, have very irregular outlines and jagged ends. A few inclusions of opaque minerals and of quartz are in chlorite; some quartz inclusions are optically continuous with a surrounding crystal. Small rosettes of pennine (<0.5 mm), with some sericite, quartz, and carbonate are in the matrix of the rock. Subhedral elongated prisms of hornblende ($X =$ yellow-green, $Y =$ olive-green, $Z =$ grass-green; $2V=82°$) in a thin-section of granite adjacent to the green rock contains rounded inclusions of quartz and feldspars. The prisms are embayed by quartz and feldspars, and some show good sieve structure. Biotite plates ($<1/5$ mm; $Z =$ dark olive-green, $Y =$ olive-green, $X =$ pale yellow) show a fair to high degree of parallel alignment, and most have frayed ends; some biotites contain minute inclusions of apatite and garnet.

Chlorite-quartz rock: A zone of dark green chlorite-quartz rock with many rounded pink potash feldspar crystals and small quartz veins one inch or less in width may be seen in a small prospect pit at locality 2-26. Red granite adjacent to the dark rock is strongly epidotized, and is cut by numerous quartz veins containing much fine-grained chlorite. Small vugs less than one inch in diameter within the epidote granite contain many small euhedral to subhedral epidote crystals. In thin-section the rocks show aggregates of discrete subhedral to anhedral epidote crystals (<0.3 mm) associated with quartz and feldspar. The epidote crystals commonly fill irregular fractures in quartz and feldspars, and have their long axes oriented parallel to the fractures. Some felted aggregates of small irregular muscovite plates (<0.2 mm) are in small sheaves or spherulite-like aggregates. Small, discrete epidote subhedra are scattered throughout muscovite crystals that are associated with a few unstrained anhedral grains of quartz. Pennine rosettes ($X =$ pale green, $Y =$ pale green, $Z =$ pale yellow; $c \wedge Z = 0°$) and minute crystals of stilbite (?) are in granulated zones. A portion of the slide contains an intergrowth of quartz and strained potash feldspar crystals similar to that in the gneissic rocks, which suggests granulation and recrystallation. Anhedral grains of quartz (<2.0 mm) vary widely in grain size, and have extremely irregular boundaries. The mineral contains many rows of liquid and dustlike inclusions, and small subhedral crystals of epidote. Boundaries between quartz and pennine are very irregular; chlorite penetrates quartz along cracks, and a few cracks leading into quartz from chlorite are lined with serpentine; these relations imply that chlorite replaced quartz during or after the period of granulation.

"*Peridotite*": Darton (1906a, p. 23) described a dike of "peridotite" of pre-Cambrian age in the ridge west of the lower part of South Fork Tongue River (Pl. 2). This dike, though deeply weathered, stands as a low ridge for a part of its length, and in places projects through Cambrian shales as at locality 2-27. The rock is dark-green, medium- to coarse-grained and granular, and is locally invaded by granite. It has been sheared along several sets of fractures, and is traversed by numerous quartz veins less than one and one-half inches thick, some of which show sheet structure. A sheared epidotized granite is the usual country rock, but one lens of "peridotite" is within green crushed rocks. The western end of the "peridotite" dike is in contact with a rock that contains approximately equal amounts of quartz and specularite. Along the northern contact of the dike the granite is silicified.

Thin-sections of "peridotite" contain subhedral to anhedral diopsides (<0.6 mm; $\gamma = 1.658$) that are extensively altered along cleavage planes and fractures to fine-grained aggregates of talc. Other crystals are strongly altered to anhedral and subhedral hornblendes ($X =$ greenish-brown, $Y =$ yellow, $Z =$ pale bluish-green), and scattered anhedral amphiboles, probably tremolite. The tremolites are colorless and monoclinic. Some diopside is altered to chlorite and talc around crystal boundaries. Commonly, hematite is disseminated in areas of chloritization and steatitization. Several aggregates (<1.5 mm) of finely crystalline zeolites resemble anhedral to subhedral feldspars in shape, and some are surrounded by ferromagnesian minerals in a crude ophitic intergrowth. A few small rounded fractured olivine grains are extensively altered to talc, chlorite, and amphibole.

"*Cortlandite*": A black, coarse-grained, poikilitic rock within red granite at locality 2-25 was tentatively called "cortlandite" by Haff (1944b). A small amount of disseminated pyrrhotite is megascopically visible within the rock. The rock is an irregular lens which trends north-northwest, and is a few hundred feet long and fifty feet wide.

Thin-sections of the rock show an uncommon intergrowth of monoclinic pyroxenes; some individual crystals have vague irregular portions of pleochroic and non-pleochroic material. The pleochroic pyroxene has axial colors of $X =$ colorless, $Y =$ neutral, $Z =$ faint green; $Z \wedge c = 40°$; $2V$ is 58°. Non-pleochroic pyroxene has $Z \wedge c = 19°$. All crystals are anhedral, and the average grain size is about two millimeters. Crystals of both kinds of pyroxene are highly poikilitic, and contain many inclusions of antigorite, quartz, talc, and carbonates, and fewer inclusions of pyrrhotite and magnetite. The outlines of some completely destroyed pyroxene crystals are outlined by the opaque minerals. Many former pyroxene crystals are now only a fine-grained aggregate of alteration minerals, which made refractive index determinations impossible.

The lens of "cortlandite" is surrounded by a medium-grained granitic rock which contains about 25 percent dark minerals (biotite and chlorite), and megascopically visible pink feldspar crystals up to one-quarter of an inch long. Under the microscope anhedral quartz and feldspars (<3.0 mm) have serrate and interlocking borders. Most small plates of plagioclase within potash feldspar are optically continuous with each other. Plagioclase (An_{32}) has fine-textured albite twinning, and is altered to fine-grained sericite and zeolites along twin planes and in pervasive areas. Twin planes of both feldspars are bent. One plagioclase grain contains small potash feldspar inclusions which resemble "advance islands" from an adjacent grain of potash feldspar. Large potash feldspar crystals seem to have spread throughout the slide and to have surrounded other minerals.

TRANSITIONAL ROCKS

A gray coarse-grained gneissic rock with about 20 percent biotite, and a few dark-colored fine-grained contorted layers, crops out about 700 feet west of Dome Lake (Fig. 1). The fine-grained layers sometimes contain large ellipsoidal clots of white feldspar averaging about 4 inches long (Pl. 7, B), but some reach 5 or 6 inches. Within the feldspar masses are coarse subhedral crystals of biotite ($\beta = 1.628$) set in a matrix of oligoclase (An_{21}). Most of the biotite is arranged in clots, parallel to the long dimension of the ellipsoids. When seen on surfaces at right angles to the elongation, the ellipsoids are almost circular in outline. Gray gneiss containing the feldspar masses is a very large inclusion within the lighter colored gray gneissic granite, which has a few large crystals of pink potash feldspar and a number of biotite-rich inclusions of varying degrees of darkness.

The elliptical outline of the feldspathic masses resemble spheroidal granites. They differ from previously described orbicular rocks (Sederholm, 1928; Eskola, 1938, *et al.*) in several ways: (1) most orbicular rocks are arranged with concentric shells of biotite-rich and biotite-free material, whereas in the Dome Lake rocks biotite is clustered at the center of the ellipsoids, (2) plagioclase crystals in most orbicular rocks are radially arranged, but in the Dome Lake rocks they are granoblastic, (3) orbicules of other rocks approach spheres rather closely, but the Dome Lake ellipsoids are "football shaped" with pointed ends.

Sederholm (1928) believed the orbicular rocks to be caused by repeated fractional crystallization in viscous magma, but thought diffusion within the magma was necessary because of the growth from the center outward (as evidenced by the radial structure), and for the spacing of the centers of individual orbicules within the rock mass. Eskola (1938) and Simonen (1940) believed that the orbicules are due to metamorphic diffusion and metasomatism, and that they grow from the center out.

The difference between truly orbicular granites and the Dome Lake rock make it obvious that their origins are not wholly similar. Clusters of biotite crystals at the centers of the Dome Lake ellipsoids make it unlikely that they are caused by fractional crystallization; to derive the ellipsoids by fractional crystallization it would be necessary for biotite to crystallize first, followed by oligoclase and quartz. This process requires crystallization to start on the discontinuous side of Bowen's reaction series (biotite), then to stop completely, and resume on the continuous side—a most unlikely procedure.

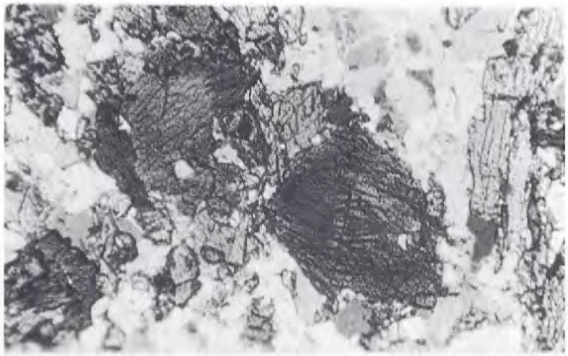
The coincidence of lineation in the surrounding rocks, elongation of ellipsoids, and elongation of biotite clots suggests that the ellipsoids were formed approximately at the same time as the surrounding rock, or at least that both were deformed at the same time. The ellipsoids were probably

Pl. 4, A. View southwest across lower valley of Little Tongue River, near Horseshoe Ranch. Pediment surface in middle ground cuts Mesozoic rocks. Flank of mountain in background is composed of flatirons of Tensleep and Madison formations.

Pl. 4, B. Sawed surface of metagglomerate from locality 1-3. Insets are quartz, feldspar, hornblende, and biotite.

Pl. 5, A. Fragmental texture of pre-Cambrian volcanic agglomerate cropping out in bank of Caribou Creek, at locality 1-3. Scale is indicated by 6-inch rule.

Pl. 5, B. Fragmental texture of pre-Cambrian volcanic agglomerate cropping out in bank of Caribou Creek at locality 1-3. Scale is indicated by 6-inch rule.

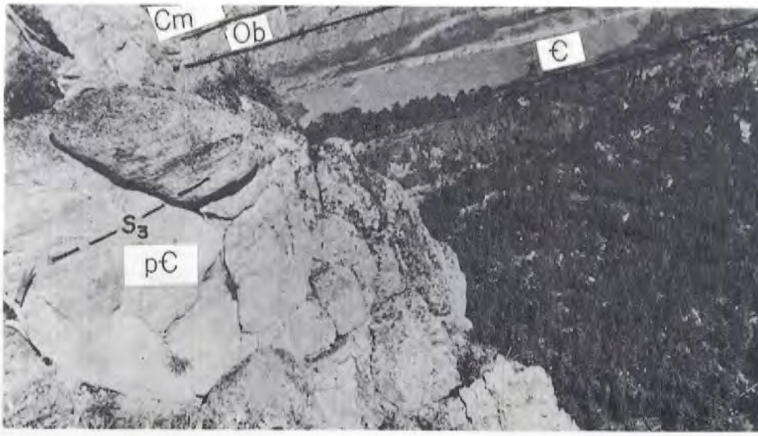


Pl. 6, A. "Rolled" and fractured hornblende crystals. Individual fragments are drawn out, parallel to s_1. Crossed nicols, 50x.

Pl. 6, B. Small fold in gray gneiss, with numerous oligoclase augen along the fold axis. Photo taken at road cut along U. S. Highway No. 16, at top of divide between North and Middle Forks of Clear Creek. Six-inch rule indicates scale.

Pl. 7, A. "Leopard Rock" (porphyritic diabase) with large phenocrysts of plagioclase which reach a maximum size of 8 inches. Locality 2-18.

Pl. 7, B. Elongated feldspathic ellipsoids with linear clots of biotite in gray gneissic granite. The outcrop is about 700 feet west of Dome Lake. Photo by R. Balk.



Pl. 8, A. View southwest across Wolf Creek Canyon. Outcrop of gneiss in foreground dips 17° northeast, parallel to upper surface of pre-Cambrian in middle ground. One set of joints parallels foliation in gneiss. Lower Paleozoic sediments crop out in distance. Photo by R. Balk.

Pl. 8, B. Clear Creek thrust fault, 7 miles west of Buffalo. View north obliquely across the strike of upturned Paleozoic beds which rest on Eocene conglomerates.

formed in one of two ways: (1) diffusion of Fe, Mg, and Al to centers (the present biotite cores), with accompanying impoverishment of the surrounding area (oligoclase-quartz ellipsoids), or (2) local separation of feldspar-rich liquid drops within a plastic rock mass with diffusion of material to the center of the drops to form biotite-rich clots.

PALEOZOIC ROCKS

This report is based on a study of the pre-Cambrian rocks and their relation to the uplift of the range, hence the Palezoic and Mesozoic sediments were not studied in detail. Some of the structural conclusions depend upon the stratigraphy of the Cambrian rocks; they will be mentioned briefly.

"Deadwood" formation: The "Deadwood" formation represents the only Cambrian rocks in the Big Horn Mountains. The formation has an average thickness of about 900 feet (Darton, 1906a, p. 23), but local variations of thickness are common as shown by the width of outcrop on plate 2. Zakis (1950, pp. 18-20) measured 1128 feet of Cambrian rocks near the south branch of Little Tongue River (Pl. 2), and 70 miles south, Carlson, (1949, pp. 24-25) measured only 897 feet. At the type locality in the Black Hills, Darton (1904, p. 382) measured about 400 feet. Closely adjacent sections measured only 20 miles apart (Carlson, 1949, pp. 24-25, and Richardson, 1950, pp. 26-28) cannot be correlated; both contain more limestone and less shale than the section near Dayton, Wyoming (Zakis, 1950, pp. 18-20). Dorf and Lochman, (1940, p. 555) found that the Big Horn Mountain Cambrian rocks could not be correlated lithologically with Cambrian rocks in southern Montana. Carlson's, Richardson's, and Zakis' sections cannot be correlated with sections measured 65 miles to the southwest (Deiss, 1938).

No break in lithology between rocks of Middle Cambrian and Upper Cambrian age can be detected in the Big Horn region (Meyerhoff and Lochman, 1938, p. 385; Deiss, 1938, p. 1165). Lochman (1950) states this is not the result of a hiatus, but to slow deposition of the upper portion of the Cambrian sections, which caused faunal zones to be sufficiently close together to suggest a time break. Thomas (1949, p. 43) concludes that Darton's Deadwood formation (1906a, pp. 23-26) includes rocks of both Middle and Upper Cambrian age, and thus represents a greater time span than the type Deadwood section in the Black Hills. He suggested that "Flathead" be used for the basal sandstone, "Gros Ventre" for the green shale unit, and "Gallatin" for the upper sandstones. Durkee (1953, p. 1544) subdivided the "Deadwood" along the east flank of the Bighorns into three formations, Flathead, Gros Ventre, and Grove Creek, using names extended from western Wyoming. The writer, however, cannot concur entirely with these subdivisions. The basal sandstone is markedly lenticular and is entirely absent in many areas, particularly near the western boundary of the area shown on plate 2. For this reason the term "Deadwood" is retained, but is placed in quotation marks to indicate that it is not exactly equivalent to the section at the type locality of the formation.

The contact of Cambrian and pre-Cambrian rocks is obscure in the Big Horn region, especially where the basal sandstone is very thin or absent, because "Deadwood" shales form rounded grassy slopes which pass imperceptibly into undulating weathered surfaces of pre-Cambrian granite in the Tongue River region. This contact is easily mapped at a certain time in the summer when wild flowers are abundant, because the soft Cambrian shales favor the growth of vivid blue *Lupine,* while a rose-colored unidentified flower is abundant on adjacent weathered granite slopes.

Most of the "Deadwood" formation in the northern Big Horns is soft green shale with interlayered brown and buff sandstone lenses, layers of blue and gray limestone, and some coarse-grained calcareous glauconitic flat-pebble conglomerate. The lenticular buff coarse-grained arkosic sandstone at the base of the section was thought to be a continuous unit by Demorest (1941), which led him to erroneous structural conclusions.

Thin-sections of the basal "Deadwood" arkosic sandstone from locality 2-28 show subangular grains of quartz and microcline, embedded in a matrix of fine-grained sericite, kaolin, and carbonate. The rock is not well sorted. Within the matrix are smaller grains of quartz, feldspar, and opaque minerals. The matrix is finely crystalline, and contains scattered spherulitic aggregates of chlorite, though the rock is not well cemented. The rock probably was formed from locally derived granitic debris, with a small amount of diagenetic chlorite. Zakis (1950) concluded that the "Deadwood" formation was formed under stable conditions in rather shallow water, with much winnowing and reworking of the clastic debris. The rocks probably were deposited in a platform or shelf environment (Lochman, 1950, pp. 42-51).

CENOZOIC ROCKS

Early Tertiary rocks: Isolated outcrops of Cenozoic rocks are common on the upland surface north and south of the central gneissic core of the range. Early Tertiary sediments flank the east side of the range north and south of Buffalo for a distance of about 50 miles (Fig. 1). These sediments overlie rocks of pre-Cambrian to Cretaceous age with marked unconformity, and are unconformably overlain by younger Tertiary rocks. The rocks of early Tertiary age are gravels and beds of boulders up to 10 feet in diameter that grade into finer clastics further from the mountain front. The coarse clastic Tertiary sediments were named "Kingsbury conglomerate" by Darton (1906a, pp. 60-62). The Kingsbury is Eocene (Brown, 1948).

Coarse clastic rocks above the Kingsbury were called "Moncrief Gravel" (Sharp, 1948). Kingsbury and Moncrief strata may be distinguished in the field because Kingsbury beds contain roundstones of Paleozoic and Mesozoic formations, but no pieces of pre-Cambrian rocks, while Moncrief rocks have mostly pre-Cambrian fragments (Sharp, 1948, pp. 1 and 5; Darton, 1906a, p. 69, *et al.*).

White River group: Outcrops of sand, volcanic ash, gravel, and boulders in the Big Horn Mountains are common on mountain divides up to about 9,100 feet altitude (Darton, 1906a, pp. 67, 68, 69, 70). Several of these outcrops are near Tongue River, and some are within the area of plate 2, at localities 2-29, 2-30, and 2-31.

The Tertiary rocks at 2-29 are gray and maroon poorly consolidated coarse-grained calcareous sandstone. Most of the grains are sub-rounded pieces of clear quartz, with a few badly altered feldspars. Layers of fine-grained light dove-colored clays are interbedded with the sandstones.

The flat-topped divide between Copper Creek and an unnamed creek to the northwest is underlain by similar deposits (2-30) about 30 feet thick, and is capped by a coarse unsorted gravel. Where the rocks are well cemented they contain fragments of pre-Cambrian, "Deadwood", Bighorn and Madison rocks, with lesser amounts of later sediments. The poorly rounded fragments range up to boulder size, have a somewhat flattened shape, and are set in a medium- to fine-grained white calcareous matrix. The conglomerate is interbedded with clay and sandstone. Pre-Cambrian granites beneath the Tertiary have an upper surface with 20 feet of relief.

The best outcrop of Tertiary rocks is one in which 15 feet of dove-colored fine-grained clay crops out on Camp Creek (2-31) in an eroded slope beside the Freezeout Point motor trail. The clay is capped by beds of coarse cobbles and boulders in a fine-grained matrix. The total thickness of the Tertiary rocks cannot be estimated at this locality due to poor exposures. Dr. P. O. McGrew identified a bone fragment found by the author at this outcrop as a medial phalanx of *Mesohippus*, thus establishing for the first time the Oliogocene age of the sediments.

STRUCTURE

The internal structures of the rock types have been discussed in the rock descriptions. This discussion treats with the interrelationships and variations of the structures. The relations between geologic structure and lithologic types are examined, and correlations are pointed out between early and late deformational periods.

Lineation

Rocks in the south central portion of the Big Horn Range (Pl. 1) along the route of U. S. Highway No. 16, show a strong lineation. The lineation generally plunges northeast, though local variations are common. Amount of plunge becomes less and direction of plunge more nearly east-northeast near Hazelton (Fig. 1, Pl. 1) where the rocks contain more amphibolite. A large hornblende schist mass crosses the divide between Canyon Creek and South Fork Crazy Woman Creek.

Northeast of the area traversed by U. S. Highway No. 16, plunge of lineation changes from northeast to southeast, but data are lacking from this region. Lineation near the Seven Brothers Lakes and near West Tensleep Lake plunges uniformly southwest (Pl. 1). The most variable lineation is along U. S. Highway No. 16, near North Fork Clear Creek (locality 1-16). This change in plunge of lineation probably implies a northeast striking pre-Cambrian fracture zone, according to data presented by E. Cloos (1946, pp. 43-44) from the Appalachians, and by Balk (1953) from New England.

Lineation in the granitic rocks, though very weak, is most common in dark biotite-rich inclusions, probably because the rock mass flowed during folding of the large compound syncline (Osterwald, 1949; fig. 3). Plastic deformation during this folding is suggested by the inclusions that are elongated parallel to s-planes, by the lineation within inclusions which plunge down the dip of s-planes, by the dismembered basic dikes, and by the absence of crushing effects in thin-sections.

S-planes and Folds

Gneissic layering, though less pronounced than lineation in gneissic rocks, is common at many localities, and is particularly strong near Powder

River Pass. Layering in gneisses (s_1) was thrown into numerous small isoclinal folds, commonly with axes plunging northeast, parallel to the lineations (page 9).

Axial plane cleavage (s_2) is most pronounced in a zone trending southwest from Sheep Mountain (Pl. 1), and passing northwest of Powder River Pass. S_2 strikes constantly northeast, and either dips steeply southeast or is vertical. Where crestal portions of folds show cleavage, the intersection of s_2 with s_1 is parallel to the lineation. Locally s_2 offsets s_1. Biotite and chlorite line the cleavage planes.

Because lineation is parallel to axes of isoclinal folds the axes of small folds probably parallel the major fold axes, and lineation probably parallels the "b" fabric axis (Sander, 1926, p. 328; Cloos, E., 1946, pp. 5, 6, *et al.*). Thus the northeast-plunging lineation in the Big Horns suggests that the direction of tectonic strike is also northeast. The southeast-plunging lineation near Seven Brothers Lakes and West Tensleep Lakes (Pl. 1) may parallel the direction of early tectonic movement, since it is approximately at right angles to "b". Caution must be used in interpreting lineations perpendicular to "b", because they may indicate only a component of movement in this direction (Cloos, E., 1947).

Most of the gneisses are probably B-tectonites because a lineation parallel to the tectonic strike is the most prominent structure (Sander, 1930, pp. 58, 220, 221; Knopf and Ingerson, 1938, pp. 68-71, 154). Where the gneisses contain stronger s-planes, but lack lineation and microscopic mineral alignment, the term S-tectonites (rocks with a single set of megascopically visible s-planes) could be applied. Rocks with strong lineation and s-planes are gradational between B-tectonites and S-tectonites (Turner, 1948, p. 199; Fairbairn, 1949, p. 6); such gradations are common in the southern Big Horns.

Folded s-planes, and microscopic "rolled" hornblende crystals (page 10) indicate that rotation was important in the genesis of the gneisses, and that these rocks are R-tectonites (Turner, 1948, p. 199). As a result of the folding, the attitude of s_1 is extremely variable, even within a single outcrop. S_2 planes are later than s_1. Large feldspar augen and masses of pegmatitic material in the crests of folds (pages 13-15) suggest that minute openings produced during folding of s_1 aided in the formation of feldspars.

Foliation (s_3) in granitic rocks: S-planes are more pronounced than lineation in the Tongue River granite area. The foliation of granitic rocks is termed s_3, and though it is probably later than s_1 and s_2 because of the general relations of the granitic and gneissic rock series, it cannot be observed in direct relation to s_1 and s_2.

The attitude and distribution of s_3 within the granite mass is extremely variable (Pl. 2). Commonly s_3 is constant for one or two miles, as along the ridge east of South Tongue River downstream from Arrowhead Lodge, but attitudes at outcrops in adjacent areas vary between wide limits. Variation of s_3 is common in individual outcrops, but tight folding occurs only along the eastern margin of the range, near Wolf Creek. The foliation strikes about north-northwest along the margins of the syncline near South Tongue River; to the east it is more irregular. Orientations of s_3 vary widely at many outcrops, and near the eastern margin of the pre-Cambrian area, northeast strikes of s_3 planes are not uncommon. Enough dip and strike measurements probably were made to arrive at a significant conclusion concerning the general structure of the mass. Though attitudes of s_3 planes exhibit much local variation, it is evident from figure 3 that (1) s_3 planes outline a large branching syncline, the "South Fork Syncline" (Osterwald, 1949, p. 37), which plunges north-northwest in the western part of the granite area and is marked by east-west strikes along the crest of Bruce Mountain which change to west-northwest at the west end of the mountain and east-northeast at the east end, and (2) east of the large syncline, s_3 planes dip more gently, and become more variable in attitude.

The northeast strikes of s_3 along the eastern margin of the granite mass (Pl. 2) may be the result of either: (1) the dominant strike of s_1 in gneisses to the south, or (2) a rotation of blocks of pre-Cambrian rocks during Laramide deformation.

Joints

Joints in gneiss: Numerous tension joints in gneisses (Pl. 1) dipping almost at right angles to lineation, or a little steeper, are probably the result of lengthening of the rock mass in a direction parallel to lineation (Fairbairn, 1949, pp. 229-230; Cloos, E., 1946, p. 37). Cross-joints commonly dip at a slight angle to the *ac* plane of the fabric axes (Turner, 1948, p. 182). Such cross-joints probably form because the rocks are left in a high state of elastic strain at the close of deformation (Turner, 1948, p. 182). Release of deforming forces causes internal forces within the rock to exceed the elastic limit, and causes rupture.

The rupture along the joints is caused by straining the atomic structure of crystals beyond the normal state of equilibrium (Bridgman, 1938). When the force field upon the various atoms of the lattice is disturbed, rupture may result, even without resorting to tensional or other forces, increase in volume, or any other of the usually accepted causes of rupture. The vector of gravity acting upon the internal force system of the crystal lattices may cause cross-joints to be deflected toward the vertical. If the joints were formed at the time the rock was deformed (not after the release of deforming pressures) granitic material, or at least mineral films, probably would have formed upon them during the growth of augen and of granitic dikes and veins. The tension joints are probably analogous to the fractures in Griggs' limestone cylinders which formed perpendicular to the axis of the cylinder when tensional stresses were relaxed (Bridgman, 1938, p. 527).

Many outcrops of gneiss show several intersecting joint planes, with the common line of intersection parallel to lineation. Dip of the joints is extremely variable, even at a single outcrop. The joints are parallel to *b*, and are probably shear factures caused by rotation during folding of s_1 planes.

Joints in granitic rocks: The granitic rocks of the Tongue River district are broken by numerous joint planes that strongly influence local topography (Darton, 1906a, pp. 16-17). The only previous attempt to correlate joints in the core of the Big Horn Range with other structural features was made by Wilson (1934) at Five Springs Creek.

All joint measurements made in the granite are plotted on plate 3, which shows considerable variation in the attitude of joints. At most outcrops visited in the field, the joints may be classified in three sets, one that is subhorizontal, and two that dip steeply. Commonly one steeply dipping set strikes northwest and the other northeast. In many localities, one joint set parallels s_3.

The poles of 264 steeply dipping joints, contoured on the upper hemisphere of an equal area net shown on plate 3 as diagram A, show a pole concentration representing a statistical joint plane that strikes north-northeast, and another which represents a plane striking northwest; both planes dip steeply. Another statistical joint surface strikes northeast and dips vertically, but the concentrations are not so well defined, and may not be significant. Sub-horizontal joints were not included on this diagram, because they probably exerted no influence on the trend of the mountains. A somewhat similar technique was used to study joints in the District of Columbia (Fellows, 1950, p. 272).

Wherever feasible, the average spacing of joints belonging to each set was measured in the field. The most prominent joint set within the granite mass was determined by the spacing measurements. To do this, a "spacing factor" was applied to each joint measurement, according to the average spacing measured at the outcrop. A joint set with an average spacing of 1 foot was assigned a "factor" of 1.0, and average spacing of 2 feet a factor of 2.0, and so on to a factor of 10.0. Spacings over 10 feet were not measured in the field, so no factor exceeds 10.0. Spacings measured in inches were converted to tenths of feet and assigned corresponding decimal spacing factors. The spacing factor for each joint set was plotted on the upper hemisphere of an equal area diagram, at the pole of the joint set (Pl. 3, diagram B). The method of plotting is analogous to that used by Newhouse, Hagner, and DeVore, (1949, p. 168) to study the structural relations of the mineral composition in rocks. Though the diagram represents the *average* spacing of joints at individual outcrops, the result is sufficiently pronounced to validate the interpretations based upon it.

The spacing diagram (Pl. 3, diagram B) shows a pronounced maximum of closely spaced joints corresponding to a plane which strikes northwest and dips steeply northeast. Lesser maxima represent northwest- and northeast-striking planes. The close correlation between the strike of the plane represented by the major maximum on the diagram, the major syncline in granitic rocks, and the trend of the Big Horn Range is apparent.

Throughout the granitic rocks sub-horizontal joints are widespread; many are shown on plate 3. They are abundant in Shell Creek Canyon along the western margin of the range, in The Horn Ridge northwest of Mayoworth in W½ T. 46 N., R. 83 W., and in the pre-Cambrian rocks south and east of Bald Mountain (Fig. 1). The sub-horizontal joints are commonly conspicuous in the landscape because, with the two steeply dipping sets, they divide granitic outcrops into cubic or rectangular blocks.

Most joints probably are the result of a change in rock volume because of movement perpendicular to the joint plane (Nevin, 1949, p. 146), or of disruption of the force field within crystal lattices (Bridgman, 1938, pp. 527-528). Horizontal joints, accordingly, should represent either a strain in which upward motion in a rock mass is easier than lateral relief, or in which the atomic force field is disrupted by unloading or by effective vertical relief of pressure. Horizontal joint sets commonly are interpreted as the result of elastic strain relieved by erosional unloading; examples are recorded from quarries in Vermont where rocks fractured *after* being exposed (White, 1946, p. 4). Many quarry faces have more horizontal joints (sheeting) near the surface of the ground than at depth (White, 1946, p. 6; Jahns, 1943, p. 71). Gilbert (1904, pp. 29-36) explained joints parallel to the rock surface in the Sierra Nevada by dilation due to unloading. The horizontal joints in the Big Horn granite probably are not caused by release of strain through erosional unloading, at least in the present geomorphic cycle, for the following reasons: (1) many pegmatite and aplite dikes follow horizontal fracture planes, as well as steep joints, (2) rock surfaces are not parallel to horizontal joints, and (3) the number of joints does not increase toward the top of the canyon walls. The horizontal joints probably are due to upward relief of stress during the last stages of deformation at the time the granite was formed.

Movement on joints: Many joint surfaces are covered with limonite, epidote, or thin films of silicified rock. A few joint surfaces are crudely slickensided. However, it is not known at what time the movement took place; it may have been much later than the opening of the fractures, perhaps even Laramide slipping along joints formed before Cambrian time.

Recurring stress fractured the rocks at different times, as shown by outcrops on the north bank of Owen Creek east of U. S. Highway No. 14, and on the north slope of Lookout Mountain, where joints cut and offset pegmatites.

Many pegmatite and aplite dikes parallel the predominent joints. A few dikes are jointed parallel to their walls, probably as the result of recurring stress. Pegmatite and aplite dikes are not all of the same age, because many early dikes are cut by late ones and many of the early dikes are folded. The dikes probably filled early fractures, were plastically deformed, and later cut by new fractures which were filled by pegmatite and aplite.

Pegmatite and aplite dikes filling early joint planes in crystallizing granite massifs are related to a pattern of parallel mineral grains related to the movement of magma (Cloos, H., 1925; Balk, 1948). Early joint patterns in the massif may be at right angles or parallel to the motion. During the later phase of crystallization, pegmatite and aplite dikes may fill the oldest joint planes, particularly in the cross-joints (those at right angles to the movement). Similar dikes in old joints in the Tongue River granite mass, however, probably are not related to magmatic flow, for the following reasons: (1) Cloos and Balk, in their studies of granite massifs, recognize arches and domes as the dominant geologic structures of intruded granite masses; the Tongue River granite has a complex synclinal structure, (2) if the early joints in the Tongue River granite were cross-joints similar to those of Balk and Cloos, linear mineral parallelism

A — Section normal to joint plane (see text) and to its dip. 157 quartz axes. Contours at 1%, 3%, 4%.

B — Section parallel to joint plane. 150 quartz axes. Contours at 0.5%, 2%, 4%, 5%.

C — Section normal to joint plane and to its strike. 150 quartz axes. Contours at 0.5%, 2%, 4%, 5%.

Fig. 4. Orientation of quartz c-axes in megascopically massive granite.

might be nearly at right angles to the dikes and joints; megascopically this parallelism is absent from the Big Horn rocks and petrofabric diagrams of quartz axes in granite (Fig. 4) show girdle patterns and not point maxima.

Most diabase dikes that cut the Tongue River granite mass are vertical and strike east-northeast, approximately at right angles to the trend of the major syncline. The diabases probably follow tension factures formed during the folding of the syncline.

The dismembered dikes, and the successive stages of deformation and recrystallization of the granite mass provide a corollary with work by Eskola (1949) on the palingenetic and metasomatic processes that were important in the post-instrusion history of many gneiss domes. These processes considerably altered the original internal structures of the intrusion. The Tongue River granite mass probably has undergone a similar "post-granite" transformation which changed its original internal structure.

Microscopic Mineral Alignment in Granite

Thin-sections from an oriented specimen were studied to determine the mineral alignment in megascopically massive granite. The specimen was collected on the western margin of the syncline. Because no s_3 planes were visible in the outcrop, the specimen was oriented with respect to joints that paralleled s_3 in nearby ledges. The orientation of quartz *c*-axes from three thin-sections at right angles to each other is shown in figure 4.

Diagrams A and B, figure 4, represent crossed girdles, with scattered maxima at various places around the girdles, but not at the periphery of the diagram. These diagrams are similar to ones with maxima at *a* and at the points where girdles cut the *bc* plane (Turner, 1948, p. 222). Three mechanisms can yield this type of pattern (Turner, 1948, p. 222): (1) Schmidt called rocks which were deformed by slip along one set of s-planes with alignment of quartz axes parallel to *a*, S-tectonite, (2) Sander and Sahama thought opposed girdles resulted from "crossed-strain", so that *a* becomes an axis of rotation and that the two surfaces formed under dynamic equilibrium, and (3) Cloos thought them due to two differently oriented systems of stress. The mineral orientation observed in the Tongue River granite probably is the result of the last phase of recrystallization and deformation and the crossed girdles are due to crossed strain, as a result of partial rotation about *a*. However, because the granite was probably emplaced during recurring plastic deformation, then Cloos' hypothesis may be applicable to the Big Horn rocks.

Zones of Crushing and Silicification

The Tongue River granite mass is cut by several elongate belts of crushed and silicified rocks, most of which trend east-northeast; a few trend north-northwest. The geometry of these belts suggests that the granite mass was broken into block-like segments at a late stage in the pre-Cambrian deformational history. Crushing was earlier than Cambrian time, because the belts are intensely silicified and epidotized, but no such mineralization is found anywhere in Paleozoic or later rocks. East-northeast trending belts are more common than north-northwest belts because many joints and foliation planes strike northwest; these planes probably provided many closely spaced zones of weakness along which adjustment took place.

The shearing and silicification are younger than diabase dikes. One of the "Leopard Rock" dikes passes laterally into a belt of crushed rock at localities 2-32 and 2-26; where sheared, the dike is weakly mineralized with a little pyrite and pyrrhotite. A few small prospects and one abandoned mine are in the sheared part, but the potential for ore is small.

Structures Involving Paleozoic and Mesozoic Rocks

Sub-Cambrian erosion surface

The western part of the pre-Cambrian Tongue River granite area is a grassy surface of moderate relief. Low, rounded hills, and ridges rise above the surface, and it is cut by numerous stream valleys less than 150 feet deep. West and northwest from the granite area the surface passes beneath horizontal "Deadwood" shales, as at 2-24.

A low ridge of granite and of crushed, silicified granites projects through the "Deadwood" formation at locality 2-22. The nearly horizontal shales surrounding the ridge show no evidence of disturbance. Throughout most of its length the "peridotite" dike of pre-Cambrian age (page 21) is surrounded by horizontal Cambrian rocks showing no trace of contact action. Near the abandoned highway junction of routes Wyoming No. 14 and U. S. No. 14, a low outcrop of granite stands above adjacent buff slabby arkosic sandstone of the "Deadwood". These examples show conclusively that the basal Cambrian rocks were deposited on a surface of moderate relief (Darton, 1906a, p. 106; Curry, 1947, p. 34), and that much of the surface is the present upland.

Hares (1950) has emphasized that most, if not all, of the Rocky Mountain Front Ranges were cut to nearly their present form in pre-Cambrian time, with only an insignificant amount of erosion during Tertiary time, and that the peaks ("pristine monadnocks") were merely "polished up a little" during the later phases. The stripped pre-Cambrian erosion surface in the Big Horn Range very probably is restricted to the gentle upland in the northern part of the granite area and no evidence supports Hares' contention that the high Big Horn peaks were formed during the pre-Cambrian time, because (1) glacial and subaerial erosion were too great and (2) too much Tertiary gravel was deposited along the flanks of the range.

The apparent dip of the stripped sub-Cambrian surface at several localities is shown on plate 2 and plate 8, A. The surface dips gently west over most of the Tongue River granite area, but east of Cutler Creek and Black Mountain it dips northeast, commonly about 20°. Sub-horizontal dikes and joints also show a complementary change in dip east of Cutler Creek and Black Mountain.

Faults and folds

Most faults in the Big Horn region are marginal to the range and are of small displacement. Along the eastern margin, normal and reverse faults strike parallel to the trend of the range, and are vertical or dip steeply northeast toward the Powder River Basin. Notable examples are at Walker Mountain and at Freezeout Point (Pl. 2). Farther south along the eastern margin, the range is overthrust eastward. The best known thrust fault crops out in a highway cut about 7 miles west of Buffalo (Fig. 1, Pl. 8, B); other thrusts crop out on Piney Creek, and Crazy Woman Creek (Fig. 1) (Blackstone, 1949). A thrust fault dips east along the west margin of the range near Five Springs Creek (Wilson, 1934). Previous interpretations of Big Horn Mountain tectonics were based on studies of faulted and folded sedimentary rocks along the margins of the range, with little attention to the pre-Cambrian core. In this paper structures in the core are emphasized.

Freezeout Point area: Freezeout Point is in sec. 3, T. 56 N., R. 88 W., near the northern boundary of the area shown on plate 2. Here a northwest-trending high-angle fault repeats the Big Horn dolomite along Sheep Creek valley. The northeastern side of the valley is the dip slope of the Big Horn dolomite, which also appears near the top of the southwestern slope. The fault plane is visible only in a small area at locality 2-31, where "Deadwood" shales are in contact with Madison limestone. Most of the fault trace is covered by White River sediments of Tertiary age, and the southern end is lost in "Deadwood" shales which show local crumpling near the approximately located fault plane. The southwestern side of the fault is apparently upthrown. Displacement probably does not exceed 1000 feet, as estimated from the approximate thickness of displaced beds.

The rocks dip approximately 5° southwest at Freezeout Point, but about three-quarters of a mile northeast, the nearly vertical Big Horn and Madison beds form prominent northwest-trending ledges; "Deadwood" shales in the intervening slopes are highly contorted and wrinkled. A fault, inferred from these observations probably trends northwest, parallel to the upturned beds and to the fault at Sheep Creek. The fault plane is visible in the stream valley at locality 2-35, at a small outcrop of pre-Cambrian rocks. The northeastern margin of the pre-Cambrian outcrop is less than 200 feet stratigraphically below the base of the upturned Big Horn dolomite, which suggests that some of the approximately 1000 feet of the "Deadwood" formation has been faulted off. These structures were mapped by Demorest (1941) as a single asymmetric anticline, with a steep flank to the west.

Fool Creek Syncline: A northwest-plunging asymmetric syncline involving rocks of Paleozoic age is north of North Tongue River, in secs. 16, 17, 18, T. 56 N., R. 88 W., (Pl. 2). The Big Horn dolomite on the southwest limb of the syncline dips steeper than that on the northeast flank. The syncline broadens rapidly north of the area shown on plate 2, and forms a large valley which slopes northwest to the Little Big Horn River. Shearing caused the thickness of the "Deadwood" formation along the southern margin of the syncline to vary.

Walker Mountain Area: Big Horn and Madison beds on the crest of Walker Mountain (sec. 24, T. 55 N., R. 87 W.) dip about 20° northeast. The mountain is separated from the Big Horn and Madison outcrops on the flank of the range by a narrow strip of pre-Cambrian gneissic granite. The dip slope of Madison limestone on the northeast side of Walker Mountain is in contact with pre-Cambrian rocks along a high-angle fault which strikes generally northwest.

The north end of Walker Mountain is also in fault contact with pre-Cambrian rocks (Pl. 2), as demonstrated by the exceptionally small thickness of the "Deadwood" formation. Numerous springs mark the fault trace along the steep south slope of Wolf Creek canyon. A similar fault at the south end of Walker Mountain was inferred by Demorest (1941, p. 172). A fault shown by Demorest (1941, Pl. 4) between Cambrian and pre-Cambrian rocks along the west side of the mountain, because the basal Cambrian sandstone is missing, probably does not exist; the sandstone is absent because of its lenticular nature.

East of Walker Mountain, the "Deadwood"-Big Horn-Madison sequence is repeated, but is broken again by a high-angle reverse fault at the top of the hogback that is the front of the range. A sharp fold in the Bighorn dolomite in the canyon of Wolf Creek shows clearly that the eastern block moved relatively upward. Displacement along the fault is less than the observed thickness of the Bighorn dolomite. The fault can be traced about 3 miles northwest along the mountain front where it passes into a small fold in the Madison formation (Zakis, 1950, pp. 64-65) (Pl. 2).

Summary: The deformation of the Big Horn Mountains, at least near Tongue River, has the following general features. Marginal fault systems are common, and pass into sharp asymmetric folds overturned toward the west; reverse faults dipping basinward near the mountain front parallel high angle faults a short distance inward. The fault systems, and the axes of folds in sedimentary rocks, are generally parallel; they are similar to the antithetic faults of H. Cloos (1928). The fault systems probably originated by stretching and uparching, caused by upward vertical movement of the basement rocks. Some northwest-trending faults are interesected by cross-faults that trend east-northeast, but no pronounced large-scale cross-faulting can be seen.

Relations Between Structures of Pre-Cambrian and Later Ages

Darton (1906a, pp. 91, 109) described the Big Horn Range as a large anticlinal uplift with subsidiary marginal faults, but later Bucher (1934, pp. 168-171) showed that the asymmetry of folds differed in different parts of the range, and he divided the range into three segments on this basis. According to Bucher, the northern and southern segments resulted from southwestward thrusting and the central segment from eastward thrusting. Demorest (1941, pp. 172-174, pl. 4) concluded that the segments were separated by "lineaments" along which differential movement had taken place. Chamberlin (1940, pp. 673-716) concluded that the lineaments were fracture zones in the pre-Cambrian rocks along which there had been recurrent movement.

The boundary between the northern and central segments was placed at a questionable "Tongue River Fault" (Demorest, 1941, p. 172) within the crystalline rocks. He stated that no good evidence of faulting could be found (1941, p. 172), and suggested that faulting had taken place upon multiple planes as shown by numerous slickensided surfaces. Several zones of shearing and silicification were mapped (Pl. 2), and movement has taken place along these zones. However, available evidence does not indicate a major lineament near Tongue River because: (1) the Paleozoic rocks along the general course of Tongue River are not displaced, and (2) fault systems at Walker Mountain and Freezeout Point are in general similar and parallel, and suggest that the most important faulting in the region was parallel to the trend of the range and not across it.

The boundary between the central and northern segments, during the later phases of uplift, was called the "Cross Creek lineament" (Demorest, 1941, p. 174). This lineament was apparently based upon the following evidence: (1) a well-defined "feature" seen from the air, (2) the slickensided pebbles and boulders in the east-west part of Cross Creek canyon, and (3) the opposing attitudes of the foliation of metamorphic rocks on opposite sides of the valley of Big Goose Creek. The existence of the Cross Creek lineament is questioned on the following grounds: (1) no definite topographic break at the Cross Creek locality is visible upon the most recent topographic maps available[1], (2) slickensided fracture surfaces are widespread in the granites, and are not restricted to the valley of Cross Creek, (3) foliation of the pre-Cambrian rocks commonly is variable (Pl. 2); opposing attitudes in adjacent outcrops are common, and they do not imply faults, and (4) zones of crushed rocks along the headwaters of South Tongue River, near locality 2-26 trend northwest and not east, as does the Cross Creek lineament of Demorest. The feature Demorest saw from the air may be the topographic expression of the change from gneissic to granitic rocks near Dome Lake. No evidence indicates that the Cross Creek lineament passes somewhere near Shell Creek; on the contrary, small faults of a few tens of feet displacement parallel the trend of the range in the canyon, and there is no noticable transverse offset.

The close correlation between attitude of closely spaced primary joints and the trend of the range near Tongue River is evident from plate 3, lending strong support to the hypothesis that the present outline of the range is controlled by pre-existing primary joint planes. The correlation in trend between the major syncline in granitic rocks, the trend of folds in Paleozoic rocks, and the trend of the range, was pointed out by Osterwald (1949, p. 38). The earlier conclusion, however, that the trend of the range was controlled by the attitudes of foliation planes only, is not substantiated by the pole diagram (Fig. 5); the primary joints probably were the major controlling structure. Cloos and Cloos (1934, p. 56) suggested that the range occupies the site of a pre-Cambrian anticline with the same general trend. This suggestion is not substantiated, because the predominent pre-Cambrian structure is synclinal.

Diagram A, plate 3, statistically suggests that the most prominent joint set in the granite trends northwest, although this set is not the most closely spaced (Pl. 3, diagram B). Several belts of crushed rock, and some slickensided fracture surfaces also trend northeast. The granitic mass probably reacted to the Laramide deformation as a group of blocks of various sizes. The outlines of the blocks were determined by pre-existing joint planes and crush zones, along which there was differential movement. Movement along narrow elongate zones, and along many closely spaced fracture planes probably caused the segmentation into blocks. The movement on individual joint planes was very slight and may have been only a fraction of a millimeter, however the aggregate movement along many joints was large. Movement of individual blocks probably was vertical, because the preserved sub-Cambrian erosion surface is nearly horizontal.

Near North Tongue River, at the contact of pre-Cambrian and later rocks, Bighorn dolomite dips steeply to vertically. Nearby crushed granite crops out, but remnants of the nearly horizontal sub-Cambrian erosion surface, about one-half mile south, are covered by a thin veneer of Cambrian shales. These facts suggest that the Fool Creek syncline was caused by downward motion of large block segments north of Tongue River, relative to upward motion of blocks to the south. As the northward blocks dropped, sediments were bent and arched, in adjustment to the underlying crystalline rocks.

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[1]U. S. Bureau of Reclamation, Missouri Basin Project, *Tongue-Powder River Basins Survey*, Preliminary Sheets, 16, 17, 13, 1949. Scale 1/24,000.

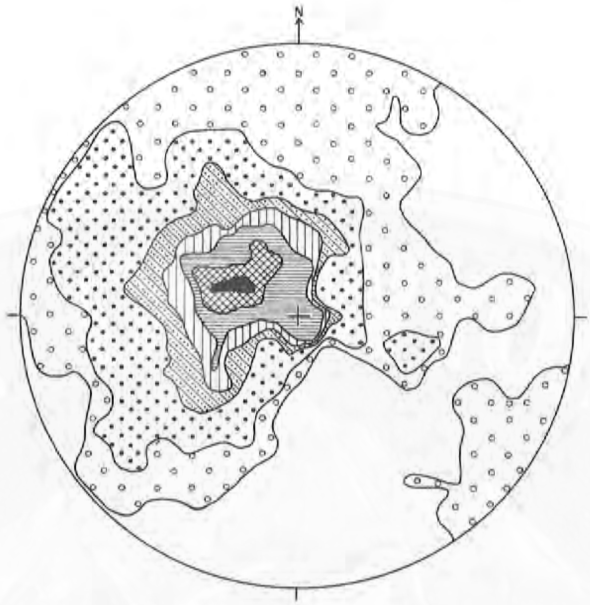

Fig. 5. Eight hundred thirty two s_3 measurements in Tongue River gneiss, plotted on upper hemisphere. Contours at 0.5%, 1.0%, 2.3%, 3.0%, 4.0% and 6.0%.

Marginal fault systems probably had an origin similar to that of the fold. Pre-Cambrian blocks marginal to the range were rotated, as shown by the eastward tilted sub-Cambrian erosion surface along the eastern flank of the range (Pl. 8, A). Rotation was accompanied by high-angle faulting in the Paleozoic rocks at Walker Mountain and Sheep Creek. Reverse faults combined with sharp folds at Walker Mountain and Freezeout Point probably were caused by crumpling and adjustment of sediments to rotation of narrow, elongate, northwest-trending blocks of pre-Cambrian rocks (Pl. 2). Rotation produced northeast dip of originally horizontal planes (dikes and joints) (Pls. 2, 3).

Because layered rocks bend more easily than massive ones, bending along zones where granitic rocks become gneissic may have aided block motion in establishing the present mountain boundaries. The granites in the core probably reacted as a coherent mass, and most movement probably took place where foliation planes were most numerous (Cloos, H., 1933, p. 239), because the present margins of the range coincide with changes to gneissic rocks.

R. T. Chamberlin (Bucher, 1934, p. 187) suggested that the pre-Cambrian rocks of the Beartooth and Big Horn ranges are large uplifted horst-like blocks, originally covered by a folded sedimentary shell; the evidence seems generally to support this contention. Little evidence in the pre-Cambrian rocks, however, indicates that large scale transverse movements were important in the Big Horn Range. The central segment of the range, which has a long gently sloping western side and a precipitous eastern flank with overthrust margins (Thom, 1947; *et al.*), may be the result of the internal structure of the gneisses, which predominate in that segment. The gneisses have strong northeast-plunging lineation and many cross-joints, as well as many other joints that intersect in the lineation direction. During uplift, a slight amount of slipping along the longitudinal joints may have caused eastward crowding and slight overthrusting. Overthrusting along the east flank was not of large magnitude because the Kingsbury gravel was deposited before thrusting (Sharp, 1948, p. 9), and it contains boulders of most of the Paleozoic formations. Thus, to produce the displacement shown by the Clear Creek thrust (Pl. 8, B) the over-riding block need only to have moved a distance equal to part of the stratigraphic thickness of the "Deadwood", Bighorn, and Madison sediments.

CONCLUSIONS

Definite Conclusions

The pre-Cambrian rocks of the Big Horn Mountains are gneisses and amphibolites in the central part of the range. A large granitic mass lies to the north and a smaller mass of similar granite to the south. The gneisses have a strong northeast-plunging lineation throughout much of the range, and locally were isoclinally folded. Granitic rocks in the northern part of the range were folded into a large northwest-plunging compound syncline. Along the southern and eastern margins of the northern granite mass the rocks are gneissic. The granitic rocks are cut by at least three joint sets, one trending northwest, one trending northeast, and one subhorizontal.

The margins of the range, and marginal folds and faults produced during the Laramide Revolution, closely coincide with the trend of the pre-Cambrian syncline, and with the closely spaced northwest-trending joint set.

A stripped sub-Cambrian erosion surface forms a nearly horizontal plain in the northern part of the range.

Probable Conclusions

The nearly flat sub-Cambrian erosion surface suggests that the range was uplifted vertically, with little tilting, except at the margins. The coincidence of trend of the pre-Cambrian syncline, of joints, and of the range suggests a partial control of Laramide structures by pre-Cambrian structures.

Tentative Suggestions

The folding and partial recrystallization of the gneisses was pre-granite in age. Thrust faults along the east base of the mountains in the central segment may be controlled by the strong northeast-plunging pre-Cambrian lineation.

Marginal faults and folds in the northern part of the range may have been caused by differential rotation of rectangular blocks of pre-Cambrian rocks, as the overlying sediments were arched and bent in response to the block movements.

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APPENDIX

TABLE 1

Chemical Composition of Rocks*

1. Caribou Creek biotite metatuff.

	Wt. %
SiO_2	57.78
TiO_2	1.14
Al_2O_3	13.78
Fe_2O_3	4.40
FeO	5.76
MnO	0.05
MgO	3.87
CaO	5.80
Na_2O	2.71
K_2O	2.62
P_2O_5	0.41
CO_2	1.15
	99.47

2. Augen gneiss, from Steamboat Rock.

	Wt. %
SiO_2	67.26
TiO_2	0.69
Al_2O_3	15.38
Fe_2O_3	1.68
FeO	2.52
MnO	0.05
MgO	1.06
CaO	3.29
Na_2O	3.65
K_2O	3.58
P_2O_5	Tr.
CO_2	0.42
	99.58

3. Gray fine-grained gneiss, from South Fork Clear Creek.

	Wt. %
SiO_2	72.32
TiO_2	0.38
Al_2O_3	13.33
Fe_2O_3	1.05
FeO	1.30
MnO	0.11
MgO	0.49
CaO	2.42
Na_2O	4.16
K_2O	3.08
P_2O_5	0.05
CO_2	0.71
	99.40

4. Red granite, from South Tongue River.

	Wt. %
SiO_2	72.37
TiO_2	0.23
Al_2O_3	13.65
Fe_2O_3	0.76
FeO	0.94
MnO	0.02
MgO	0.30
CaO	2.29
Na_2O	3.09
K_2O	4.77
P_2O_5	0.07
	98.49

*Analyses by H. B. Wiik, Helsingfors

TABLE 2

Locations Referred to in Text

Location	Description
1-1	Near the crest of the ridge approximately 1 mi. west-northwest of Powder River Pass.
1-2	Along crest of ridge southeast of Powder River Pass.
1-3	Old highway road cut at Caribou Creek.
1-4	West of West Tensleep Lake.
1-5	Near summit of Sheep Mountain.
1-6	Along the ridge northeast of Powder River Pass.
1-7	Along the ridge 2½ mi. northwest of Powder River Pass.
1-8	¾ mi. northwest of Powder River Pass.
1-9	Contact of pre-Cambrian and Paleozoic rocks in highway cut, 7 mi. west of Buffalo.
1-10	In a highway cut on U. S. Highway No. 16, approximately 1½ mi. east of Powder River Pass.
1-11	Road cut on U. S. Highway No. 16, near The Pines Guest Ranch.
1-12	Road cut on U. S. Highway No. 16, 150 ft. north of East Tensleep Creek bridge, near Meadowlark Lake.
1-13	Fifty feet north of U. S. Highway No. 16, approximately 3½ mi. east of Powder River Pass.
1-14	On the crest of hill 10,399 (Cloud Peak Quadrangle).
1-15	Northern boundary of large schist mass on crest of range near headwaters of Middle Fork Crazy Woman Creek.
1-16	North Fork Clear Creek, near U. S. Highway No. 16.
1-17	In ridge northwest of Seven Brothers Lakes.
1-18	Prominent outcrop cut by U. S. Highway No. 16, ¼ mi. south of South Clear Creek.
2-1	Ridge east of South Fork Tongue River, in sec. 34, T. 56 N., R. 88 W.
2-2	Road cut on U. S. Highway No. 14, southwest of Steamboat Rock in N½ sec. 29, T. 56 N., R. 87 W.
2-3	In the "Box Canyon" of Tongue River, downstream from the abandoned settlement of Rockwood, in sec. 24, T. 56 N., R. 88 W.
2-4	Along the south side of Wolf Creek, north of Walker Mountain.
2-5	Crest of Bruce Mountain.
2-6	About 300 ft. west of the secondary road in NW¼ sec. 7, T. 55 N., R. 87 W., near abandoned log cabin.
2-7	NE¼ sec. 17, T. 54 N., R. 88 W.
2-8	SE¼ sec. 1, T. 54 N., R. 88 W.
2-9	Outcrop along north bank of West Fork South Tongue River, in sec. 9, T. 54 N., R. 88 W.
2-10	Along the ridge east of Prospect Creek bridge on U. S. Highway No. 14.
2-11	Along the west side of sec. 31, T. 55 N., R. 88 W., approximately 1500 ft. west of U. S. Highway No. 14.
2-12	Along the north bank of Owen Creek in N½ sec. 29, T. 55 N., R. 88 W., approximately 1000 ft. east of U. S. Highway No. 14.
2-13	On north slope of Bruce Mountain in NW¼ sec. 26, T. 54 N., R. 88 W.
2-14	At a prominent outcrop east of West Fork South Tongue River, in SW¼ sec. 16, T. 54 N., R. 88 W.
2-15	A conical hill on north edge of the summit of Bruce Mountain, in NE¼ sec. 26, T. 54 N., R. 88 W.
2-16	Along the old logging road ½ mi. southeast of a large marsh in NW¼ sec. 18, T. 54 N., R. 87 W.
2-17	In SE¼ sec. 33, T. 56 N., R. 87 W.
2-18	NE¼ sec. 25, T. 56 N., R. 89 W.
2-19	1½ mi. east of Cutler Creek in sec. 19, T. 56 N., R. 87 W.
2-20	Dike in center sec. 24, T. 56 N., R. 88 W.
2-21	At west end of dike, near small quarry pit, south of Twin Buttes in sec. 25, T. 56 N., R. 89 W.
2-22	On small ridge, approximately 1/3 mi. south of Bear Lodge Resort.
2-23	SW¼ sec 31, T. 56 N., R. 88 W.
2-24	NW¼ sec. 32, T. 56 N., R. 88 W.
2-25	Prospect Pit, the Hope No. 2 Claim in W½ sec. 32, T. 56 N., R. 88 W., owned by H. C. Cooper.
2-26	W½ sec. 21, T. 56 N., R. 88 W.
2-27	SE¼ sec. 31, sec. 32, T. 56 N., R. 88 W.

2-28 South side of road cut on U. S. Highway No. 14, in W½, NE¼ sec. 32, T. 56 N., R. 88 W.
2-29 At road cut on U. S. Highway No. 14, ¾ mi. north of Arrowhead Lodge.
2-30 Near Copper Creek, in SE¼ sec. 21, and SW¼ sec. 22, T. 55 N., R. 88 W.
2-31 On Camp Creek, in SW¼ sec. 5, T. 56 N., R. 88 W.
2-32 On the ridge between North Fork and South Fork of Tongue River, in secs. 28, 29 T. 56 N., R. 88 W.
2-33 A low ridge near Prospect Creek Bridge on U. S. Highway No. 14.
2-34 Sec. 5, T. 56 N., R. 38 W.
2-35 Approximately 1/3 mi. southeast of a cow camp in sec. 34, T. 57 N., R. 88 W.
2-36 Headwaters of South Tongue River, sec. 32, T. 54 N., R. 87 W.

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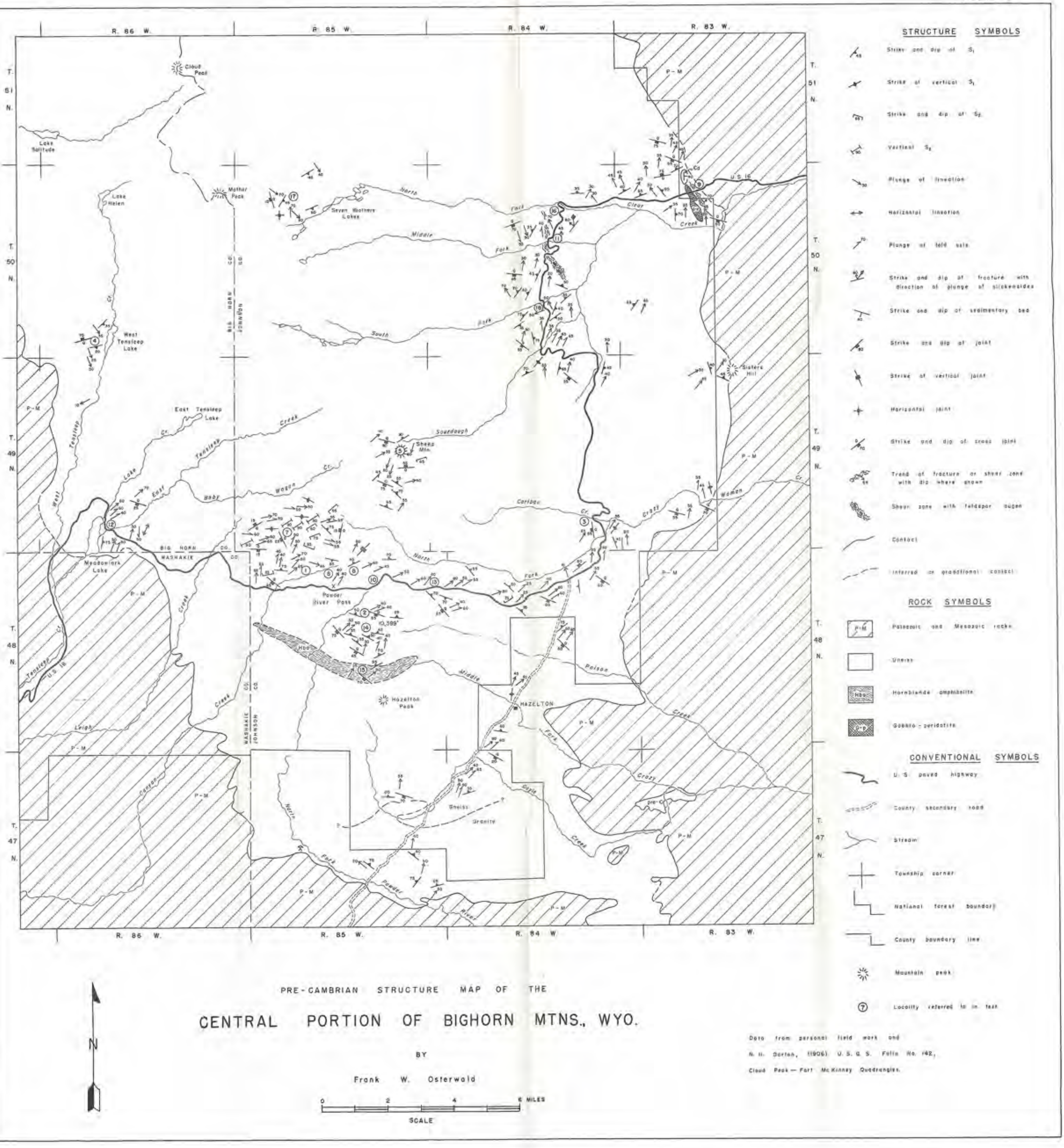


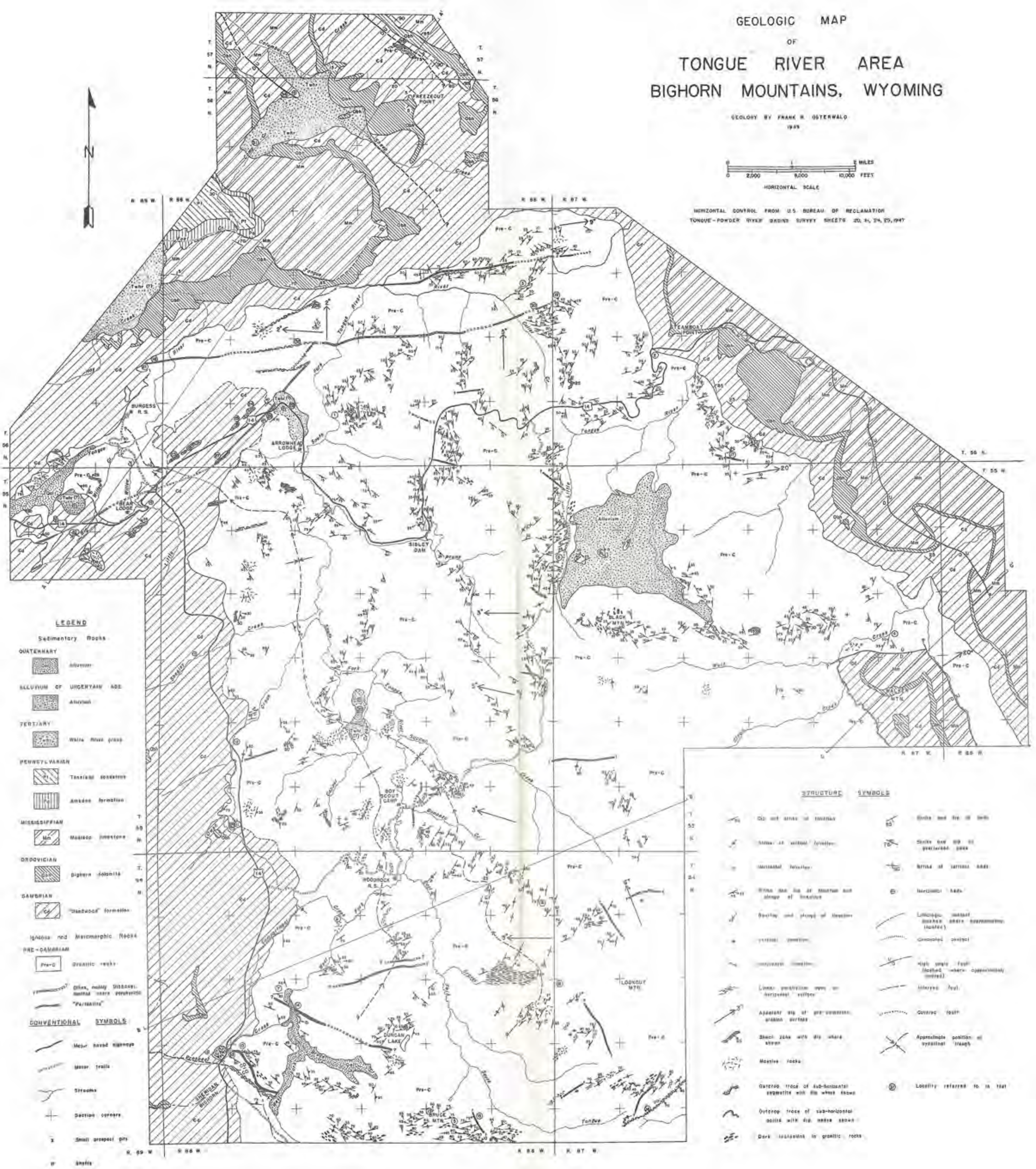
PRE-CAMBRIAN STRUCTURE MAP OF THE

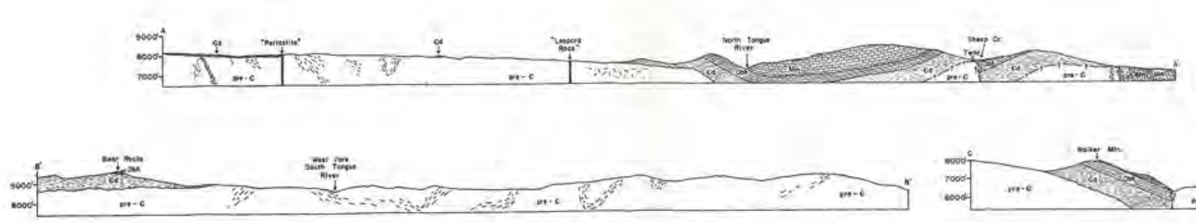
CENTRAL PORTION OF BIGHORN MTNS., WYO.

BY

Frank W. Osterwald

Data from personal field work and N. H. Darton, (1906) U. S. G. S. Folio No. 142, Cloud Peak — Fort McKinney Quadrangles.





CROSS SECTIONS

HORIZONTAL SCALE EQUALS SCALE OF MAP

Plate 2. Geologic map of Tongue River area.

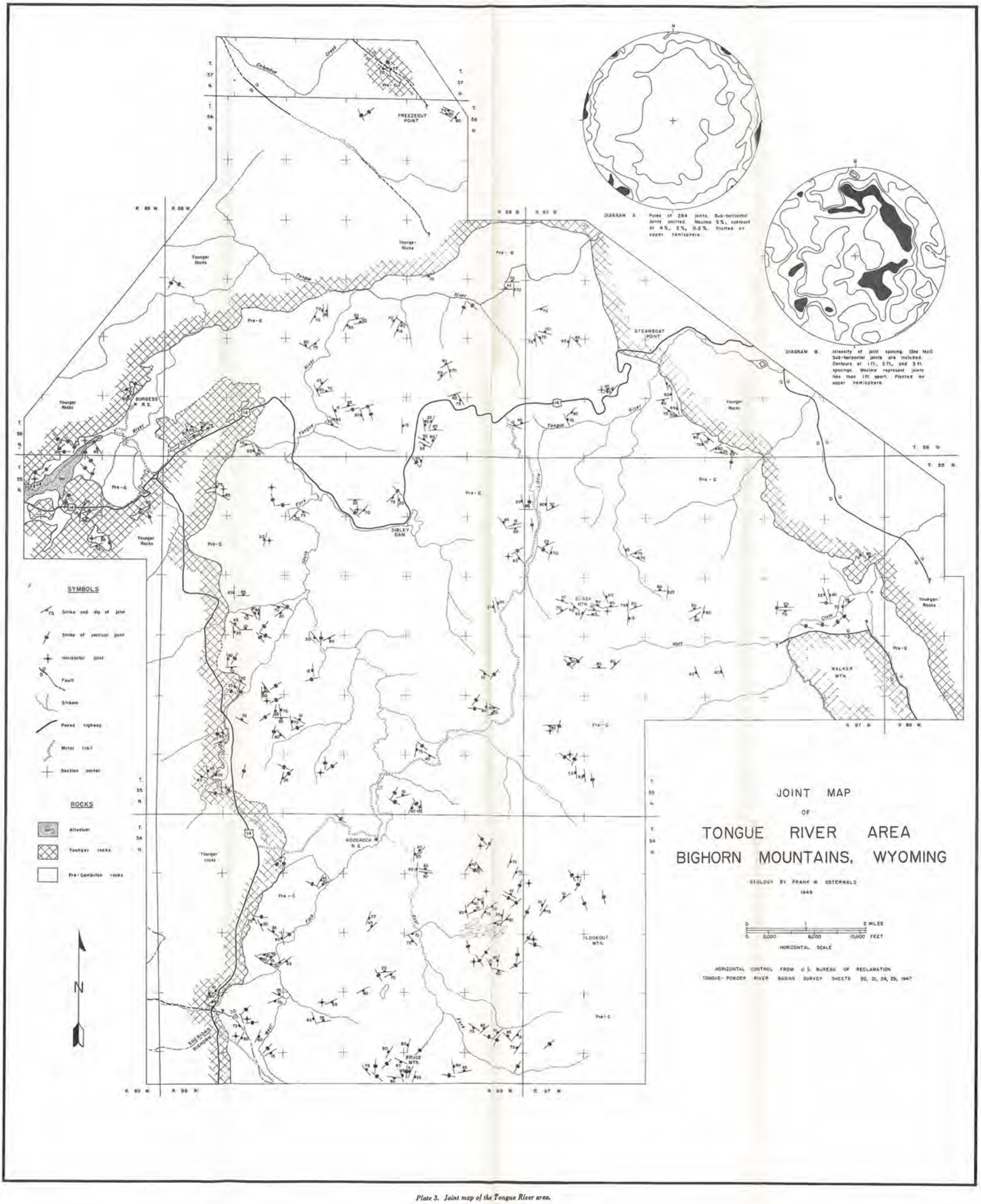


Plate 3. Joint map of the Tongue River area.