Geology of the Berne Quadrangle Black Hills South Dakota

GEOLOGICAL SURVEY PROFESSIONAL PAPER 297-F

Prepared partly on behalf of the U.S. Atomic Energy Commission

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By JACK A. REDDEN

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PEGMATITES AND OTHER PRECAMBRIAN ROCKS IN THE SOUTHERN BLACK HILLS

GEOLOGY OF THE BERNE QUADRANGLE, BLACK HILLS, SOUTH DAKOTA

By JACK A. REDDEN

ABSTRACT

The Berne quadrangle, in the southern Black Hills, S. Dak., is underlain largely by Precambrian metamorphic rocks. Numerous bodies of granitic pegmatite and quartz-rich veins are also of Precambrian age. Paleozoic sedimentary rocks, including the Deadwood Formation (Upper Cambrian), Englewood Formation (Devonian and Lower Mississippian), and Pahasapa Limestone (Lower Mississippian), are present only in the westernmost part of the quadrangle as part of the rim of sedimentary rocks around the Black Hills. Quaternary deposits consist of small amounts of terrace gravel and alluvium.

The Precambrian rocks are separated into two sequences by the Grand Junction fault. The rocks west of the fault comprise, from youngest to oldest, the Mayo, Crow, Bugtown, Loues, and Vanderlehr Formations. The Loues Formation, consisting largely of mica-rich and biotite-garnet-rich schists, and the Vanderlehr Formation, largely made up of amphibole schist, quartzite, metaconglomerate, schist, and calcite-tremolite gneiss, are newly named formations. The total thickness of the five formations is approximately 14,000 feet. East of the fault are four lithologic units whose combined thickness is approximately 5,000 feet. These units include (1) quartzite and quartz-mica schist (2) quartz-biotite-garnet schist, (3) mica schist, and (4) quartzite. The sequence of these lithologic units and their correlation with formations west of the fault have not been fully ascertained.

The composition, textures, and structures indicate that most of the metamorphic rocks were originally sedimentary, but probably amphibole schists in the Vanderlehr and Crow Formations were originally basaltic flows. Parts of the Loues Formation and the quartz-biotite-garnet schist unit were originally banded shales and black shales, and the rest of the Loues Formation and the mica schist unit were uniform high-alumina shales. The lithologic and chemical similarities of the parent material thus indicate that the Loues Formation may be equivalent to these two schist units. The Bugtown and Mayo Formations and the quartzite and quartz-mica schist unit were largely graywacke, subgraywacke, impure sandstone, and siltstone. Thin grunerite-quartz lenses made up a chert-iron carbonate iron formation. Many small bodies of amphibolite were diabase dikes and sills.

Many different types of quartz veins contain such minerals as andalusite, sillimanite, kyanite, staurolite, garnet, feldspar, cordierite, black tourmaline, brown tourmaline, ilmenite, magnetite, wolframite, talc, and gold. Most of the veins are in wallrocks that tend to be rich in the minerals found in the veins. Physical conditions during the formation of most veins and associated metamorphic rocks must have been very similar, but hydrothermal fluids were presumably a factor in the formation of the veins.

Approximately 1,150 granitic pegmatites occur in the quadrangle. These occur mainly in the southeastern and eastern parts near the main mass of pegmatite and granite in the area surrounding Harney Peak. Most of the pegmatites are small and concordant and are without internal structures. Others contain internal units consisting of zones, fracture fillings, replacement bodies, and layers. Pegmatite in the northwestern part of the area is gneissic and has been deformed. The main minerals of the pegamites are plagioclase (albite-oligoclase), quartz, perthite, and muscovite. The common accessory minerals are apatite, tourmaline and garnet; the less common accessory minerals include beryl, biotite, spodumene, amblygonite, graftonite, lithiophilite-triphylite, columbite-tantalite, cassiterite, uraninite, loellingite, allanite, chalcopyrite, chrysoberyl, sillimanite, and secondary uranium minerals.

Early, intermediate, and late structures result in a complex arrangement of the metamorphic rocks in the mapped area. The earliest structures, which delineate the basic structural framework, include a north-northwest-trending major syncline and an adjacent overturned anticline, together with as well as such associated minor structures as folds, lineations, and axial-plane foliation. Structures of intermediate age include a northeast-trending foliation and associated minor folds and lineations. These intermediate structures may have been caused by the emplacement of a large mass of granite and pegmatite in a domal structure around Harney Peak. Late structures are associated with a large dome in the northwestern part of the area and with a smaller dome in the southeastern part. The smaller dome warped and refolded earlier folds and faults; similar refolding of the early structure is inferred around the larger dome. Although some of the youngest structures are postmetamorphic, and the early and intermediate structures formed before or during metamorphism, there is no conclusive evidence that a great difference in time elapsed between the formation of the three groups of structures.

The original sedimentary and volcanic rocks have been converted into medium- to high-grade metamorphic rocks in which many of the primary structures, especially bedding, are excellently preserved. The sillimanite isograd crosses the area and essentially coincides with the boundary of the pegmatite-bearing rocks. Differences in texture and mineralogy on opposite sides of the isograd are small. Certain mineral assemblages and relict textures and the distribution of sillimanite suggest that the rocks have undergone two distinct metamorphisms. The first was more typical of regional metamorphism, and the second, of

thermal metamorphism. The thermal metamorphism probably was caused by the emplacement of large masses of granite and pegmatite in the Harney Peak area and at depth. Metasomatism occurred adjacent to some pegmatites.

Sheet mica, feldspar, beryl, scrap mica, columbite-tantalite, and lithium minerals have been produced from the pegmatite deposits. Discovery of additional sheet mica deposits is not probable. Most of the large feldspar reserves would require beneficiation. Several fine-grained pegmatites contain spodumene of potential economic value. Micaceous pseudomorphs after fine-grained spodumene have a distinctive texture that is useful to geologists searching for additional deposits. Considerable reserves of pure quartz occur in two large veins.

Numerous small gold mines have been abandoned and have apparently been unprofitable. The two largest gold mines, however, are both in places structurally favorable for ore localization and may warrant additional exploration.

INTRODUCTION

The Berne 7½-minute quadrangle in the southern Black Hills, S. Dak. (fig. 121), is underlain by about 95 percent metamorphic rocks, 4 percent sedimentary rocks, and 1 percent pegmatites (pl. 34). The metamorphic rocks are part of the structurally complex Precambrian terrane, largely metasedimentary in origin, that forms the north-northwest-trending core of the Black Hills. Paleozoic sedimentary rocks at the west edge of the quadrangle dip gently westward, away from the core of crystalline rocks. More than 1,100 separate pegmatites occur in the metamorphic rocks of the quadrangle. Pegmatites are most abundant in the eastern part of the quadrangle—on the west flank of the Harney Peak Granite. Some of the pegmaties contain deposits of mica, feldspar, beryl, and lithium minerals; such deposits are numerous in the southern Black Hills and have made this region one of the foremost pegmatite-mining areas of the United States.

The area is easily accessible. U.S. Highway 16 crosses the eastern part of the quadrangle. Gravel secondary roads are common, and lumbering and mining trails cross areas between the secondary roads. The Chicago, Burlington, and Quincy Railroad is nearly parallel to U.S. Highway 16 in the east-central and northern parts of the quadrangle and has a siding at Berne and a freight depot at Custer.

The topography is moderately rugged. Altitudes range from about 5,330 feet along French Creek in the southeastern part of the quadrangle to 7,166 feet on Bear Mountain. Locally ridges and valleys are very steep, but the maximum relief per mile is only about 1,000 feet. Paleozoic sedimentary rocks cap flat-topped ridges and hills along the west edge of the quadrangle and form a prominent cuesta that overlooks the small low-lying hills underlain by metamorphic and igneous rocks to the east. The Danby Park area is relatively high and flat. The topography is most rugged in the northeasternmost part of the mapped area and along the east boundary of the quadrangle.

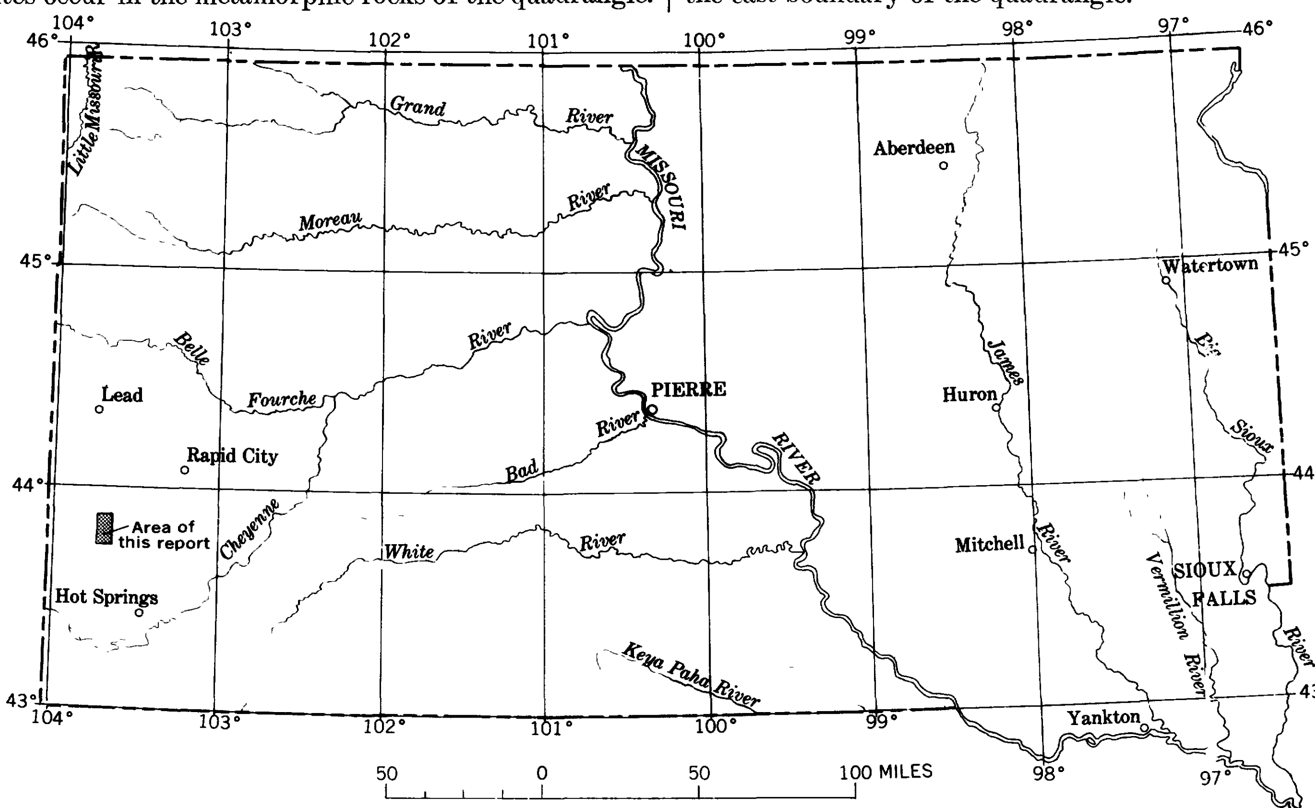


FIGURE 121.—Location of the Berne quadrangle (shaded), South Dakota.

PREVIOUS WORK

In 1875, Henry Newton and W. P. Jenney made the first geologic reconnaissance of the Black Hills (Newton and Jenney, 1880). Van Hise (1890) made additional studies of the structure and metamorphism of the Precambrian rocks. Nevertheless, it was not until 1925, when the "Central Black Hills Folio" by N. H. Darton and Sydney Paige (U.S. Geol. Survey) was published, that a good geologic map on a reasonably accurate base became available. This map, at a scale of 1:125,000, shows few details of the metamorphic rocks.

Since 1925, topical studies and geologic maps of small areas in the Precambrian rocks have been completed by several geologists. Among these publications are noteworthy contributions to the knowledge of the metamorphic geology and ore deposits of the Lead area in the northern Black Hills by geologists of the Homestake Mining Co. (Noble and Harder, 1948; Noble and others, 1949), and papers on several different phases of Black Hills geology by Runner (1921, 1928, 1934, 1943). Runner's 1943 paper specifically discusses many of the geologic problems in the Berne quadrangle. Balk (1931) studies the relationships of the granite and metamorphic rocks around Harney Peak. Ore deposits in the Precambrian rocks have also been described by Darton and Paige (1925) and by Connolly and O'Harra (1929). Other studies of Black Hills geology pertinent to this report are specifically cited in text.

The pegmatites of the southern Black Hills have been studied intensively by the U.S. Geological Survey since World War II. Most of the Black Hills pegmatite mines have been mapped and the geologic data published (Page and others, 1953; Sheridan, 1955). Fisher (1942, 1945) described several of the pegmatite mines. Many papers on individual pegmatites, mineral occurrences, isotopic age determinations, and other phases of Black Hills geology involving the pegmatites have also been published.

The general geology of the central Black Hills (including Precambrian metamorphic rocks) was studied most recently by Redden (1963) in the Fourmile quadrangle, which lies directly south of the Berne quadrangle. In the Fourmile quadrangle, it was possible to divide the metamorphic rocks into stratigraphic units and to study problems of the origin and structural emplacement of the pegmatites. Stratigraphic units defined in the Fourmile quadrangle have been traced into the Berne quadrangle. The geology of about 6 square miles in the southeastern part of the Berne quadrangle, previously described in a preliminary report by Lang and Redden (1953), is modified in the present report.

FIELDWORK AND ACKNOWLEDGMENTS

A. J. Lang, Jr., began mapping of the southeastern part of the Berne quadrangle in 1948. Redden joined Lang for 1 month in 1949 and continued the work during June–December 1950. Redden and Robert Lawthers mapped the southeastern part of the Berne quadrangle in 1952, and Redden mapped the rest of the quadrangle during the summers of 1955–57.

Initial mapping was done on aerial photographs enlarged to a scale of approximately 1:12,000. In 1954 a new topographic base map was published, and post-1954 mapping was done on this new base enlarged to a scale of 1:10,000 but was compiled at a scale of 1:20,000.

About 6 square miles in the southernmost part of the quadrangle was mapped in 1948–49 as a part of the beryllium program of the U.S. Geological Survey on behalf of the Division of Raw Materials, U.S. Atomic Energy Commission.

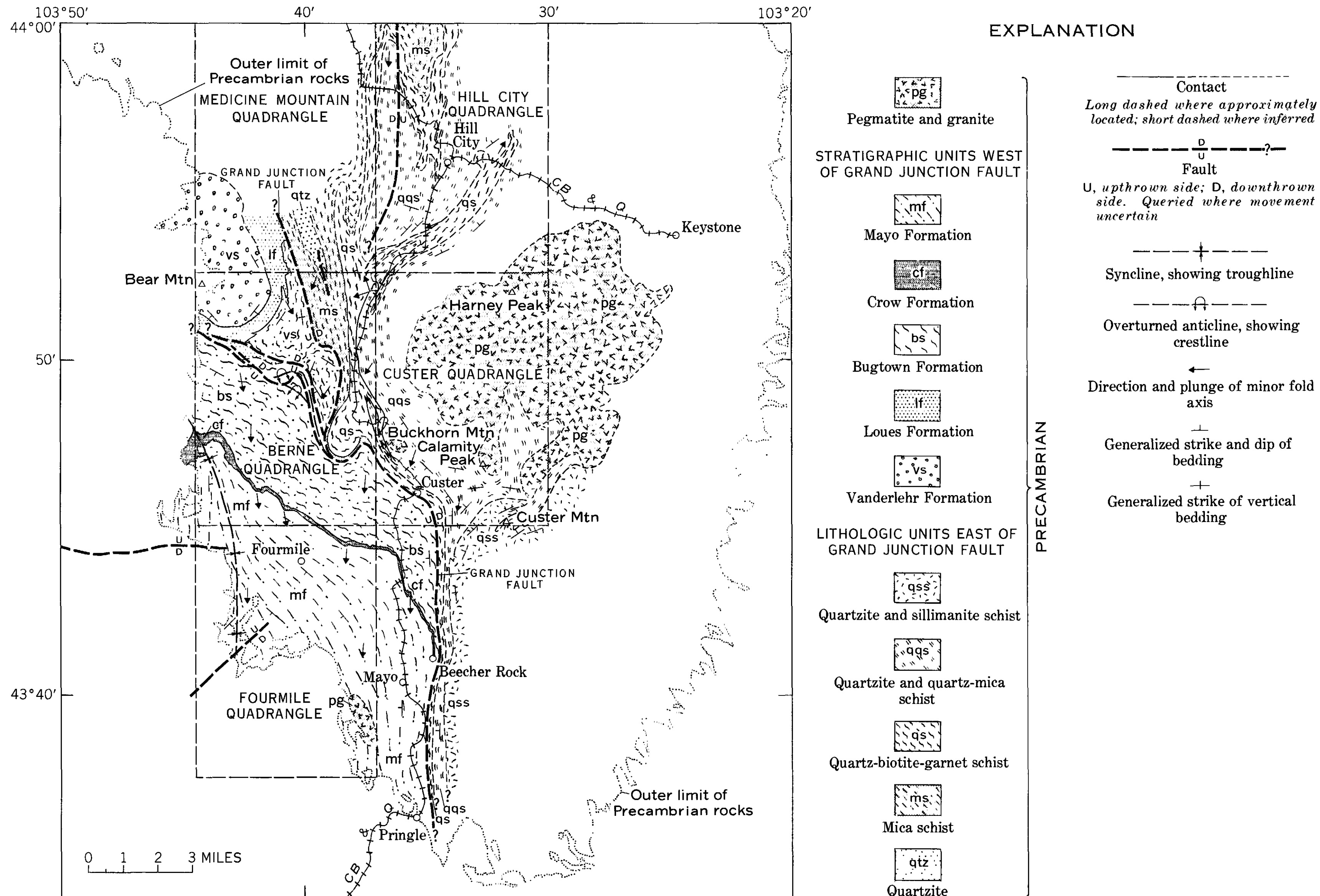
The geologic interpretations in this report, though based mainly on evidence noted in the Berne quadrangle, were made partly as a result of evidence from unpublished preliminary maps of the geology in the Hill City area by R. G. Wayland and in the Keystone area by J. J. Norton.

GEOLOGIC SETTING

The Black Hills uplift is an oval dome about 120 miles long in a north-northwesterly direction and about 50 miles wide. The inner half of the dome consists of Precambrian crystalline rocks and the outer half consists of younger, outward-dipping sedimentary rocks. Several bodies of Tertiary igneous rocks intrude both the metamorphic and the sedimentary rocks of the northern Black Hills. The dome was uplifted in early Tertiary time, and the Precambrian core has been exposed by post-Eocene erosion. The core of the Black Hills consists dominantly of metamorphic rocks (>95 percent); the rest of the core is made up of postmetamorphic granite and pegmatite.

Granite and pegmatite are most abundant in an area of about 40 square miles in the core of the subsidiary, Harney Peak dome, which was named for the highest peak in the Black Hills (fig. 122). Many thousands of pegmatites are exposed on the flanks of this dome.

The Berne quadrangle is west of the Harney Peak dome. The structural effects of the dome are barely perceptible in this quadrangle. However, the stratigraphic units do show a general curvature and flattening of dips near the east edge of the quadrangle. South of the Harney Peak dome near Custer, the metasedimentary rock units seem to diverge—some units trend northwest into the Berne quadrangle and some trend northeast around the east flank of the dome.



Geology of Hill City area after unpublished map by R. G. Wayland. Outer limit of Precambrian rocks and base modified after Darton and Paige (1925)

Figure 122.—Generalized geologic map of part of the southern Black Hills, S. Dak.

Another dome is present in the northwestern part of the Berne quadrangle and in the adjacent Medicine Mountain quadrangle; it is similar to, but smaller than, the Harney Peak dome. This dome, called the Bear Mountain dome by Runner (1943, p. 431), dominates the geology of this part of the area (fig. 122). Westerly dipping Paleozoic sedimentary rocks cover the western part of this dome.

Metamorphic rocks in the Fourmile quadrangle, south of the Berne quadrangle, have been divided into the Bugtown, Crow, and Mayo Formations. These rocks have a combined thickness of about 15,000 feet; they rim a large open syncline which extends northward into the southwestern part of the Berne quadrangle.

METAMORPHIC ROCKS

The metamorphic rocks of the Berne quadrangle are separated into two groups by the Grand Junction fault. Rocks west of the fault include the three formations (the Bugtown, Crow, and Mayo) which were defined in the Fourmile quadrangle by Redden (1963), and two previously undefined underlying formations that are exposed around the Bear Mountain dome (fig. 122). The rocks east of the fault are divided into four major lithologic units. These rocks cannot be readily correlated with rocks west of the fault; however, enough lithologic similarities were noted to suggest that additional mapping in nearby areas would probably demonstrate that the rocks on both sides of the fault are, at least in part, equivalent in age.

Rocks in the center of the Bear Mountain dome cannot be lithologically distinguished from those near the center of the Harney Peak dome. Runner (1943, p. 449) noted the rock similarities and stated that the rocks in the centers of the two domes are equivalent. Grunerite-rich rocks and associated rocks in the lower part of the Bugtown Formation are lithologically similar to certain units on the northeast flank of the Harney dome near Keystone. They also resemble units in the Homestake mine area at Lead, S. Dak. Very similar lithologic units are present in the Keystone area. However, final correlation can be made only after mapping in these areas is complete.

Aside from such speculative correlations as these, the rock units of the Berne and Fourmile quadrangles cannot at present be correlated with other formations in the Black Hills.

STRATIGRAPHIC UNITS WEST OF GRAND JUNCTION FAULT

VANDERLEHR FORMATION

A sequence of variegated rock types consisting of amphibole schist, garnet schist and gneiss, quartzite, meta-arkose, metaconglomerate, calcite-tremolite gneiss, biotite-plagioclase gneiss, and various other metamorphic rocks is exposed in a semicircular area around Bear Mountain. This sequence is believed to contain the oldest Precambrian rocks west of the Grand Junction fault and probably the oldest in the region. It is here named the Vanderlehr Formation after Vanderlehr Creek (pl. 34), which flows across most of the formation. The Vanderlehr Formation extends at least 4 miles north-northwestward into the Medicine Mountain quadrangle (fig. 122).

The type section, and the best exposed section in the area, extends along Vanderlehr Creek from a few hundred feet west of bench mark, 5,821 to a point approximately 1,000 feet north of the northwest corner of sec. 26, T. 2 S., R. 3 S., and thence to the north along a small valley to the contact with the Paleozoic rocks. The upper contact of the formation, which appears to be conformable in the limited exposures, is placed where the amphibole schist is in contact with graphitic schist in an overlying biotite and garnet-rich schist unit. The base of the sequence is not exposed . The maximum exposed thickness of the formation is about 2,000 feet.

The formation has six main lithologies. One of these appears twice in the section, and thus there are seven stratigraphic units, which, from top to bottom, are as follows:

Unit	*Thickness (ft)*
7. Upper amphibole schist	75–165
6. Mica and graphite schists	105–350
5. Lower amphibole schist	100–200
4. Quartz-mica garnet schist	600–700
3. Calcite-tremolite gneiss	250–300
2. Quartzite, meta-arkose, and metaconglomerate	400–530
1. Biotite-plagioclase gneiss	200+

The amphibole schist lithology is shown separately on plate 34 to delineate the structure. The other subdivisions are not shown separately, but all can be recognized and traced several miles northward into the Medicine Mountain quadrangle.

Runner (1943, fig. 3) divided this same rock sequence into six units which correspond in part to the subdivisions just listed. Runner, however, grouped subdivisions 6 and 7 in what he termed "an amphibolitic quartzite," and he treated the intrusive bodies of massive amphibolite (pl. 34), as a separate stratigraphic unit. Also, Runner mapped the lowermost stratigraphic unit as igneous rock, whereas it is actually biotite-plagioclase gneiss containing only a few low-dipping sills of pegmatite.

BIOTITE-PLAGIOCLASE GNEISS

The lowest exposed rock unit in the Vanderlehr Formation is biotite-plagioclase gneiss. The gneiss crops out adjacent to Bear Mountain in an area of very low dips near the center of the dome. Only about 200 feet of section is exposed, and much of this is concealed by debris from the Paleozoic rocks around Bear Mountain. The contact with overlying resistant quartzitic rocks is concealed. The abundance of coarse-grained, originally clastic, rocks in the overlying unit suggests that this contact is unconformable.

The biotite-plagioclase gneiss is gray to brown and of moderately coarse grain size, especially in its biotite flakes, which are 1–5 mm across. Some exposures exceptionally rich in mica would more aptly be called schist. Small lenses of biotite and feldspar impart a gneissic structure to the rock; larger scale bedding or layering is not conspicuous. The major minerals are quartz, biotite, and plagioclase (oligoclase-andesine); however, some specimens are rich in muscovite. Typical modes of the gneiss are given for samples 6–22 and 6–23 in table 1.

QUARTZITE, META-ARKOSE, AND METACONGLOMERATE

Thick-bedded quartzite, meta-arkose, and metaconglomerate form a unit at least 500 feet thick that occurs above the biotite-plagioclase gneiss. The quartzite is most abundant in the upper part of the unit, and meta-arkose and metaconglomerate, in the lower part. All varieties of rock are locally interbedded with each other, and thin beds of biotite gneiss or schist occur between the other thick beds.

Beds of the quartzite average 3 feet thick, but some are as much as 6 feet thick. Some beds are pure white and sugary textured, whereas others are grayish brown and glassy. Thin sections of the grayish-brown types are rich in microcline and plagioclase and have large grains of quartz and feldspar that appear to be relict detrital grains. The mode given for sample 4–12 in table 1 is typical of the grayish-brown quartzites.

The meta-arkose is a pink to tan medium- to coarse-grained gneiss that closely resembles a gneissic granite in hand specimen. It occurs as thick beds, however, some of which contain indistinct crossbedding. The mineralogy of typical samples of the meta-arkose is given in table 1 (samples 7–28, 4–13, 4–14). The rock is almost completely recrystallized, but the shapes of some of the coarser grains of quartz and plagioclase suggest modified and flattened detrital grains. Thinner beds richer in biotite and plagioclase form the less resistant beds between the massive meta-arkose beds.

The metaconglomerate beds are generally interlayered with meta-arkose and, less commonly, with gray-brown quartzite. These layers commonly contain flattened, elongated pebbles of quartz or of sparse feldspar (fig. 123 *A*. *B*) pebbles, the largest are generally less than 1 inch thick, are normal to the foliation, and 2–3 inches wide and 3–4 inches long. Commonly the long axes point in a southerly direction. The pebbles are generally white to gray quartz and make up 5–15 percent of the rock. The matrix is either a gray-brown quartz-rich rock or a coarse-grained quartz-microcline-plagioclase gneiss similar in texture and composition to the meta-arkose.

TABLE 1.—*Modes of rock units*

[Tr.,

	Upper amphibole schist	Mica and graphite schist		Lower amphibole schist			Quartz-mica-garnet schist					
Sample	6–29	6–3A	5–15	6–5	6–3	7–26	6–7	6–36A	7–30	6–36F	6–7A	6–36A
Quartz	10	32	5	19	15		70	75	55	50	10	8
Muscovite		4	84				Tr.		15		45	
Biotite	Tr.	32	5				10	18	15		5	4
Phlogopite												
Plagioclase	15			15	17		10		4		20	20
(Percent anorthite in plagioclase) [1]	(40)			(40)	(35)				(25)		(45)	(42)
Microcline		32	4									
Staurolite							2				5	
Garnet							6	5	10		5	
Cordierite											Tr.	
Sillimanite												
Hornblende	73			65		98				7		60
Tremolite												
Actinolite	Tr.				32							
Grunerite						Tr.				40		
Calcite	Tr.				10					3		
Chlorite	Tr.				12		1	1		Tr.	5	3
Epidote					13					Tr.		
Tourmaline		Tr.							Tr.		Tr.	
Apatite							Tr.	Tr.	Tr.	Tr.		
Zircon								Tr.	Tr.		Tr.	
Magnitite		1										
Ilmenomagnetite										Tr.		2(?)
Opaques			2	1	Tr.	Tr.						
Graphite											5	
Sphene						Tr.		Tr.			Tr.	2
Rutile												

[1] Approximate, where determined.

DESCRIPTION

6–29. Amphibole schist.
6–3A. Biotite-quartz-microcline schist.
5–15. Muscovite schist.
6–5, 6–3, 7–26. Amphibole schist.
6–7, 6–36E, 7–30. Quartz-biotite-garnet schists.
6–36F. Quartz-grunerite gneiss.
6–7A. Muscovite-plagioclase-staurolite-garnet schist.
6–36A. Amphibolite bed or layer.
6–13, 6–13D, 4–10C, 4–10D. Calcite-tremolite gnesiss.

CALCITE-TREMOLITE GNEISS

A unit approximately 275 feet thick and composed mostly of calcite-tremolite gneiss overlies the unit containing quartzite, meta-arkose, and metaconglomerate. Some beds in the unit consist of micaceous gneiss and schist, and others are relatively pure calcite marble. Approximately 60 feet of ledge-forming quartzite occurs near the middle of the unit. Above this quartzite along upper Vanderlehr Creek is the only good exposure of calcareous rocks. Elsewhere the calcareous rocks have been eroded into gentle valleys on each side of the quartzite.

The calcite-tremolite gneiss is generally gray to almost white and forms beds, or groups of beds, as much as 50 feet thick. The rock is coarse grained; white tremolite rosettes and sheaflike aggregates as much as 1 inch in diameter make up 10–50 percent of the rock. The β index of refraction of the tremolite ranges from approximately 1.615 to 1.625. Calcite occurs as a medium- to coarse-grained mosaic of grains in nearly monomineralic layers. Light-brown medium- to coarse-grained phlogopite (β index, 1.600–1.605) is the third most abundant mineral in most samples. Typical modes of the different gneiss samples are given in table 1 (samples 6–13, 6–13D, 4–10D). Accessory minerals in the gneiss include apatite, talc, graphite, and epidote. Talc and epidote occur adjacent to small quartz veins; also talc may be a vein mineral. The small quartz veins commonly contain a few grains of chalcopyrite.

In the steep cliffs along Vanderlehr Creek (pl. 34, C–2) a somewhat schistose amphibolite sill(?) is in contact with the calcite-tremolite gneiss. Adjacent to the amphibolite, the normal, white to gray calcite-tremolite gneiss has been altered to green calcite-actinolite gneiss. Away from the contact the change from the tremolite to actinolite is gradual, over a distance of several feet.

The major minerals in the interlayers, or interbeds, of micaceous gneiss and schist are biotite or phlogopite, plagioclase, quartz, tremolite-actinolite, and calcite; the common accessories are tourmaline, apatite, and zircon. Typical modes are given in table 1 (samples 4–10, 4–10A, 4–10B). Much of the mica is light brown and has relatively low indicies of refraction, indicating that it is similar in composition to phlogopite. The range in composition of plagioclase of different beds is from oligoclase to high andesine. Many of the andesine grains are strongly reverse zoned, so that the outer parts contain about 10 percent more anorthite than the interiors. Cordierite is an abundant constituent of some beds (table 1, sample 6–362). The cordierite is medium grained and occurs as irregular untwinned grains. Tiny inclusions are abundant throughout the cordierite; some are zircon, surrounded by very faint pleochroic halos. Part of the cordierite is altered to chlorite along basal(?) partings.

of the Vanderlehr Formation

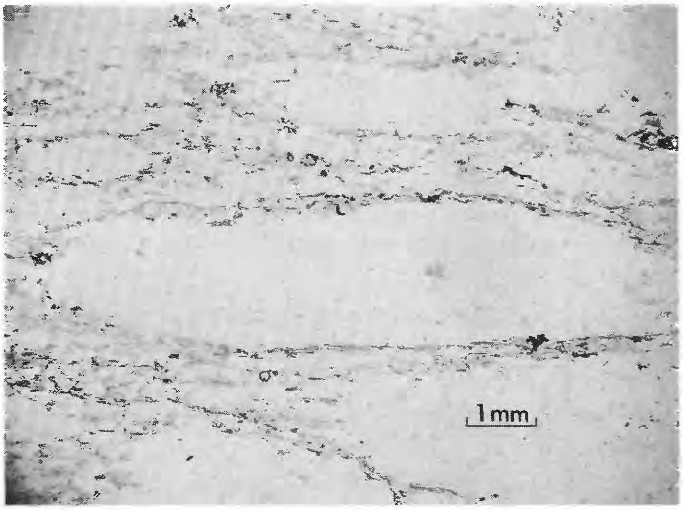
trace]

	Calcite-tremolite gneiss										Quartzite, meta-arkose, and metaconglomerate							Biotite plagioclase gneiss	
6–13	6–13D	4–10C	4–10D	4–10	4–10A	4–10B	6–362	4–11	4–11A	6–13H	4–12	7–28	4–13	4–14	4–15	6–21	6–33	6–22	6–23
	6			18	25	45	27	94	72	66	70	33	61	66	75	58	80	41	44
	5								5	3	1	3	3	15	4	16	8	4	6
				48	45		12	Tr.		4	Tr.	2	Tr.	4	Tr.	4	Tr.	20	10
5	25	30	7			15			6										
	2(?)			30	25	25	27		3	26		15				1	8	35	40
				(40)	(35)	(18)	(12)		(80)	(14)		(9)					(20)	(30)	(16)
						Tr.		5	5	Tr.	25	48	35	15	20	20	4	Tr.	
							Tr.												
							1							Tr.					
							27												
							Tr.												
45	20	10	7																
50	40	60	85	2	5	14			2		3								
	2						5											Tr.	
									6										
				Tr.	Tr.	Tr.	1		Tr.		Tr.	Tr.	Tr.	Tr.	Tr.	Tr.			
		Tr.		Tr.	Tr.	Tr.	Tr.		Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.
				Tr.	Tr.	Tr.							Tr.		Tr.			Tr.	
							Tr.						Tr.(?)						
		Tr.													Tr.		Tr.		
						Tr.	Tr.		1										
													Tr.		Tr.		Tr.		

OF SAMPLES

4–10, 4–10A, 4–10B. Biotite or phlogopite schist.
6–362. Cordierite schist.
4–11, 4–11A, 6–13H. Quartzite.
4–12. Quartzite.

7–28, 4–13, 4–14. Meta-arkose.
4–15, 6–21, 6–33. Metaconglomerate.
6–22. Quartz-plagioclase-biotite gneiss.
6–23. Quartz-plagioclase-muscovite gneiss.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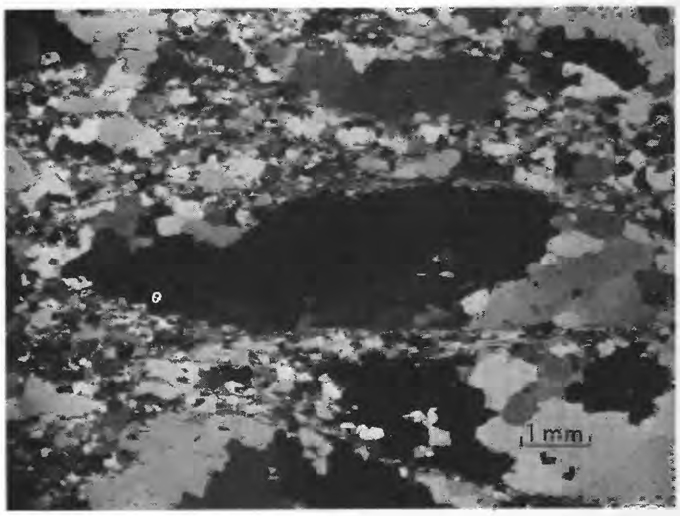
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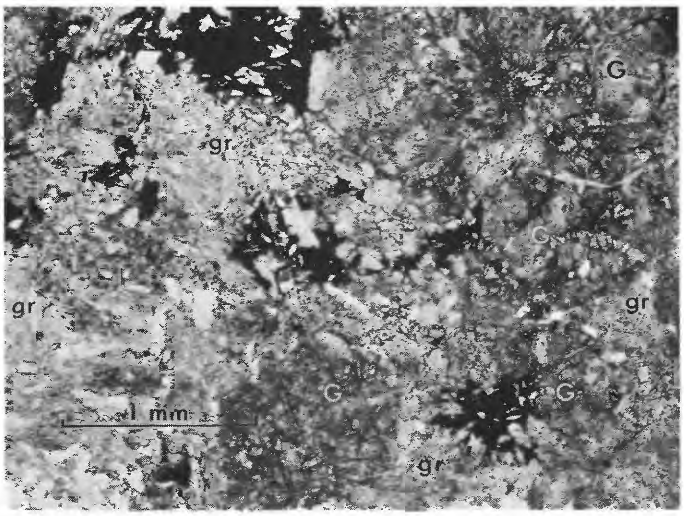
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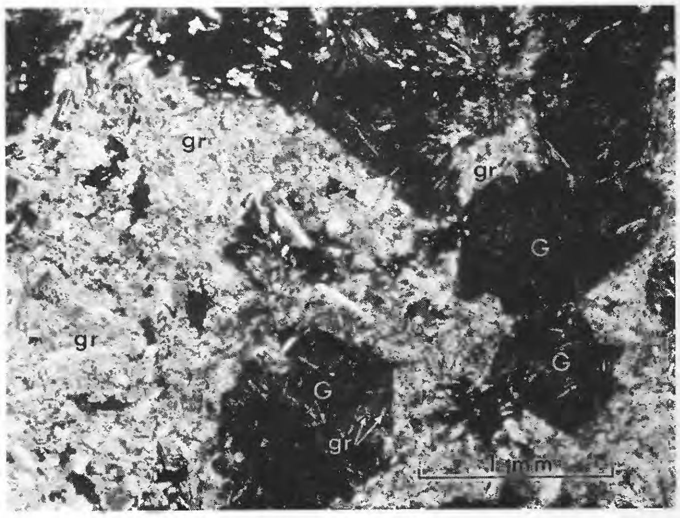
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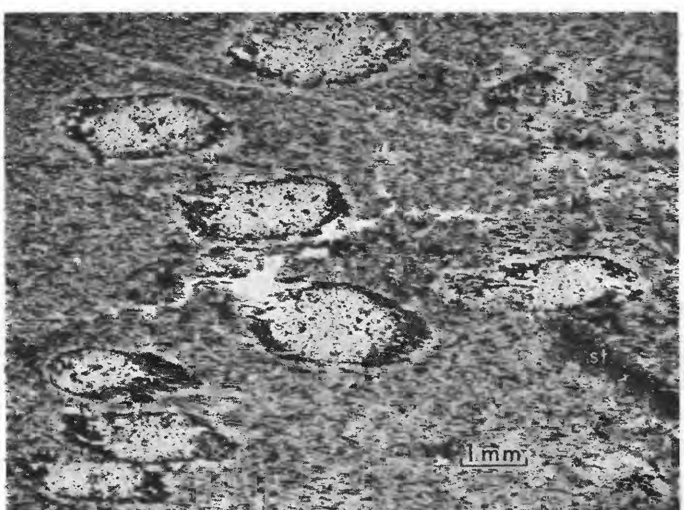
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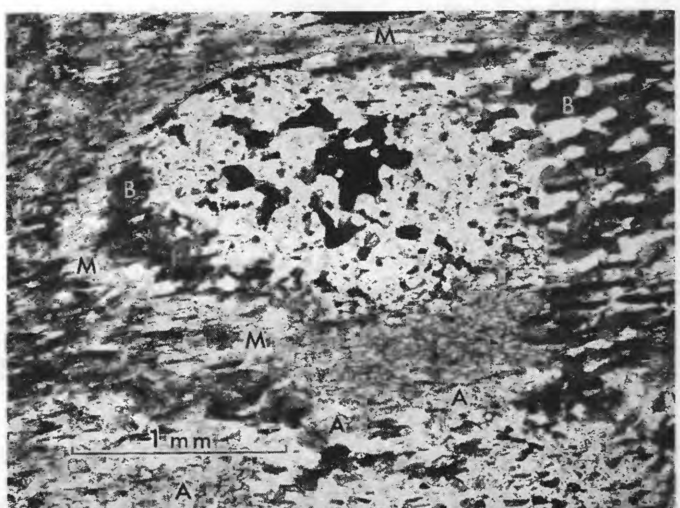
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Quartzite in the calcite-tremolite gneiss unit includes both an impure grayish-brown variety and a very pure white to pink sugary-textured type. Both types occur in beds 1–8 feet thick. The pure variety consists of quartz, a little microcline, and a trace of mica (table 1, sample 4–11). The grayish-brown variety contains plagioclase and mica as the chief impurities but also contains accessory microcline, calcite, epidote, tourmaline, and sphene (table 1, samples 4–11A, 6–13H).

QUARTZ-MICA-GARNET SCHIST

Above the calcite-tremolite gneiss unit is about 650 feet of predominantly garnet-bearing quartz-mica schists that contain a few interbeds of many other types of rock. The unit is best exposed along the east side of White House Gulch; its other exposures are limited largely to float and to small outcrops.

Bedding is conspicuous; most beds are only a few inches thick, but some are as much as 3 feet thick. The most abundant rock type is quartz-biotite-garnet schist, and a few beds consist of biotite-garnet schist, quartzite, quartz-grunerite gneiss, calcite-actinolite gneiss, muscovite-graphite schist, or amphibolite.

The garnet- and biotite-rich schists are generally hematite stained to brownish gray, but thin biotite-rich beds are nearly black. Most of the schist is medium grained. Garnet crystals, however, are as much as 1 cm across, are generally flattened parallel to the bedding, have few crystal faces, and show evidence of rotation and breaking (fig. 124). Most garnet crystals contain many inclusions of the rock matrix. Some oligoclase is generally present in the thicker quartzose beds, and staurolite is sparsely distributed in a few beds. Modes are given in table 1 for two samples (Nos. 6–7, 7–30) of these schists.

The quartzite, quartz-grunerite gneiss, and calcite-actinolite gneiss are generally interlayered in thin iron-stained beds in the upper part of the quartz-mica-garnet schist unit. Total thickness of the interlayers is probably less than 100 feet. The quartzite is generally fine to medium grained and has a distinct sugary texture. The quartz-grunerite gneiss consists of nearly pure layers of sugary-textured quartz and alternating layers

FIGURE 123.—Metamorphic rocks. *A*, Flattened quartz pebbles from metaconglomerate of the Vanderlehr Formation. Quartz, muscovite, and traces of iron ore, biotite, and zircon constitute the rock matrix. Plane-polarized light. *B*, Cross-nicol view of *A*. *C*, Meta-iron-formation from the Grand Junction mine. Grunerite (gr) forms matrix and radial replacement(?) aggregates within garnet (G). Opaque mineral is magnetite. Plane-polarized light. *D*, Cross-nicol view of *C*. *E*, "Spots" in schist of the Bugtown Formation, showing rims of biotite around quartz-rich interiors. Matrix consists of fine-grained quartz, muscovite, and biotite. Single grains of staurolite (st) and of garnet (G) are indicated. Plane-polarized light. *F*, Enlargement of one of the "spots" in same thin section as *E*, showing biotite rim (B) and less apparent outer muscovite-rich rim (M). Interior consists of quartz, opaque material, and wisps of biotite. Andalusite (A) forms poikilitic grains below "spot."

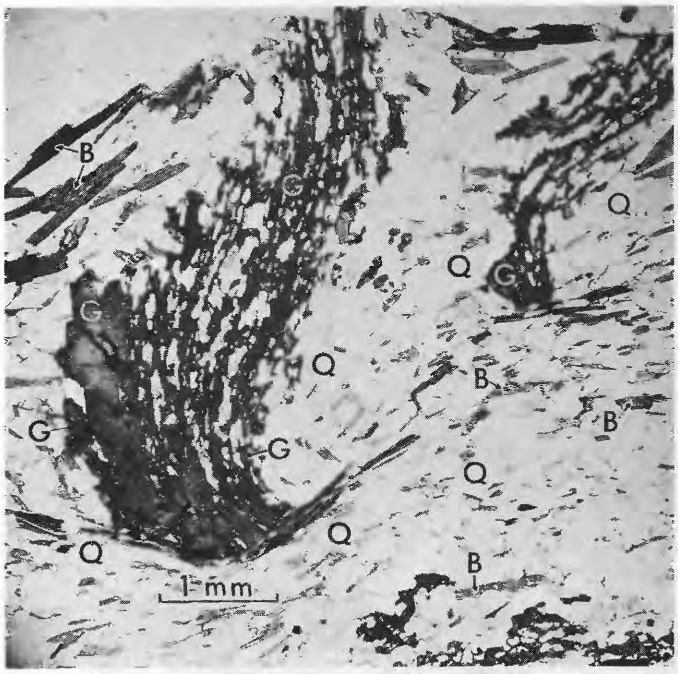
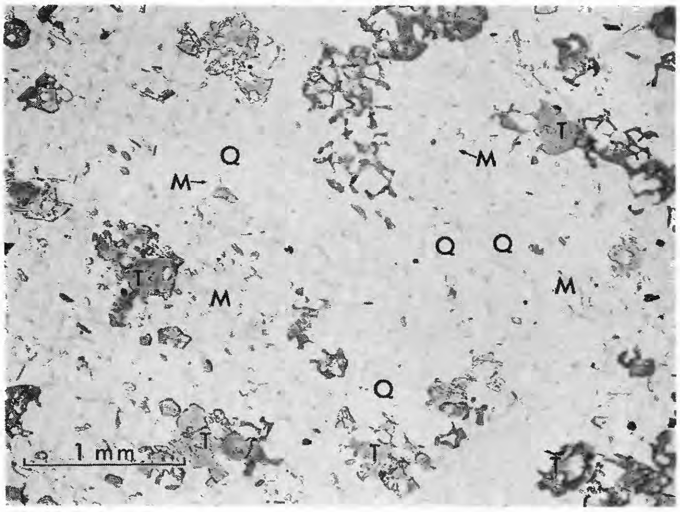


FIGURE 124.—Deformed garnet from the Vanderlehr Formation near Bear Mountain. Skeletal garnet (G) has been rotated more than 90° clock-wise and is elongated subnormal to the foliation indicated by biotite laths (B). The rest of the rock is quartz (Q).

of nearly pure pale-green to brown coarse-grained rosettes of grunerite. The grunerite has a β index of refraction of approximately 1.695; in thin sections it is straw colored, strongly birefringent, and abundantly twinned. Green pleochroic hornblende, commonly associated with the grunerite, has a β index approximately the same as that of the grunerite, but its birefringence is much weaker. Calcite is also a constituent of the quartz-grunerite rock. In some beds the grunerite appears to form in a reaction zone between quartz and calcite; if so, the carbonate probably was originally rich in iron and magnesium. A mode of the quartz-grunerite gneiss is given in table 1 (sample 6–36F).

Muscovite-graphite schist, which occurs only as a few thin beds in the garnet schist unit, is bluish gray, has silky luster, and is fine grained. Some beds contain coarse-grained garnet and staurolite crystals, as in sample 6–7A, table 1. In places these minerals are replaced by muscovite (fig. 125F). The graphite generally is evenly distributed throughout the rock matrix in very small blebs.

Amphibolite occurs as a few thin layers in the cliffs along White House Gulch. It is a mottled coarse-grained rock containing white blebs of quartz and feldspar interspersed between hornblende aggregates. In mineral composition, the amphibolite is similar to the overlying amphibole schists (table 1, sample 6–36A), but the hornblende is much coarser grained and has a sheaflike habit



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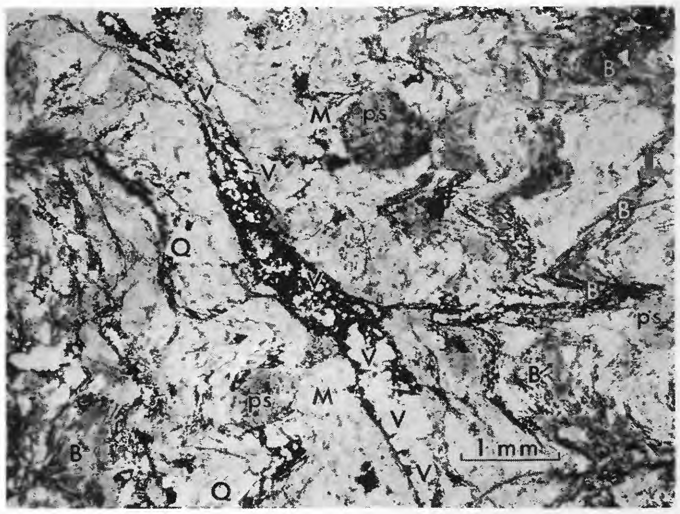

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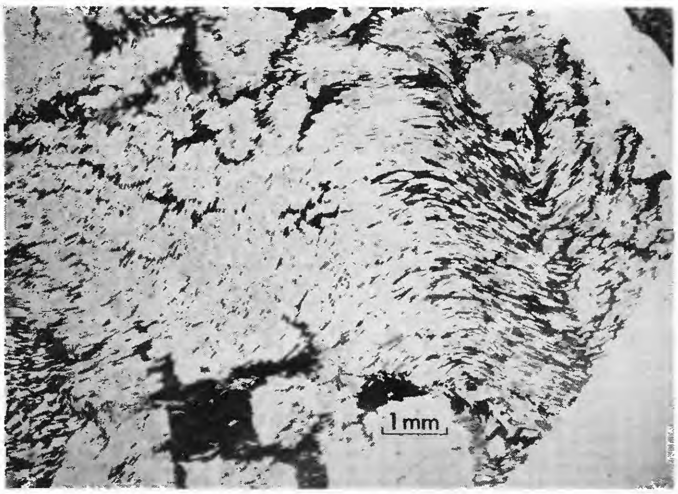

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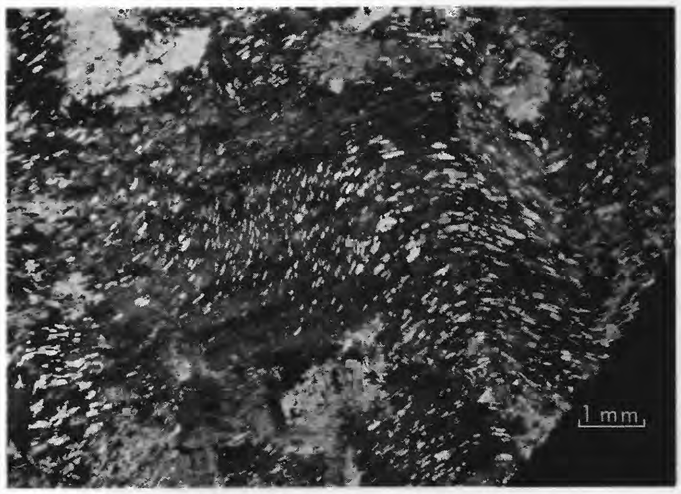

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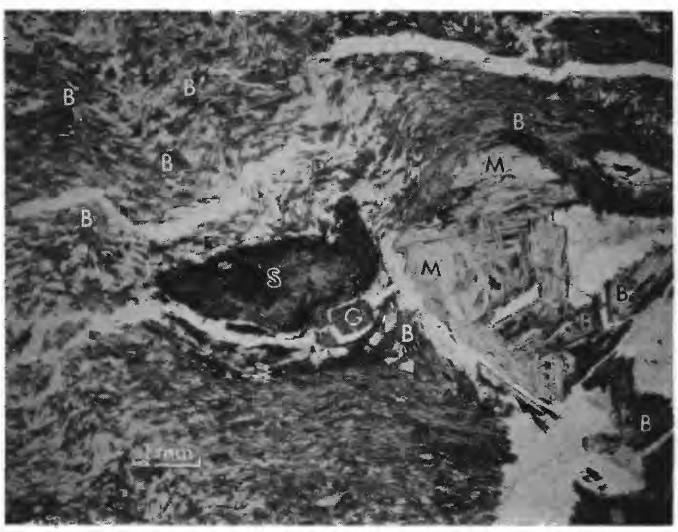

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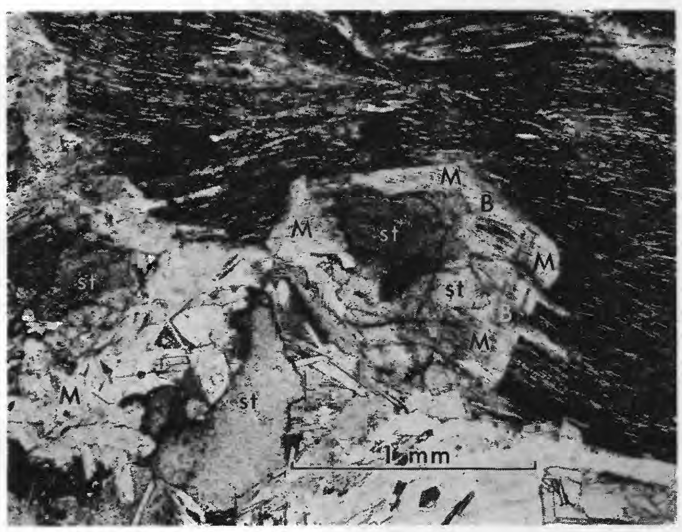

F

quite unlike the hornblende of the amphibole schists. In grain size the rock in these layers resembles the amphibolite bodies shown on the geologic map (pl. 34). Boudinlike segments in these amphibolite layers are fairly common; some beds 3 feet thick neck down and pinch out and then reappear along strike in a distance of only about 3 feet.

AMPHIBOLE SCHIST

Amphibole schist occurs as two separate units that cannot be distinguished lithologically. The lower unit occurs above the quartz-mica-garnet schist and is 145 feet thick. The upper amphibole schist unit is about 180 feet stratigraphically above the lower unit, forms the uppermost part of the formation, and is 180 feet thick. Both units consist almost entirely of fine-grained green to dark-green amphibole schist. A few thin beds of sugary-textured quartzite occur in both units. Excellent exposures of almost 80 feet of continuous section of the lower unit are present just south of the junction of Vanderlehr Creek and White House Gulch.

From a distance outcrops of the amphibole schists appear dark green and massive, but up close they show considerable layering, which is caused by differences in the amount of amphibole. The layers are generally 0.5–1.5 inches thick, but some are 10 inches or more thick. Although most of the layers are uniform in thickness, boudinage is fairly common, and thicker layers neck down and adjacent thinner layers curve inward toward the end of the boudins.

Many outcrops of the amphibole schist contain a low percentage of pebble-sized elliptical aggregrates consisting largely of calcite (commonly removed by weathering) and quartz. Some of these aggregates also contain small amounts of chlorite, biotite, actinolite, and ilmenite. The aggregates are flattened parallel to the schistosity, and their long axes form a distinct lineation.

The schist contains 65–70 percent amphibole; the remainder consists mostly of andesine and quartz in approximately equal amounts. In a few places calcite, chlorite, and epidote are moderately abundant. The accessories in most of the schist are biotite, chlorite, and ilmenite. A typical mode of the amphibole schist is given in table 1; (No. 6–5); the modes of variants of the amphibole schist are also given in table 1 (samples 6–3, 7–26).

Nearly all of the amphibole occurs as needles of green pleochroic hornblende (β=1.675) 1–3 mm long. Most of the long axes are randomly oriented in the plane of the schistosity, but a few of the larger grains form an indistinct lineation. Larger radial grains of actinolite(?) (β=1.655) are common in a few of the calcite-bearing amphibole schists and in some of the quartz-calcite aggregates. A few grains of colorless cummingtonite, found in two thin sections, have a β index of refraction about equal to that of the associated hornblende, but the grains are more abundantly twinned. Some grains have one end consisting of green hornblende and the other end of colorless cummingtonite. Both segments of such grains go to extinction simultaneously under crossed nicols.

Most of the plagioclase and quartz are finer grained than the hornblende, and few single grains are more than 0.1 mm in diameter. The plagioclase is median andesine (Ab_{60}) in composition, and has no prominent zoning or twinning, although some grains appear to be slightly reverse zoned.

The amphibole schist units contain many quartz-tourmaline veins. These veins were not observed in other rocks; thus, the float from these veins was useful in mapping the amphibole schist. Most of the veins are several feet thick, but some are only a few inches thick. They consist of about 15 percent brown tourmaline in a matrix of gray to white quartz. The short, stubby tourmaline crystals are generally 1–2 mm in diameter, but some are much larger. Many crystals have been bent or broken, and the cracks have been healed with quartz. Some weathered parts of the veins are iron stained and cellular, as gossan, and the unweathered veins probably contain pyrite or other iron-bearing minerals.

MICA AND GRAPHITE SCHIST

The mica and graphite schist unit consists largely of nonresistant micaceous and graphitic schists. Its average thickness is 180 feet, but wide variations occur. The limited exposures indicate that both the upper and the lower parts of the unit consist of graphite-bearing schist containing thin beds of sugary-textured white quartzite.

FIGURE 125.—Replacement textures of metamorphic rocks. *A*, Altered quartz-mica-feldspar schist from the upper part of the Bugtown Formation. Equant grains containing a few tiny dark inclusions are quartz (Q), and the clear interstices are microcline (M). Poikilitic gray tourmaline grains (T) enclose rounded blebs of quartz. Plane-polarized light. *B*, Altered garnets from the quartz-biotite-garnet schist unit near the Park school, showing pseudomorphs (ps) of fine-grained muscovite and chlorite after garnet. Late crenulations are outlined by biotite (B) laths. Rest of rock is quartz (Q) and microcline (M). A limonite-quartz veinlet (V) cuts across the section. Plane-polarized light. *C*, Altered quartz-biotite schist from the quartz-biotite-garnet schist unit. Large outlined microcline porphyroblasts have enveloped structure of the late crinkle folds. Dark grains are biotite. Plane-polarized light. *D*, Crossed-nicol view of *C*, showing light-colored quartz inclusions that outline the original rock structure. *E*, Sillimanite schist from the quartzite and quartz-mica schist unit. Sillimanite aggregates (S) deformed along crenulations marked by biotite (B). Euhedral muscovite (M) and biotite aggregate in right center outlines a replaced staurolite grain. White matrix is quartz and some untwinned oligoclase. (Garnet (G) adjacent to sillimanite was broken during thin section preparation.) Plane-polarized light. *F*, Muscovite-staurolite-graphite schist from the Vanderlehr Formation. Staurolite (st) forms relict core of larger zoned outline now replaced by muscovite (M) and biotite (B). Tiny blebs of graphite darken the fine-grained blades of muscovite and form the dark bands in the replaced staurolite. Plane-polarized light.

Hematite stain is common in weather float, and sulfide minerals probably occur in the fresh rock.

Micaceous schist is abundant in the middle of the unit. Most of the mica is muscovite, but biotite is dominant in places. (See table 1, sample 6–3A.) Some beds are rich in microcline, and others contain small staurolite crystals. Two modes of the schist are given in table 1 (samples 6–3A, 5–15). Thin beds containing variable amounts of mica, quartz, feldspar, and garnet are abundant and are a conspicuous feature of the outcrops.

A few thin beds of sugary-textured quartzite and thin beds rich in cummingtonite or grunerite are interbedded with the micaceous schist. Some of these amphibole beds also contain small amounts of calcite.

ORIGIN OF FORMATION

The quartzose gneisses and schists in the two lower lithologic units of the Vanderlehr Formation clearly originated from clastic sedimentary rocks. Most of the gneisses and schists have recognizable relict detrital grains or pebbles, and they also contain accessory minerals—such as zircon and tourmaline—typical of clastic rocks. Garnet-rich schists and associated rocks in the quartz-mica-garnet schist unit have compositions and bedding structures which indicate that they were sandstone and siltstone prior to metamorphism; and they, too, contain accessory minerals typical of clastic rocks.

The high-alumina minerals and the thin bedding in the mica and graphite schist unit suggest that this unit is a metamorphosed shale. Possibly it was black, for it has a high graphite content and probably has sulfides in the fresh rock.

The calcite-tremolite gneiss was apparently derived from carbonate sedimentary rocks. It lacks the common heavy minerals and probably was originally a moderately pure dolomitic limestone. Some of the pure sugary-textured quartzites that are interlayered with the calcite-tremolite gneiss do not contain either the zircon or the tourmaline that is typical of the clastic quartzites, and so were possibly chert layers, rather than sandstone beds.

The origin of the amphibole schist units is less clear. These units are similar in composition to the nearby sills and dikes of amphibolite, which were derived from gabbro or diabase. In texture and structure, however, the schists are so different from the metamorphosed intrusive rocks that the two rock types probably do not have the same origin. Rather, the schists were probably basaltic tuffs and flows which may, or may not, be related to the same magma as the associated amphibolite sills and dikes. A pyroclastic origin is suggested by thin layers that resemble beds in the amphibole schists. The possibility that the amphibole schists are made up partly of lava flows is implied by the presence of quartz-calcite aggregates that may have originally been amygdules. Actinolite, epidote, calcite, and quartz are known to occur in amygdules (Harker, 1939, p. 106–111). The amphibole schists show little resemblance to other amphibole-bearing rocks of the quadrangle that are believed to be metasedimentary.

LOUES FORMATION

Overlying the Vanderlehr Formation is 600–800 feet of uniform and generally well exposed interbedded biotite-garnet and muscovite-rich schists of Precambrian age. The unit is here named the Loues Formation after Loues Creek, which flows along and across the formation south of Bear Mountain. The type section is along the north boundary of the quadrangle, where exposures are the most nearly complete, although the rocks are folded (pl. 34). Other prominent exposures are in the outcrop belt north of Vanderlehr Creek.

Both the lower and the upper contacts of the Loues are apparently conformable. Neither contact is exposed, but both can be approximately located within a few feet. The upper contact is placed where dark thin-bedded biotite-rich schists of the Loues Formation are adjacent to overlying thick-bedded gray quartz-mica-feldspar schists.

In the narrowest and apparently least folded section, due west of Junction Ranger Station (pl. 34, D–3), the thickness is approximately 600 feet. At the type section, along the north boundary of the quadrangle, the outcrop is about 0.7 mile wide. At this locality and others where the formation is folded or faulted, or both, the true thickness may be no more than 600 feet.

Although the Loues Formation is shown as a single unit on the geologic map (pl. 34), it can be roughly separated into three subdivisions. The lower ½–⅓ of the formation is made up predominantly of thin- and well-bedded quartz-biotite and biotite-garnet schists that contain minor amounts of several rocks types; the middle third of the formation consists of more massive mica schists; and the upper part is similar to the lower part of the formation.

The thin-bedded schists in the lower part of the formation are grayish black and are commonly hematite stained. The thinner beds, 0.1–1 inch thick either contain as much as 50 percent of reddish-brown to black garnet in a biotite-rich matrix or consist almost entirely of biotite. The thicker beds consist largely of quartz and biotite and variable amounts of plagioclase and microcline. A few beds are very rich in microcline. The mode of sample 6–26, given in table 2, is typical of some thicker beds.

About 20 feet of porous black graphitic schist occurs

TABLE 2.—*Modes of rocks of the Loues Formation*

[Tr., trace]

Sample	6–26	5–10	5–12A	5–12B	5–11A	5–12	5–13B	5–7
Quartz	59	27	44	35	----	Tr.	4	31
Muscovite	3	----	2	3	5	----	----	18
Biotite	33	52	48	50	24	----	----	15
Plagioclase	----	10	----	----	5(?)	----	----	15
(Percent anorthite in plagioclase) [1]	----	(16)	----	----	----	----	----	----
Microcline	5	5	5	5	50	----	----	----
Garnet	----	5	----	5	----	----	----	5
Cordierite	----	----	----	----	15	----	----	Tr.
Staurolite	----	----	----	----	----	----	----	10
Tremolite	----	----	----	----	----	50	----	----
Actinolite	----	----	----	----	----	----	48	----
Calcite	----	----	----	----	----	49	48	----
Chlorite	----	Tr.	----	1	----	----	----	4
Tourmaline	Tr.	Tr.	Tr.	Tr.	----	----	----	Tr.
Rutile	----	----	----	----	Tr.	----	----	----
Sphene	----	----	----	----	Tr.	----	----	----

[1] Approximate, where determined.

DESCRIPTION OF SAMPLES

6–26, 5–10, 5–11A, 5–12A, 5–12B. Quartz-biotite, rutile-garnet, and microcline- and cordierite-bearing schists from the lower part of the Loues Formation.

5–12, 5–13B. Calcite-tremolite and calcite-actinolite gneiss from the lower part of the Loues Formation.

5–7. Quartz-mica-staurolite schist from the middle part of the Loues Formation.

at the bottom of the lower thin-bedded schists and forms the basal part of the formation. South of Bear Mountain this graphitic schist is overlain by a bed of iron- and graphite-stained sugary- to glassy-textured quartzite 2–10 feet thick. In the overlying 90 feet of section, fine-grained biotite-rich schist beds are interlayered with thin biotite-garnet beds and a few thin layers, or beds, of calcite-tremolite or calcite-actinolite gneiss. The schistose interbeds are commonly rich in microcline, as well as in quartz and mica; some also contain cordierite. Modes of these rocks are given in table 2 (samples 5–10, 5–12A, 5–12B, 5–11A, 5–12, 5–13B).

At the north end of the quadrangle, a bed of sugary- to glassy-textured quartzite with graphite and iron stains occurs about 100 feet stratigraphically above the lower contact. This quartzite bed may be equivalent to the lowermost quartzite to the south. The bed can be traced for more than 1 mile to the north and is more than 30 feet thick in a few places. It contains some lenses of iron-stained carbonate, and the carbonate is commonly separated from the quartzite by a grunerite-rich layer.

The middle, mica-schist part of the Loues Formation is predominantly an alumina-rich rock that locally forms prominent silvery-gray outcrops, which appear massive from a distance. The schist has thin layers or laminae of biotite and chlorite, which may be relict bedding, and a few garnet-rich beds 0.5–1 inch thick. The dominant rock type is medium-grained mica schist that consists of 35–70 percent micaceous minerals. Other constituents in the rock are quartz, plagioclase, and various aluminum silicate minerals (table 2, sample 5–7). Muscovite, or a white mica, is the most abundant micaceous mineral, but biotite and aggregates of chlorite form 10–20 percent of the rock. Locally staurolite and cordierite form in porphyroblasts 1–2 cm long. The staurolite crystals are euhedral and are fresh, but the cordierite forms black pointed rodlike grains, which are commonly altered to mica and chlorite in their outer parts or throughout the grain. Many of these rodlike cordierite grains have a southward orientation of their long axes parallel to fold axes. Small garnet crystals are found in most specimens of the mica schists, but seldom does the garnet (or for that matter the staurolite and cordierite) make up more than 5 percent of the rock. The thin garnet-rich interbeds contain as much as 75 percent of dark-brown garnet. These garnet-rich beds are tan to brown and are lighter colored than the other garnet-rich beds of the formation.

In the area south and west of Vanderlehr Creek, the mica schists in the middle part of the formation are less massive, lack staurolite and cordierite, and contain more quartz and less mica than the schists to the north. The mica is coarser grained, however, and the rock is more schistose than that farther north. The change in texture is gradational along strike.

A few thick beds of gray fine- to medium-grained quartz-mica schist occur in the mica schists near the middle and in the upper part of the Loues Formation. These quartz-mica schist beds contain less than 30 percent mica and almost no chlorite. The major minerals present are quartz, muscovite, and biotite; a small percentage of plagioclase is also present. Some of the thicker quartz-mica schist beds contain small ellipsoidal masses of calc-silicate gneiss. The long axes of the masses are commonly parallel to the fold axes, and the short axes are perpendicular to the most prominent foliation. Although variable in size, most of the ellipsoids are about 4×10 inches in their thickest cross section, and about 4–5 feet long. One ellipsoid, collected 0.6 mile south of bench mark 6,293, was 7×13 inches in cross section and more than 4 feet long. The ellipsoids consist of such calcareous minerals as diopside, plagioclase (bytownite-anorthite), epidote, calcite, and andradite-rich garnet, as well as quartz, microcline, hornblende, and biotite. The hornblende and biotite generally occur as a well-formed outer layer around a core of the calcium-rich minerals. In some specimens calcite forms the core of the ellipsoids. Ellipsoids in similar host rocks in the adjacent Fourmile quadrangle have been more fully described (Redden, 1963). Runner and Hamilton (1934) described ellipsoids of several areas in the Black Hills and concluded that they are metamorphosed calcareous concretions. These ellipsoids are similar to the metaconcretions in Tennessee described by Emmons and Laney (1926) and in Ontario by Pettijohn (1940). Pettijohn noted that the metaconcretions

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occur mainly in the thicker, originally sandier beds, just as they do in the Berne and Fourmile quadrangles.

The upper thin-bedded biotite-garnet schist part of the Loues Formation is very similar to the lower part of the formation but is only about 100 feet thick in most of the area. It evidently thins to the north and is only a few feet thick at Vanderlehr Creek. The exposures of the Loues Formation shown on plate 34 (in secs. 35 and 36) consist almost entirely of these well-bedded schists. The schists are dark gray to black and thin bedded and are rich in biotite and garnet. Locally, thick beds of lighter colored more massive quartz-mica and quartz-mica-feldspar schists are interbedded with the darker thin-bedded schists. These more quartzose schists commonly contain ellipsoids of calc-silicate minerals.

The composition and structures of schists in the Loues Formation indicate that they are metamorphosed shales. Modes of mica schists from the middle part of the Loues Formation suggest that the schists contain as much as 20 percent or more aluminum oxide (Al_2O_3). The thin-bedded lower part of the formation is rich in aluminum, but also has more iron and magnesium than the middle part. Some of the thin beds contain sulfides and graphite, which indicate that the lower part of the formation was originally a black shale. The few sugary-textured quartzite and carbonate beds near the base of the formation probably originated as chert and iron- and magnesium-bearing carbonate beds. The upper part of the formation was a more nearly uniform shale containing a few silty beds and some iron- and aluminum-rich beds that formed the more massive quartz-mica schist and the thin garnet-rich beds.

BUGTOWN FORMATION

The Bugtown Formation of Precambrian age was originally named and described in the Fourmile quadrangle (Redden, 1963), but the name was derived from the excellent exposures along Bugtown Gulch in the Berne quadrangle. As originally defined, the lower contact of the formation occurred where massive quartz-mica-feldspar schist typical of the Bugtown was in contact with unnamed garnetiferous schists. The Bugtown Formation is here redefined to include these garnetiferous schists and additional quartz-mica-feldspar schists, as well as other rock types. The contact between the Bugtown Formation and the underlying Loues Formation is placed at the first appearance of massive quartz-mica-feldspar schists above the dominantly thin-bedded micaceous schists of the Loues Formation. The upper contact of the Bugtown Formation is at the base of the amphibole schists of the Crow Formation.

The additional rock units included in the Bugtown Formation increase the thickness of the formation from

FIGURE 126.—Outcrop of the Bugtown Formation. Upper quartz-mica-feldspar schist unit in the SE¼ sec. 19, T. 3 S., R. 4 E. Note tight folds and thick beds that become thinner on limbs of the folds. Prominent schistosity is parallel to the axial planes of the folds.

an estimated 4,000 feet to possibly 7,000 feet. The lower part of the formation is so intensely folded and deformed, however, that the estimated thickness may not be at all representative of the true thickness.

The type section of the revised Bugtown Formation is along North Bugtown Gulch and northeastward to Atlantic Hill (pl. 34). Some of the better exposures of the lower part of the formation extend northward from the Saginaw mine to the Loues Formation.

Outcrops of the Bugtown Formation are generally plentiful and prominent and are composed of medium- to thick-bedded schist (fig. 126). The most quartzose schist is widely exposed on the higher hills, but the parts of the formation rich in ferromagnesian and aluminous minerals are exposed, though not as well, in and along the valleys.

The Bugtown Formation is divided into lithologic units on plate 34 to help indicate the structural changes across the wide outcrop area. The main structural distinction is that between the commonly garnetiferous quartz-mica schists near the middle of the formation and the generally massive quartz-mica-feldspar schists that make up most of the formation. Also, meta-iron-formation occurs as thin discontinuous units within the schist units. The schist units have gradational contacts and a variable composition and probably cannot be traced as separate units in other places in the Black Hills. The main features of the schist units are as follows:

1. Upper quartz-mica-feldspar schist, medium- to thick-bedded, gray to gray-brown. About 1,000 feet above the lower contact, the schist is more micaceous and contains some garnet and very sparse staurolite and graphite in a 400- to 500-foot-thick interval. Ellip-

soids of cali-silicate minerals are abundant in thicker beds. Unit is approximately 5,000–6,000 feet thick.

2. Quartz-mica schist, thin- to medium-bedded, gray to dark-gray; commonly garnetiferous. Contains beds of quartz-mica-feldspar schist, garnet-biotite schist, staurolite schist, meta-iron-formation, and small amounts of other varieties of schist. Sillimanite schist occurs in the sillimanite metamorphic zone. Unit is approximately 700 feet thick.
3. Lower quartz-mica-feldspar schist, medium- to thick-bedded, gray to gray-brown. Contains beds of quartz-mica schist and locally a thin unit of meta-iron-formation. Ellipsoids of calc-silicate minerals are abundant in thicker beds. A thin, apparently discontinuous, lens of quartz-mica schist occurs in the lower part of this unit.

The lower quartz-mica-feldspar schist forms an outcrop belt adjacent to the Loues Formation and underlies Atlantic Hill in the nose of a fold. The quartz-mica schist forms an extensive outcrop belt between the Junction Ranger Station and the Grand Junction mine. This belt extends to the south around a fold nose at Atlantic Hill and to the northwest past the Saginaw mine. The upper quartz-mica-feldspar schist underlies a sizable area around, and to the south of, Graveyard Gulch along the Grand Junction fault and crops out in a 2- to 3-mile-wide belt adjacent to the Crow Formation.

QUARTZ-MICA-FELDSPAR SCHIST

Outcrops of the quartz-mica-feldspar schist are abundant. The rock is typically medium to thick bedded; single beds are as much as 10 feet thick. Large outcrops of massive rock—where bedding is not recognizable—are sparse. A detailed description of this schist was given in the description of the Bugtown Formation in the Fourmile quadrangle (Redden, 1963).

The schist is medium grained and gray to gray brown; it consists of about 65–70 percent quartz, 25 percent mica, and 5–10 percent oligoclase. The muscovite-biotite ratio is about 2:1. Garnet is an accessory mineral in only a very few of the more mica-rich thinner bedded parts of this rock unit. In places the more micaceous schist also contains small light-colored "spots," 2–4 mm across, which are apparently a relict of some metamorphic mineral. Metagrit, consisting of a few somewhat flattened granule- to grit-sized pebbles of quartz and feldspar in a massive quartzose matrix, occurs as either a single bed repeated by folding or a group of separate beds in the lower part of the upper unit.

Some of the thick schist beds contain as much as 5 percent ellipsoids of calc-silicate rock, identical with those described previously in the upper part of the Loues Formation. In several of the more massive outcrops, the distribution of ellipsoids helps to delineate the bedding.

North of the Highland Lode mine (pl. 34, D–9), especially in the southwestern part of section 18 (pl. 34, D–7), altered quartz-mica-feldspar schist is exposed in a few places. Though much of the fabric of the schist is preserved, the mica in the rock is replaced by tourmaline and microcline, which give the rock a sugary texture. A thin section of the rock is shown in figure 125*A*. These outcrops of altered or metasomatized schist are generally cut by a few thin stringers of pegmatite, but the pegmatite forms only a small fraction of the entire outcrop.

A different type of altered schist occurs as scattered outcrops in the upper schist unit in the area north of French Creek and east of the Big Spar No. 1 mine (pl. 34, F–8). This schist contains large microcline porphyroblasts, 1 inch or more in diameter, disseminated throughout thick beds. This type of schist nearly always occurs adjacent to an exposure of pegmatite. The microcline-bearing beds cannot be traced along strike.

QUARTZ-MICA SCHIST

The quartz-mica schist unit is generally thinner bedded, is richer in ferromagnesian minerals, and contains less quartz than the quartz-mica-feldspar schist units. The unit tends to underlie valleys and to be poorly exposed. Interbeds of more massive and quartzose schist are also present and form possibly the only sizable outcrops in areas of sparse exposures. Garnet-rich biotite schists are abundant locally. Scattered beds contain staurolite; sillimanite-bearing schist is present in the area of high-metamorphic-grade rocks. Also present are "spotted" or micaceous schists, a few thick beds of gray granular quartzite, and a very few thin beds of slabby clastic quartzite or metagrit. The uppermost part of the unit is darker and is richer in biotite and garnet, whereas the lower part is lighter colored and more nearly like the quartz-mica-feldspar schist units.

The main rock type is a light- to dark-gray micaceous schist that contains a few scattered garnet porphyroblasts and 30–45 percent biotite and muscovite; the rest of the schist is composed largely of quartz and a small percentage of oligoclase. This rock locally grades into a dark well-bedded garnet-biotite schist containing as much as 40 percent dark-red to almost black garnet porphyroblasts in a biotite-rich matrix. Sulfides are common accessory minerals in the darker beds. Some interbeds contain staurolite porphyroblasts, most of which are about 2 mm long, but some of which are more than 5 mm long. The garnet-biotite schist occurs at several different stratigraphic levels, but it is thickest near meta-iron-formation. Exceptionally wide exposures of the rock occur near the Grand Junction mine (pl. 34, E–3).

At least two distinct beds of brown to dark-gray metagrit, each about 2 feet thick, are locally present. These beds are identical in appearance with the metagrit beds described in the quartz-mica-feldspar schist unit.

The "spotted" micaceous schist generally contains garnet, as well as sparse staurolite, andalusite, and (or) sillimanite. The schist has many disseminated light-colored "spots" rimmed by a biotite-rich layer, as shown in figure 123 *E* and *F*. The light-colored centers are mostly small blebs of quartz and opaque minerals. The spotted schist is found throughout the quartz-mica schist unit but tends to be most abundant near the lower contact, especially in the area southwest of Atlantic Hill.

The sillimanite-bearing schists contain waxy white ellipsoidal aggregates of sillimanite 2 mm–1.5 cm long. The aggregates tend to increase in size to the southeast.

META-IRON-FORMATION

Several thin units consisting largely of alternating grunerite-rich and quartz-rich layers, or, in a very few places, of massive grunerite schist or massive quartzite, occur at various places in the lower half of the Bugtown. This rock is the metamorphosed equivalent of the iron-formation, as defined and described by James (1954, 1955). The units of meta-iron-formation are no more than a few tens of feet thick and are exposed in discontinuous outcrops which are, at least locally, distinct stratigraphic markers in the more nearly uniform schists of the Bugtown.

One unit of meta-iron-formation occurs about 50 feet stratigraphically above the lower contact of the Bugtown, and two other units occur in the quartz-mica schist unit. A few hundred feet southeast of the Penobscott mine (B–6), sparse pieces of meta-iron-formation float occur at a somewhat higher stratigraphic level than the other units; however, because of lack of further evidence to indicate the presence of this unit, it was not shown on the geologic map (pl. 34).

In general the different units cannot be distinguished lithologically from one another, although individual outcrops differ greatly in grunerite or quartz content. The quartzite of the meta-iron-formation is in lenses or beds that are 0.5–1.5 inches thick, sugary to glassy textured, and white to gray. These are interlayered with a brown to green grunerite-rich rock, some of which also contains quartz, hornblende, garnet, magnetite, and apatite. Lenses of iron-rich carbonate, separated from quartzite by a layer of grunerite, occur southwest of Atlantic Hill. The quartzite is sparse in some parts of the meta-iron-formation units, yet in others it forms entire outcrops that contain little or no amphibole. For example, the outcrop about 0.4 mile northeast of the Minnie May mine (pl. 34, F–8) is entirely composed of quartzite.

The grunerite forms medium- to coarse-grained aggregates of either stubby prismatic crystals or semiradial aggregates. The crystals are generally randomly oriented, and the rock is massive. In thin sections the grunerite is colorless to light yellow, strongly birefringent, and abundantly twinned. The β index of refraction averages 1.685. A characteristic feature of many larger grains in thin sections is distinct striations parallel to (001), which are similar to those described and shown by Miles (1943, p. 29).

Most specimens of the meta-iron-formation also contain magnetite, garnet, and a dark-blue-green hornblende. The hornblende is strongly pleochroic (X, light green to yellowish green; Y, dark green; and Z, dark blue green); about one-third of the hornblende is intergrown and optically continuous with the grunerite. In many of the intergrowths the blue-green hornblende forms only the tip of the grunerite grains. The β index of refraction of the hornblende nearly equals that of the grunerite but may be as much as 0.005 lower. In a few samples the hornblende also has striations parallel to (001). The garnet in the meta-iron-formation forms dark-reddish-pink crystals as much as 5 mm in diameter. Some grains are flattened and broken and are partly rimmed by green hornblende.The hornblende seemingly formed somewhat later—partly from alteration of the garnet. In other samples tiny needles of grunerite occur as radial aggregates in the garnet crystals and embay the outer crystal faces of the garnet (fig. 125*C*, *D*).

A few outcrops almost due west of the Junction Ranger Station are rich in actinolite and also contain augite. The augite is generally associated with broken and brecciated quartz-rich lenses of the meta-iron-formation. It consists of large crystals as much as 0.8 inch across which have a β index of refraction about 1.705.

The meta-iron-formation was probably a mixture of chert and iron-rich carbonate minerals prior to metamorphism. The grunerite and other iron-silicate minerals seemingly formed by a reaction between the thin chert beds and the interbeds of iron-rich carbonate.

CROW FORMATION

The Crow Formation of Precambrian age is a thin incompetent poorly exposed unit of diverse rock types that overlie the Bugtown Formation. The Crow forms a narrow outcrop belt in the southwestern part of the Berne quadrangle. The formation consists of gneisses and schists rich in ferromagnesian, carbonate, and potassium minerals. The type locality is Crow Creek in the south-central part of the Berne quadrangle. Because the type section and other exposures of the Crow Formation in both the Berne and the Fourmile quadrangles were fully described in the Fourmile report (Redden, 1963), only brief summaries of the many rock types of the Crow Formation are given in this report.

Both the upper and the lower contacts of the Crow Formation are readily identifiable. The upper contact is placed above a white quartzite bed, and the lower contact is placed below the amphibole schists that overlie quartz-mica-feldspar schist of the Bugtown. The quartzite is well exposed, and the upper contact is easily followed. The map pattern on plate 34 suggests, however, that the upper contact interfingers with the overlying rocks in at least two localities. The uppermost part of each interfingering tongue or lens of the formation is generally a thin quartzite bed, also.

The formation is about 60 feet thick at the south edge of the quadrangle, 150 feet at its type locality, and possibly as much as 1,000 feet thick in the axis of the syncline near the Western Star mine (pl. 34, A–7). The interfingering of the upper contact indicates that some of the difference in thickness was original, but the great thickness in the nose of a major fold probably is largely a result of plastic deformation.

The major types of rock from the bottom to the top of the Crow Formation are (1) amphibole schist, (2) calcite-hornblende gneiss, (3) calc-silicate gneiss, (4) cordierite-biotite schist, (5) microcline-biotite schist, (6) calc-silicate gneiss, and (7) quartzite. Other types of rock, as well as gradations between adjacent types, occur in some exposures.

In the area near the main synclinal fold axis, the Crow Formation consists largely of amphibole schist. The amphibole schist is medium grained and light to dark green and consists largely of hornblende or actinolite and plagioclase; it also contains small amounts of quartz, biotite, and magnetite. The amphibole generally makes up about 60–90 percent of the rock. Near the Western Star mine, thin lenses of rocks rich in cordierite and plagioclase and lenses of calc-silicate rock are interlayered with the amphibole schist. The outcrop 1,600 feet N. 52° E. of the Western Star shaft contains 5–10 percent of ilmenite in coarse-grained plates 0.1–2 inches across. These plates are embedded in a medium-grained matrix of pure actinolite (ε index of refraction = 1.655). Tiny limonite-filled cavities in this same rock contain actinolite crystals and minute rutile needles.

The calc-silicate gneiss commonly occurs as lenses, generally in the upper part of the formation; it is exposed mainly along Crow Creek and near the Western Star mine. Beds and layers of almost pure tremolite exposed near the intersection of North and South Forks French Creek are in the stratigraphic position ordinarily occupied by magnesium-rich calc-silicate gneiss. The two rocks probably have very similar chemical compositions, and the difference in mineralogy may be caused by the difference in metamorphic grade. Thus, the two rock units—though mineralogically different—are probably stratigraphically equivalent.

The calcite-hornblende gneiss and the calc-silicate gneiss consist of thin light-colored layers that are rich in calcite and are interstratified with darker layers rich in hornblende or actinolite, diopside, biotite, and sphene. At the type locality near Crow Creek, the light-colored layers also contain plagioclase (bytownite-anorthite), microcline, and scapolite, and some of the minerals generally present in the dark layers. A few of the light-colored layers are sugary-textured quartzite.

The cordierite-biotite schists are mottled dark gray to brownish gray and are medium to coarse grained. Oligoclase-andesine is locally abundant. The cordierite occurs as equant grains as much as 5 mm across, but most are about 1 mm in diameter. The grains are bordered by biotite blades and tiny grains of microcline.

The cordierite-biotite schist has interbedded units a few tens of feet thick that consist of thin-bedded nearly black fine-grained schist containing microcline, biotite, and muscovite. The rock somewhat resembles slate but is slightly coarser grained. A published analysis of this rock (Redden, 1963) shows that it contains about 10 percent potassium oxide (K_2O).

The bed of quartzite at the top of the Crow Formation is sugary to glassy textured, is generally impregnated with graphite and limonite, and is about 4 feet thick in most exposures. On the noses of folds near the main synclinal axis, it apparently thickens to as much as 30 feet. The quartzite bed is generally well exposed and, locally, where the dip is steep, forms wall-like exposures. Similar, but considerably thinner, beds are present, though sparse, lower in the formation.

Kyanite and traces of sillimanite occur with quartz, plagioclase, and mica in the wide exposure in the southeastern part of sec. 24 (pl. 34, C–8). In this same area green chrome mica, fuchsite, occurs in beds adjacent to the kyanite-bearing schist. Fuchsite was also observed in small outcrops about 900 feet N. 74½° E. from the Western Star shaft (pl. 34, A–7).

Detailed studies of the petrology and structure of the Crow Formation in the adjacent Fourmile quadrangle (Redden, 1963) indicated that the formation is derived from basaltic flows, impure carbonate rocks, potassium-rich shale, and probably chert.

MAYO FORMATION

The youngest of the metasedimentary rocks of Precambrian age exposed west of the Grand Junction fault is a thick sequence of interbedded quartz-mica-feldspar schist, quartz-mica schist, and staurolite- and garnet-rich schists which has been named the Mayo Formation in the Fourmile quadrangle by Redden (1963). The resistant rocks of the Mayo Formation are well exposed in the core of the syncline in the southwesternmost part of the quadrangle. About 4,000 feet of the formation is exposed in the Berne quadrangle, but at least 14,000 feet is exposed in the Fourmile quadrangle (Redden, 1963).

The Mayo Formation can be divided into three generalized lithologic units in the Berne quadrangle, but the boundaries between these are indistinct and, therefore, are not shown on plate 34. The lowermost unit includes nearly 1,400 feet of predominantly thick-bedded to massive quartz-mica-feldspar schist. A few beds of gray staurolite and garnet schist occur about 80 feet above the base of this unit. These beds locally contain pale-green laths of kyanite in the western part of the area, and they also have waxy knots of sillimanite in areas above the sillimanite-zone boundary. Several beds of calc-silicate gneiss were mapped in the Fourmile quadrangle; in the southern part of the Berne quadrangle these beds rapidly lens out and thus are not shown on the geologic map (pl. 34).

The lower unit is overlain by approximately 800 feet of interbedded quartz-mica-schist, quartz-mica-feldspar schist, and staurolite- and garnet-bearing schists, and these in turn are succeeded by about 1,800 feet of quartz-mica-feldspar schist. The upper unit contains a 5–10-foot thick bed of metaconglomerate and metagrit, which crops out about 2,000 feet east of long 103°42′30″ along the south edge of the quadrangle. This bed was traced to the southeast for several miles in the Fourmile quadrangle, but it extends to the northwest into the Berne quadrangle for only about 400 feet.

The quartz-mica-feldspar schists of the upper and lower units are light gray to gray, medium grained, and thick bedded, and similar to the massive schists of the upper part of the Bugtown Formation. In the Mayo Formation, however, the feldspar-bearing schists are somewhat coarser grained and contain a larger amount of relict grains that are about 1 mm in diameter. Calc-silicate ellipsoids are abundant in the more massive beds.

In the middle unit, beds of gray medium-grained quartz-mica-schist alternate with darker beds of porphyroblastic staurolite and garnet schist and beds of quartz-mica-feldspar schist. The beds range in thickness from a fraction of an inch for the dark beds to several feet for the quartz-mica schist.

In addition to these three major units, a few tens of feet of fine-grained actinolite schist and graphite-bearing mica schist occur adjacent to the large quartz vein 2,800 feet east-northeast of the southwest corner of the quadrangle. These two rock types form a narrow subunit which can be traced, largely by exposures in prospect pits, for almost 1 mile in a N. 5° W. direction. Similar rocks are not known to occur elsewhere in the Mayo Formation.

In its general lithology, the Mayo Formation does not differ greatly from the Bugtown Formation. However, the Mayo is richer in such minerals as garnet and staurolite. Also, biotite is commonly more abundant than muscovite, whereas the reverse is true in the Bugtown Formation. Generally the Mayo is more diversified and does not have as much of the uniform thick-bedded quartz-mica-feldspar schist that is characteristic of much of the Bugtown Formation. The original sediments of the two formations were probably very similar, but the Mayo probably contained more fine-grained clay. The different rock types in the Mayo Formation are more fully described, and numerous modes are presented, in the report on the Fourmile quadrangle (Redden, 1963).

AMPHIBOLITE

Amphibolite occurs in approximately 60 separate sills and dikes in the west half of the quadrangle. Many of these occur in the Mayo Formation near the major synclinal axis, and several are near the middle of the Vanderlehr Formation. Only four other bodies—three small dikes and a poorly exposed oval mass—occur in the other metamorphic formations. Most of the amphibolite bodies are sills or are parallel to an axial-plane foliation, and only a few appear to be discordant irregular bodies or dikes. The sills and dikes range in thickness from about 10 feet to several hundred feet and may be more than 1 mile long. Many sills in the vicinity of the Wabash mine (pl. 34, B–8) are folded, or at least follow folds in the Mayo Formation.

The amphibolite is dark green to nearly black. Its nearly massive character is generally not interrupted except by weak foliation parallel to that in the country rock. In a few outcrops near the contacts with the country rock, however, the amphibolite is gneissic to almost schistose. Most specimens have a typical "salt and pepper" texture, wherein grains of light-colored plagioclase are in a matrix of dark hornblende. The average mineral composition of the amphibolite intrusives in the Mayo Formation is about 50 percent hornblende, 45 percent plagioclase (andesine or, in places, oligoclase), and minor amounts of magnetite and sphene. Amphibolite is somewhat richer in hornblende in the Bear Mountain area and more commonly contains garnet. The outer parts of many of the amphibolite bodies contain a small percentage of biotite and quartz, and a few contain 10–20 percent garnet. The garnet-rich specimens have less hornblende, and the garnet clearly formed at the expense of hornblende. Most of the plagioclase is strongly reverse zoned and untwinned.

A thin wedge-shaped sill is exposed in the vertical quartzite cliffs along the upper part of Vanderlehr Creek. Where the sill is about 10 feet thick it has the typical mineralogy of the amphibolite. Where it is only a few feet thick, however, it is composed of about 45 percent biotite, 20 percent hornblende, 15 percent oligoclase, 10 percent quartz, and 10 percent calcite.

The amphibolite seems unquestionably to have originally been an intrusive igneous rock. The crosscutting relationships as well as the similarity in chemical composition to diabase and gabbro permit little doubt on this point. Furthermore, partly altered but recognizable inclusions of country rock occur in the amphibolite in the Fourmile quadrangle (Redden, 1963). Runner concluded (1943, p. 456) that the amphibolite bodies in the Bear Mountain area are of sedimentary origin. His conclusion was probably a consequence of mapping that had not been sufficiently detailed to separate the intrusive amphibolites from the amphibole schists and other rocks of the Vanderlehr Formation; thus, rocks of different character and different origin were treated as the same.

LITHOLOGIC UNITS EAST OF GRAND JUNCTION FAULT

Four separate major lithologic units of Precambrian age are shown on plate 34 in the area east of the Grand Junction fault. From east to west along the north edge of the quadrangle these units are (1) quartzite and quartz-mica schist, (2) quartz-biotite-garnet schist, (3) mica schist, and (4) quartzite. The mica schist unit is repeated west of the quartzite unit. The quartzite either is in the nose of a fold or is a lense in the mica schist; which occurrence is correct cannot be determined from the evidence currently available.

The stratigraphic relations of these four units are not fully known. Some of the units may correlate with formations west of the Grand Junction fault. Because of so many uncertainties, however, it seems best to describe the units lithologically and to postpone the formal naming or correlation of the units until more information is available from adjacent areas.

QUARTZITE AND QUARTZ-MICA SCHIST

A distinctive unit consisting largely of impure quartzite and quartz-mica schists extends along the east side of the Berne quadrangle. The rocks generally dip moderately to the west, are considerably deformed, and may be as much as 1,400 feet thick in the widest exposure in the Berne quadrangle. The western part of this unit consists mainly of aluminous schist about 800 feet thick, and the eastern part is impure quartzite or quartzose schist. These rocks extend to the southeast into the Custer quadrangle; the quartzite also extends to the northeast into the Hill City quadrangle, but the aluminous schist becomes very thin or lenses out entirely.

The quartzites are thick bedded and brownish gray. They are interbedded with quartz-mica-feldspar schist and quartz-mica schist. Outcrops cap the ridges in a crescent-shaped area in the extreme east-central part of the quadrangle and are especially abundant on St. Elmo Peak, about 0.5 mile east of the northeast corner of the quadrangle. The quartzites are medium grained and locally contain aggregates of quartz that appear to be relict grains. Individual beds are 4–6 feet thick and consist of 75–90 percent quartz, 5–10 percent each of biotite and feldspar, and accessory muscovite, zircon, tourmaline, garnet, apatite, and magnetite. These beds grade into massive quartz-mica-feldspar schist beds that are similar to those in the Bugtown Formation. Ellipsoids of calc-silicate rock are abundant at some localities.

The aluminous schist to the west, which is structurally higher, consists largely of light-gray to brown quartz-mica schist. It is interlayered with more micaceous beds—some are silvery medium-grained mica schist rich in sillimanite, and some contain small amounts of andalusite and staurolite. All gradations occur from micaceous schist to massive or thick-bedded types similar to the impure quartzites. A few thick beds have grit-sized aggregates of quartz and feldspar in a quartz-rich matrix. Some of the more quartzose beds have calc-silicate ellipsoids indistinguishable from those in the Bugtown Formation. One such ellipsoid, from a railroad cut about 600 feet south of the track crossing Tenderfoot Gulch, contained a 2-inch long crystal of bronzite.

The andalusite-bearing schist forms at least two subunits near the west contact of the unit. The larger of these subunits is possibly as much as 75 feet thick. It contains large flat aggregates of andalusite crystals that form conspicuous knots 3–4 inches across; one aggregate—found near the railroad due west of Thunderhead Mountain—was 18 inches long, 6 inches wide, and about 1.5 inches thick. Many, or most, of the andalusite aggregates are replaced by muscovite, and a few, by sillimanite. The staurolite in some interbeds is commonly replaced by muscovite and sillimanite and forms aggregates of needles that form lumps or knots in the schist (fig. 125*E*).

Near the boundary between the impure quartzite and the schist is a discontinuous 5–10-foot thick bed of glassy-textured streaked quartzite. Locally quartzite exposures also contain gunerite-quartz gneiss typical of the meta-iron-formations commonly associated with similar quartzite in the Bugtown Formation.

QUARTZ-BIOTITE-GARNET SCHIST

A relatively thin, but unusually distinctive, lithologic unit characterized by thin-bedded schists structurally overlies the quartzite and quartz-mica schist unit. This unit has been traced through the eastern part of the Berne quadrangle and northward to the north end of the Hill City quadrangle. To the south the unit has been traced for nearly 15 miles to the town of Pringle (fig. 122).

Quartz-biotite-garnet schist is dominant, but the unit also contains many beds of quartz-biotite-muscovite and quartz-biotite schist. Less abundant rock types in the unit include biotite schist, muscovite-biotite schist, biotite-garnet schist, graphitic schist, quartz-microcline-biotite schist, plagioclase-biotite schist, quartzite, quartz-cummingtonite-gunerite gneiss, and amphibolite.

Excellent exposures are along Spring Creek and in railroad cuts near the Berne siding (pl. 34, G–5). Elsewhere the unit is poorly exposed; the few natural outcrops are dark gray to almost black and are commonly iron stained. In many areas, especially where the beds have low dips, trees are virtually absent on soil derived from this rock unit. This is especially noticeable in the several square miles of "park" around the Park School.

The eastern contact with the quartzite and quartz-mica schist unit is well exposed on the south side of Spring Creek. Apparently it is conformable. Locally the contact seems to be somewhat gradational, and lithologies typical of the quartzite and quartz-mica schist unit occur within the thin-bedded schist.

In the best exposed sections the quartz-biotite-garnet schist unit is an estimated 500 feet thick. Folds are visible in most exposures, and some localities exhibit tight folds deformed by younger folds, therefore, the exact thickness is difficult to determine. In the northeastern part of the quadrangle the unit has an outcrop width of approximately 1 mile and dips steeply, but the apparent large thickness is caused by at least three folds that are outlined by lithologic units just north of the quadrangle boundary (fig. 122).

Thin-bedded medium-grained generally dark schists are characteristic of the entire unit. Beds range in thickness from less than 0.1 inch for some of the biotite-rich schists to as much as 2 feet for the more quartzose schist. In general the darker schists, rich in biotite and garnet, are thinner bedded than the lighter colored, more quartzose, feldspathic beds. Probably all gradations occur from the lighter colored beds (which are largely quartz-biotite-muscovite schist) to the darker ones (which are predominantly quartz-biotite-garnet schist). Representative modes of some of the beds are given in table 3. Some of the dark beds contain more than 40 percent garnet, which is generally porphyroblastic and dark brownish red to almost black and which has an index of refraction of 1.805–1.815.

Most of the thin layers of quartzite and quartz-cummingtonite-grunerite gneiss are less than 1 inch thick and lens out along strike. Most of the fine- to medium-grained quartz in both rocks is sugary textured, although in a few exposures it is coarse grained and resembles vein quartz. Locally the quartzite beds have been so tightly folded that only fold crests remain, in a series of "rods." Figure 127 shows some intricacies of the structure. Most of the cummingtonite-grunerite is in coarse-grained white to pale-green fiberous aggregates. The β index of refraction of the cummingtonite-grunerite ranges from 1.675 to 1.700. Green hornblende is commonly associated with the cummingtonite-grunerite and has approximately the same β index. The associated minerals are noted in the modes of samples 126, 127, and 130 (table 3).

Amphibolite forms a single bed (or possibly two beds) that can be traced discontinuously through most of the length of the quardangle; this bed is shown separately on plate 35 in the Park School area. It is only about 3 feet thick in its thickest exposures (in fold noses) and locally pinches out on the limbs of folds. The amphibolite is light green to green and medium to coarse grained. Some samples resemble gneiss, and others are nearly schistose. The rock consists largely of actinolite, quartz, epidote, and sphene.

Certain minor lithologies in the quartz-biotite-garnet schist unit are present only in the area around Park School and in the area southwest of Thunderhead Mountain. A notable example is the microcline-rich

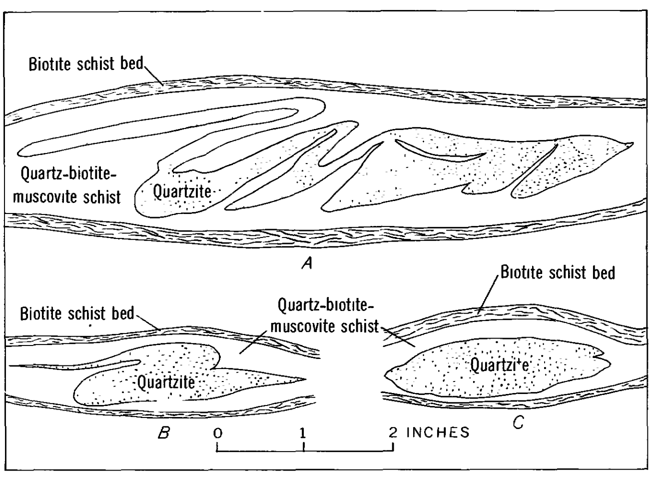


FIGURE 127.—Early folds in quartzite beds. Railroad cut 800 feet west of bench mark 5,752, Park dome area, quartz-biotite-garnet schist unit. Natural size.

TABLE 3.—*Modes of rocks in the quartz-biotite-garnet schist unit*

[Tr., trace]

Sample	124		125		126	127	128		129	130	131	132	133	134	134–1	135	136	136–3	136–27
Quartz	40	58	----	30	20	30	----	45	56	25	46	50	70	15	12	30	29	35	15
Biotite	20	40	80	35	49	30	50	18	35	----	42	40	26	5	30	30	25	35	30
Muscovite	----	----	Tr.	----	----	----	----	----	----	----	----	1	----	10	1	3	2	3	2
Microcline	----	----	----	----	----	----	50	30	5	----	----	----	----	----	----	----	30	20	52
Plagioclase	----	----	----	2	25	----	----	----	----	----	----	5	----	----	----	20	12	5	----
(Percent anorthite in plagioclase)[1]	----	----	----	(40)	(40)?	----	----	----	----	----	----	----	----	----	----	(30–40)	----	(20)	----
Garnet	40	2	20	32	3	10	----	7	2	15	12	4	----	----	----	4	----	2	----
Cordierite	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----	12	----	----	----
Cummingtonite-grunerite	----	----	----	----	Tr.	30	----	----	----	60	----	----	----	----	----	----	----	----	----
Hornblende	----	----	----	----	Tr.	----	----	----	----	----	----	----	----	----	----	----	----	----	----
Actinolite	----	----	----	----	----	----	----	----	----	----	----	----	----	57	45	----	----	----	----
Zoisite	----	----	----	----	----	----	----	----	----	----	----	----	----	----	Tr.	----	----	----	----
Sphene	----	----	----	----	----	----	----	----	----	----	----	----	----	6	8	----	----	----	----
Zircon	Tr.	Tr.	Tr.	Tr.	Tr.	----	----	----	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	----	----	Tr.	----
Apatite	----	----	----	----	----	----	----	----	----	----	----	----	Tr.	6	4	----	----	----	----
Tourmaline	----	----	Tr.	1	2	----	----	----	----	----	----	----	----	----	----	----	----	Tr.	1
Rutile	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----	Tr.	----	----
Graphite	----	----	----	----	----	Tr.	----	----	1	Tr.	Tr.	Tr.	----	----	----	----	----	Tr.	----

[1] Approximate where determined.

DESCRIPTION OF SAMPLES

124, 125. Light and dark beds of quartz-biotite-garnet schist.
126, 127. Quartz-biotite-garnet schist containing lenses of cummingtonite-grunerite.
128. Microcline-biotite schist beds.
129, 131, 132, 133. Quartz-biotite schist beds.
130. Grunerite-quartz-garnet gneiss.
134, 134–1. 10-inch layer of amphibolite.
135. Quartz-biotite-plagioclase-cordierite schist adjacent to amphibolite in sample 134.
136, 136–3, 136–27. Microcline-biotite quartz schists adjacent to amphibolite in sample 134.

schist beds that occur stratigraphically within a few tens of feet of the amphibolite bed. Microcline constitutes nearly 30 percent of the rock, and in a few places as much as 60 percent. The microcline occurs as lath-shaped white almost fibrous porphyroblasts as much as 2 cm long. The porphyroblasts cross the schistosity of the rock in a random fashion and envelope relicts that have schistose fabric (fig. 125 *C*, *D*). The lathlike shape differs from the more equidimensional shape of microcline porphyroblasts near the pegmatites in the Bugtown Formation. The associated minerals are indicated in the modes of samples 136, 136–3, 136–27, in table 3. Garnet crystals are broken and strung out, and some are replaced by micaceous minerals (fig. 125*B*). The microcline-rich schists and associated thin-bedded biotite-rich schist are severely crumpled and deformed. A strong slip cleavage occurs along the axial planes of tiny crenulations, such as those shown in figure 125*B* and *D*. Thin sections show that the micas are bent and broken along the planes of the slip cleavage.

Graphitic schist occurs as dark thin beds in various parts of the unit. These beds and the rocks interbedded with them are generally rich in biotite and garnet. Sulfide minerals make up as much as 15 percent of some of these dark beds. Both the graphite and the sulfides are thought to have been primary rather than introduced during a subsequent mineralization. The high arsenic content of a composite sample of the quartz-biotite-garnet schist unit given in table 4 suggests that much of the sulfide is arsenopyrite. Table 4 also shows that the unit is somewhat rich in many metals, but the metal content is not significantly different from the average for shales as given by Rankama and Sahama (1950, table 5.52). The analysis is high in arsenic and manganese compared to average figures, but not sufficiently high to require later addition of these constituents. The high content of metals, graphite, and sulfide would be expected in a black shale, which was almost certainly the original nature of this rock.

TABLE 4.—*Spectrographic analyses, in percent, of a composite sample of the quartz-biotite-garnet schist unit, Park School area*

[Sample 57-3930. Values cited as less than (<) are limits of sensitivity for those metals. Semiquantitative spectrographic analysis by Uteana Oda, U.S. Geol. Survey]

Ag	<0.0001	Fe	>5	Sb	<0.01
As	.07	Ge	<.002	Sn	<.001
B	.01	In	<.005	Sr	.007
Ba	.05	La	<.01	Ta	<.01
Be	<.0001	Mg	1.5	Ti	.3
Bi	<.001	Mn	.5	Tl	<.01
Cd	<.005	Mo	.001	V	.03
Co	.002	Nb	<.002	Y	.005
Cr	.007	Ni	.005	Zn	<.03
Cu	.01	Pb	.005	Zr	.02

MICA SCHIST

Approximately 2,000 feet of aluminous schists, in which the predominant rock is massive mica schist, constitutes a lithologic unit in the northeastern part of the Berne quadrangle. This unit (pl. 34) lies east of the Grand Junction fault, and similar rocks have been traced northward across the Medicine Mountain quadrangle and into the northwestern part of the Hill City quadrangle (fig. 122). The eastern part of this unit, which is structurally its lower part, has layers of dark thin-bedded schists that are lithologically the same as the quartz-biotite-garnet schist unit. Thus the contact between the massive mica schist and the quartz-biotite-garnet schist is a gradational one. The western contact of the mica schist is the Grand Junction fault. However, a

275-306 O - 68 - 4

unit of quartzites, shown separately on plate 34, occurs near the middle of the mica schist along the north border of the quadrangle. The quartzite may be in the nose of a fold, but currently available evidence does not exclude either the possibility that it is cut out by a fault or the possibility that its southern extremity is the site of a facies change.

The mica schist unit forms large jagged outcrops and ridges of nearly continuous exposures, especially in the area between Tenderfoot Gulch and Spring Creek. It consists largely of massive mica schist that is homogeneous for thicknesses of 100–250 feet, but at many localities garnet-rich schist is interbedded with the mica schist. Other rock types are quartz-biotite-garnet schist and quartz-mica schist. The quartz-biotite-garnet schist in small outcrops cannot be distinguished easily from the much thicker quartz-biotite-garnet schist unit to the east. Quartz-mica schist occurs mainly in a body about 200 feet thick that can be traced from the vicinity of the Old Bill mine (pl. 34, F–4) north-northwestward approximately parallel to the Grand Junction fault for almost 2 miles, and there apparently lenses out.

The massive mica schist forms large silvery-gray outcrops. Indistinct laminae, a few millimeters to 1 cm thick, are caused mostly by variations in biotite content. Many schist exposures near Tenderfoot Gulch have small-scale rodding or streaking caused by crenulations along a minor foliation that crosscuts the major foliation of the rock. Chlorite and biotite, in small porphyroblasts or in coarser grained aggregates several millimeters across, are most commonly oriented at a high angle to the dominant foliation. The mica schist is very fine grained except for the micaceous aggregates and a few porphyroblasts. Quartz and plagioclase grains are 0.1 mm across or smaller, and garnets are generally about 0.15 mm across and only seldom as large as 1–2 mm.

The mineral composition of the mica schist is shown in table 5 (samples 7–32, 6–22, 5–5, 5–23C, 6A, 6–21, 6–11, 6–10, 6–6, 5–23A, 5–23B). Most of the modes contain a total of about 90 percent of muscovite, biotite, and quartz in the following ranges: muscovite, 30–50 percent; biotite 10–30 percent; and quartz, 10–30 percent. Plagioclase is a major constituent in many places, and garnet and magnetite or ilmenomagnetite are common. Some of the rocks contain porphyroblasts of cordierite, staurolite, andalusite, and apatite. Tourmaline and zircon are common accessory constituents.

Staurolite, the most widespread of the aluminum silicate porphyroblasts, occurs in some of the beds throughout the mica schist unit. Cordierite and andalusite are most abundant in the area north of Tenderfoot Gulch but are present as far south as the Old Bill mine.

Staurolite forms brown euhedral crystals 1 centimeter or more long, but the crystals are generally smallest in the northern part of the area. Crystals have many poikilitic inclusions of most of the groundmass minerals except biotite, which (for the most part) has been replaced by staurolite. Some staurolite porphyroblasts are pseudomorphically altered to muscovite aggregates.

Cordierite is present only in massive mica schist; it forms large spindle-shaped black crystals as much as 4 cm long and 1 cm thick that commonly have pointed ends and are oriented parallel to the dominant lineation of the schist. Some crystals are bent at their ends. The cordierite is filled with inclusions of ilmenite or magnetite and also has inclusions of zircon with intense yellow pleochroic halos.

Andalusite forms aggregates of large gray skeletal crystals, which make conspicuous knots or lumps on the weathered schist outcrops; some of the knots are as much as several inches across. Mode 20 (table 5) indicates the kinds and amounts of minerals contained in sample 5–24. Most of these aggregates consist of only about 50 percent andalusite. The other minerals are the same as those in the groundmass of the schist and, except for muscovite, occur in approximately the same proportions as in schist. The muscovite content, however, is so low in and near andalusite (table 5, modes 12, 13) that the andalusite must have formed from muscovite. This reaction works both ways, for andalusite is replaced by fine-grained muscovite (fig. 128).

Magnetite and ilmenomagnetite are common accessory minerals in the mica schist. In some specimens the magnetite forms more than 5 percent of the rock, mostly as well-formed octahedrons that average nearly 0.8 mm in diameter. The ilmenomagnetite occurs as platy grains that are about one-fifthieth the size of the magnetite grains. Polished sections show that these platy grains consist of about 10–20 percent ilmenite in parallel laths.

The interbedded garnet-rich rock and mica schist consists of thin light-brown to tan garnet-rich beds, ranging from 0.5 to 3 inches in thickness, interbedded with mica schist beds that are similar to the silvery massive mica schist. The garnet beds make up 10 percent of some exposures, and mica schist the remainder. The garnet beds consist of garnet, quartz, plagioclase, and biotite (table 5, mode 19); some contain as much as 70 percent garnet, giving the rock a massive and granular appearance. The index of refraction of the garnet ranges from about 1.793 to 1.799. One specimen with an index of refraction of 1.798 ± 0.003, X-rayed by Edward J. Young, U.S. Geological Survey, was found to have a unit cell of 11.58 A; these data indicate that the sample has a composition of about 65 percent spessartite, 20–25

TABLE 5.—*Modes of rocks of the mica schist unit*

[Tr., trace]

	Berne quadrangle																				Medicine Mountain quadrangle				Hill City quadrangle, northern part					
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30
Sample	7-32	6-22	5-5	5-23C	6A	6-21	6-11	6-10	6-6	5-23A	5-23B	6-8	5-23E	5-22A	5-17	5-20	6-1	6-2	5-19	5-24	RG-1	RG-2	RG-3	RG-4	5B	5C	5A2	5b1	5a1	5A
Quartz	42	34	33	31	32	30	25	22	15	14	7	43	5	---	38	50	43	44	30	8	40	15	31	27-33	40	39	36	31	30	24
Muscovite	22	45	45	34	35	25	46	47	60	39	45	6	27	60	12	8	---	17	---	4	20	35	45	38	42	33	35	36	27	48
Biotite	33	11	15	27	12	15	18	13	5	16	14	17	24	15	26	25	7	17	---	10	7	5	15	18	14	15	24	19	38	25
Plagioclase	Tr.	3	Tr.	---	5	4	2	11	15	25	26	4	10	---	---	Tr.	35	4	20	7	4	3	2	4-10	---	13	---	---	---	---
(Percent anorthite in plagioclase)[1]	---	---	---	---	(15-20)	---	(22)	(20)	(20)	(16)	(20)	(20)	(20)	---	---	---	(25)	(20)	(30)	---	(16)	(16)	(15)	(18)	---	(15-20)	---	---	---	---
Staurolite	Tr.	---	---	---	---	4	---	---	---	---	---	---	---	---	13	6	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Andalusite	---	---	---	---	---	5	---	---	---	---	---	25	27	---	---	---	---	---	---	61	---	---	---	---	---	---	---	---	---	---
Garnet	Tr.	3	2	5	4	8	2	2	1	1	Tr.	2	1	4	9	10	12	10	47	---	4	3	2	2	1	Tr.	4	12	2	---
Cordierite	---	---	---	---	---	---	---	---	---	---	---	0	---	10	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Magnetite	---	2	5	Tr.	1	4	6	---	---	5	---	2	5	5	---	---	---	---	---	10	---	---	2-3	3	2	Tr.	---	---	---	---
Ilmenomagnetite	Tr.	1	Tr.	2	Tr.	Tr.	Tr.	4	2	Tr.	5	Tr.	Tr.	4	---	---	Tr.	Tr.	---	---	---	1	Tr.	Tr.	---	---	---	1	1	Tr.
Chlorite	---	---	Tr.	1	10	5	1	---	2	Tr.	3	1	---	---	---	Tr.	1	7	3	---	25	37	2	2	---	Tr.	---	1	Tr.	Tr.
Apatite	Tr.	Tr.	Tr.	Tr.	---	Tr.	---	---	Tr.	Tr.	Tr.	---	Tr.	Tr.	---	Tr.	Tr.	Tr.	---	---	---	---	---	---	Tr.	---	Tr.	---	---	Tr.
Zircon	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	---	---	Tr.	Tr.	---	Tr.	---	Tr.	---	Tr.	Tr.	Tr.	---	---	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.
Tourmaline	Tr.	Tr.	Tr.	Tr.	---	Tr.	---	---	Tr.	Tr.	Tr.	Tr.	1	Tr.	---	---	---	---	---	---	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.
Opaques	---	---	---	---	---	---	---	---	---	---	---	---	---	---	2	1	---	Tr.	---	---	---	---	---	---	---	---	---	---	---	---
Sphene	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	Tr.	---	---	---	---	---	---	---	---	---	---	---	---

[1] Approximate, where determined.

DESCRIPTION OF SAMPLES

1-11. Mica schists from north of Tenderfoot Creek.
12, 13. Andalusite schists.
14. Muscovite-biotite-cordierite schist.
15, 16. Mica-quartz-staurolite schists.
17. Quartz-oligoclase-garnet schist.
18. Quartz-mica-garnet schist.
19. Garnet-rich bed in mica schist.
20. Large andalusite crystal.
21-24. Mica schists from the lower part of the staurolite zone in Reno Gulch.
25-30. Mica schists from the garnet zone.

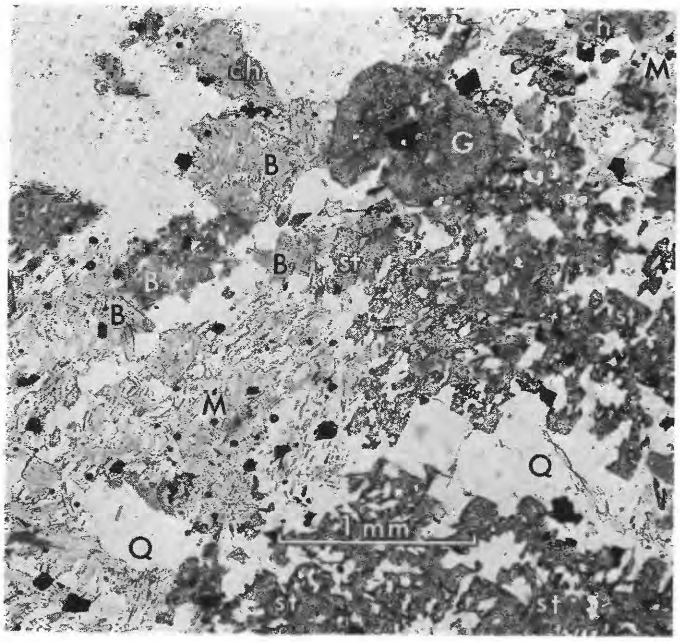


FIGURE 128.—Staurolite-andalusite schist from the mica schist unit. Little of the original mica of the schist has been preserved in this rock. Part of it has been replaced by poikiloblastic staurolite (st). Poikilitic biotite (B) is oriented transverse to the foliation. Large aggregate of muscovite (M) at left of staurolite is probably after andalusite, which is present as scattered remnants elsewhere in the section. Garnet (G), quartz (Q), and magnetite (opaque) are the remaining minerals. Plane-polarized light.

percent almandite, and 10–15 percent pyrope (Fleischer, 1937).

Most exposures within 100 feet of the quartzite unit north of Spring Creek (pl. 34) consist of dark thin-bedded biotite-garnet schist beds interlayered with quartz-mica schist. The darker beds, averaging 0.9 inch thick, consist mainly of biotite and garnet but contain some graphite. The higher biotite content of these beds distinguishes them from the garnet beds that elsewhere are interbedded with mica schist. The lighter-colored quartz-mica schist beds are several inches to several feet thick.

Except for the occurrence of some garnet-rich beds of quartz-rich schist beds within the unit, the mica schist unit is a quite uniform rock. Table 5 contains 30 modes of various types of rock in the unit. The average of these modes (neglecting trace constituents) and the calculated chemical composition (both in percent) are as follows:

Mode		*Calculated chemical composition*	
Muscovite	31	SiO_2	60. 0
Quartz	30	Al_2O_3	21. 6
Biotite	17	TiO_2	. 2
Oligoclase	7	Fe_2O_3	
Garnet	5	FeO }	8. 2
Staurolite	1. 5	MgO	
Andalusit7	3	CaO	. 3
Cordierite	1	Na_2O }	7. 4
Chlorite	1. 5	K_2O	
Magnetite	1. 5	H_2O	2. 3
Ilmenomagnetite	1. 5		

Though the errors in this method of determining the composition of the rock may be substantial, the "average" rock in the mica schist unit almost certainly contains at least 20 percent Al_2O_3 and 7 percent alkalies. The rock composition is therefore that of an aluminum-rich shale.

The spectrographic analyses in table 6 confirm the high content of alkalies. Sample S–1 apparently has a high plagioclase content or possibly contains paragonite (although paragonite was not identified by X-ray analysis of similar rocks). The minor-element contents given in table 6 were compared with those in six spectrographically analyzed samples of quartz-mica schists collected by J. J. Norton (written commun., 1958) from metamorphosed impure sandstones and graywackes of the Keystone area. The mica schists of the Berne quadrangle are richer in Na, K, Mn, Fe, Mg, Co, Cr, Cu, Ni, and Sc, and poorer in Zr. All the elements more abundant in the Tenderfoot schists are also more abundant in shales than in the coarser grained clastic sediments (Rankama and Sahama, 1950).

TABLE 6.—*Semiquantitative spectrographic analyses of mica schist from the mica schist unit*

[Results in percent. Elements looked for but not detected: P, As, Au, Bi, Cd, Dy, Er, Eu, Gd, Ge, Hf, Hg, Ho, In, Lu, Pr, Pt, Re, Rn, Ru, Sb, Sn, Sm, Ta, Tb, Te, Th, Tl, Tm, U, W. M, major constituent; Tr., trace. Analyst: J. C. Hamilton, U.S. Geol. Survey]

Field No.	S–1	S–2	S–6	S–10	S–11	S–14
Lab. No.	245645	245646	245647	245648	245649	245650
Si	M	M	M	M	M	M
Al	M	M	M	M	M	M
Fe	7	7	7	7	7	7
Mg	3	3	3	3	3	3
Ca	.7	.3	.3	.7	.7	.7
Na	7	3	1. 5	3	3	3
K	7	7	7	7	7	7
Ti	.3	.3	.3	.3	.3	.3
Mn	.7	.15	.3	.3	.3	.3
Ag	0	.00015	0	0	0	0
B	.07	0	.007	.007	.007	.03
Ba	.15	.15	.15	.15	.15	.15
Be	.0007	.00015	.00015	.0003	0	0
Ce	.015	.015	Tr.	Tr.	.015	.015
Co	.007	.003	0. 003	0. 003	.003	.0015
Cr	.03	.015	.015	.015	.015	.015
Cu	.0015	.015	.007	.007	.015	.0015
Ga	.007	.003	.003	.003	.003	.003
La	.007	.007	.007	Tr.	.007	.007
Li	Tr.	0	0	0	0	0
Mo	0	.0015	0	0	0	0
Nb	Tr.	.0015	Tr.	.0015	.0015	.0015
Nd	Tr.	Tr.	Tr.	Tr.	.007	Tr.
Ni	0. 007	0. 007	0. 007	0. 007	.007	0. 003
Pb	.007	.003	.003	.003	.003	.007
Sc	.003	.003	.003	.003	.003	.003
Sr	.015	.015	.007	.015	.015	.01
V	.015	.015	.015	.015	.03	.03
Y	.003	.0015	.003	.003	.007	.007
Yb	.0003	.0003	.0003	.0003	.0007	.0007
Zn	.03	Tr.	0	Tr.	0	0
Zr	.015	0. 015	.015	0. 015	.015	.015

QUARTZITE

The unit consisting largely of impure quartzite at the north end of the quadrangle, sec. 17, T. 2 S., R. 4 E., forms a conspicuous ridge that extends north-northwest through most of the Medicine Mountain quadrangle. The unit is about 400 feet thick at the north end of the Berne quadrangle but is nearly 700 feet thick about 2 miles farther north. If the unit is in the nose of a fold, then the true thicknesses would be about half those given.

The quartzite forms gray to dark-gray beds as much as 15 feet thick. Most of it is medium grained and largely recrystallized, but some of the larger quartz grains appear to be relict clastic grains. The average quartz content is about 80 percent; nearly equal amounts of mica and feldspar (microcline and oligoclase) make up most of the other 20 percent.

Interbedded with the thick quartzite beds, but rarely exposed, are a few thin beds of gray quartz-mica schist and thin-bedded biotite-garnet schist. The biotite-garnet schist is dark and consists largely of 1- to 2-inch-thick beds rich in garnet and biotite, interlayered with quartz-biotite schist.

SEQUENCE AND POSSIBLE CORRELATION OF LITHOLOGIC UNITS

The stratigraphy of the lithologic units east of the Grand Junction fault is not completely known, nor has the relation of these units to the rocks west of the fault been established. That the two groups of rock correlate with each other is probable, but available data are inadequate to support firm conclusions.

The sequence west of the fault, from oldest to youngest, is Vanderlehr, Loues, Bugtown, Crow and Mayo Formations. Reconnaissance indicted that metamorphic rocks near Harney Peak (fig. 122) are lithologically very similar to the Vanderlehr Formation. Furthermore, these rocks are overlain by rocks that could well be the Loues Formation at a higher metamorphic grade. The quartzite and quartz-mica schist unit in the eastern part of the Berne quadrangle is lithologically similar to the Bugtown; east of the Berne quadrangle the schist has units of meta-iron-formation indistinguishable from those in the Bugtown.

The quartz-biotite-garnet schist and mica schist units have no counterparts west of the Grand Junction fault, unless they are equivalent to the much thinner Loues Formation. This correlation is highly probable for the upper part of the Loues Formation is a mica schist much like the mica schist unit, and the lower part is a thin-bedded schist similar to the quartz-biotite-garnet schist unit. If this correlation should be correct, the sequence on each side of the fault would be as follows:

Sequence west of Grand Junction fault	*Sequence east of Grand Junction fault*
Mayo, Crow	Not present.
Bugtown	Quartzite and quartz-mica schist unit.
Loues, upper part	Quartz-biotite-garnet schist unit.
Loues, lower part	Mica schist unit.

Section *A–A′*, plate 34, shows a structural arrangement east of the fault that is consistent with this stratigraphic interpretation. The key feature is an anticline that is truncated by the fault on the west but has successively younger, but overturned, units to the east. The defect in this interpretation is that the quartzite in the core of this fold would then correspond to some part of either the Loues or the Vanderlehr Formation, and it is quite unlike either of them.

Nevertheless, the general similarity of the rocks on the two sides of the fault is inescapable. The difficulties in arriving at precise correlations may be caused by facies changes and by undetected faults and folds that will become known as mapping continues in adjacent areas.

QUARTZ VEINS

Thousands of quartz veins, cutting all types of metamorphic rocks, occur in the Berne quadrangle, though most are too small to be shown on the geologic map. The quartz veins in the quadrangle range in thickness from a fraction of an inch to more than 100 feet. The large vein at the Grand Junction mine (pl. 34 E–3) reaches this maximum thickness.

Accessory minerals are abundant in the veins, and in many places the same minerals occur in the wallrocks, either as local alteration effects or as widely distributed minerals. In table 7 the distinguishing features of the veins are tabulated, and the veins are divided into 14 types, based on their mineralogic character. Generally the minerals other than quartz do not exceed 5 percent of a vein. The most notable exception is in the veins containing brown tourmaline; exposures across the full width of such veins and several feet along strike may contain about 10–50 percent tourmaline. All minerals listed in table 7 occur in the veins; some of the minerals, especially graphite and tourmaline, are also present in the adjacent wallrock.

The pure, or nearly pure, quartz veins are both concordant and discordant; they are generally small but range from microscopic size to very large veins, such as the one northwest of the North Pole Spring (pl. 34, B–10). The quartz-gold veins are nearly all discordant and most trend N. 20°–60° E. The contacts of these veins are very sharp, and the adjacent wallrock is generally altered by the addition of much tourmaline and graphite. Slickensides are common on the walls of such veins, and fragments of wallrock are bent and broken. Most of the quartz-gold veins are in the Bugtown Formation, but some are also present in all other formations, except possibly the Vanderlehr.

Other varieties of veins are generally smaller than the quartz and quartz-gold types. They are generally concordant or nearly concordant, though in detail their

TABLE 7.—*Description of quartz-rich veins*

Vein type	Mineral assemblages (excluding quartz)	Size range		Wallrock (formation or type)
		Thickness (feet)	Length (feet)	
Quartz	±[1] Graphite and ± tourmaline	<1 ->100	<1 -500	Locally found in all rock types. Predominantly in quartz-rich wallrocks.
Quartz-gold	Graphite, black tourmaline, muscovite, chlorite, arsenopyrite(?) and gold. (These minerals also found in altered wallrock of vein.)	<1 - 4.5	10 -200	Predominantly in Bugtown Formation, but a few occur in all other units except the Vanderlehr Formation.
Quartz-tourmaline (brown)	Brown tourmaline, hematite, magnetite, and sulfide(?).	<1 - 4	2 -150	Upper and lower amphibole schists of the Vanderlehr Formation.
Quartz-tourmaline (black)	Black tourmaline, graphite, plagioclase, chlorite, and muscovite.	<1 - 2	5 - 50	Mayo, Crow, and Bugtown Formations, and the mica schist, quartz-biotite-garnet schist, and quartz-mica schist units.
Quartz-feldspar	Oligoclase ± microcline, chlorite, biotite, garnet, muscovite, tourmaline, and apatite.	<1 - 3	5 -150	Quartz-mica-feldspar schists of the Mayo and Bugtown Formations and quartz-biotite-garnet schist unit.
Quartz-garnet	Garnet, oligoclase, chlorite, biotite, muscovite, tourmaline, and apatite.	<1 - 1	5 - 20	Loues Formation, mica schist unit, and quartz-biotite-garnet schist unit. Also in the garnet schist of the Vanderlehr Formation.
Quartz-staurolite	Staurolite, garnet, ± biotite, ± muscovite, chlorite, and ± andalusite.	<1 - 1	5 - 20	Mayo Formation, mica schist unit, and quartz-biotite-garnet schist unit.
Quartz-andalusite	(1) Andalusite, oligoclase, chlorite, biotite, muscovite, garnet, ± sillimanite, ± staurolite, ± cordierite, ± zircon, ± tourmaline, ± ilmenite, ± magnetite, ± bornite, ± chalcocite, ± chalcopyrite, and ± malachite.	<1 - 4	5 - 50	(1) Micaceous schists in the Loues Formation and mica schists unit.
	(2) Andalusite, oligoclase, chlorite, biotite, apatite, garnet, staurolite, and ± sillimanite.			(2) Micaceous schists from the quartzite and quartz-mica schist unit.
Quartz-sillimanite	Sillimanite, oligoclase, biotite, chlorite, andalusite, muscovite, garnet, and ± staurolite.	<1 - 3	5 - 50	Sillimanite-bearing schists of the middle part of the Bugtown Formation and micaceous schists from the quartzite and qartz-mica schist unit.
Quartz-ilmenite	Ilmenite. (1) chlorite, calcite and rutile. (2) Apatite, andalusite, biotite, chlorite, and ± oligoclase.	(1) <0.1- 1 (2) <0.1- 1	(1) 2 - 10 (2) 2 - 10	(1) Amphibole schists of Crow Formation. (2) Mica schist unit.
Quartz-magnetite	Magnetite, chlorite, and biotite	<1 - 0.5	1 - 5	Mica schist unit.
Quartz-tungsten	(1) Wolframite, graphite, tourmaline, kyanite, andalusite, sillimanite, and ± biotite.	(1) <2.5- 3.0	(1) 40	(1) Mica schist unit.
	(2) Wolframite, graphite, and tourmaline	(2) <1-3 20-60	(2) 50	(2) Mica schist unit.
Quartz-chlorite	Chlorite, garnet, biotite, and oligoclase	<1 - 2	2 - 15	Upper part of the Loues Formation.
Quartz-talc	Talc, calcite, chlorite, ± chalcopyrite, and malachite.	------	0.1- 0.5	Calcite-tremolite gneiss unit of the Vanderlehr Formation.

[1] ±, Mineral may or may not be present.

shapes tend to be irregular and podlike. Some varieties occur only in certain types of wallrock or contain the same accessory minerals as the wallrock.

The quartz–brown tourmaline veins occur only in the amphibole schists of the Vanderlehr Formation. Excellent exposures of these veins occur northwest of Roetzel Deer Camp (pl. 34, C–3). Many of the brown tourmaline crystals have been broken and are cemented by quartz. Magnetite, altered in part to hematite, occurs as small particles in and around the crystals. No tourmaline was noted in the adjacent wallrock.

Quartz–black tourmaline veins were observed in most of the other rock types in the quadrangle. Some veins with sharp crosscutting contacts appear to be transitional to the quartz-gold veins, and others may be transitional to other types of veins.

The quartz-feldspar veins occur largely in plagioclase-bearing quartz-mica-feldspar schist wallrocks in the Bugtown and Mayo Formations. The veins and their characteristic halos of dark minerals in the Fourmile quadrangle were described in more detail by Redden (1963). Although these veins generally have minerals that occur in most of the pegmatites (except for the mineral chlorite), there is no other evidence that the veins are transitional to the pegmatites.

The aluminosilicate minerals garnet, staurolite, andalusite, sillimanite, and cordierite are locally abundant in the quartz veins in the aluminous and iron-rich schists of the Loues Formation and in the mica schist, quartz-biotite-garnet schist, and quartzite and mica schist units. The quartz-garnet and quartz-staurolite veins are generally small and only a few inches thick. The garnet-bearing veins are mainly nodular and irregularly shaped; they contain garnet crystals as much as 2 cm across and abundant oligoclase and chlorite. The staurolite-bearing veins are more tabular than the garnet-bearing veins and generally contain few additional accessory minerals. In some veins the staurolite occurs largely outside the quartz-rich center of the vein and forms a nearly solid plaster of crystals along the vein wall. The wallrock may or may not contain staurolite. Veins rich in staurolite are scarce, but many of the andalusite-and sillimanite-rich veins contain a few crystals of staurolite and garnet.

The quartz-andalusite and quartz-sillimanite veins are generally larger and more abundant than the other aluminosilicate veins, although they are clearly transitional to the other types. The andalusite-rich veins are most abundant in the mica schist unit and in the andalusite-bearing micaceous schist in the lower part of the quartzite and quartz-mica schist unit. A few andalusite-bearing veins were noted in the mica schist of the upper part of the Loues Formation. Many quartz-andalusite-bearing boulders from the quartzite and quartz-mica schist unit occur along the lower part of Tenderfoot Gulch near U.S. Highway 16. The andalusite forms crystals as much as 2 inches in cross section and 10 inches long. Some crystals have graphite inclusions and others are rimmed by muscovite. Many other minerals may be present. For example, one vein about 2,300 feet N. 21° E. from the Old Bill mine has the assemblage quartz, andalusite, oligoclase, sillimanite, staurolite, garnet, cordierite, biotite, muscovite, chlorite, tourmaline, zircon, ilmenite(?), and an undetermined sulfide. The cordierite occurs as small yellow anhedral crystals or as large bluish-gray hexagonal crystals that have an excellent basal parting. According to spectrographic analyses made by J. C. Hamilton, U.S. Geological Survey, both varieties contain 0.15 percent Be. This confirms Follinsbee's (1941) conclusion that cordierite can accept considerable beryllium in its beryl-like structure. Almost black to deep blue cordierite occurs in other andalusite-rich veins in the cordierite-bearing schists of the mica schist unit. Some of the andalusite-rich veins in the more massive micaceous parts of the mica schist unit also contain chalcopyrite, bornite, chalcocite, malachite, and magnetite or ilmenite. The sulfide and oxide minerals generally occur only as a few scattered crystals. Oxidation of the copper sulfides results in some malachite-stained outcrops of schist.

Most of the quartz-sillimanite veins are in the more micaceous schists in the quartzite and quartz-mica schist unit and in the aluminous middle part of the Bugtown Formation in the area above the sillimanite isograd. Sillimanite also occurs in some veins in the mica schist unit in areas below the sillimanite isograd. In many of these veins, andalusite is nearly as abundant as the sillimanite, and the two minerals may be intergrown. Commonly the sillimanite is white to gray silky to dull waxy aggregates. It also forms long laths, which are as much as 12 inches long in veins in the quartzite and quartz-mica schist unit south of the Gold Fish mine (pl. 34, G–5). In some veins sillimanite needles appear to have replaced the more euhedral crystals of andalusite, but in other veins the textural relationships of the two minerals suggest that they formed concurrently.

The quartz-ilmenite and quartz-magnetite veins, generally only a few inches thick, occur mainly in the mica schist unit in wallrocks containing ilmenite and magnetite. A few veins containing these minerals and also calcite and rutile are in ilmenite-rich amphibole schist in the lower part of the Crow Formation. The ilmenite in the veins within the mica schist unit occurs as flat plates that are several inches across. Excellent hand specimens can be obtained from cliffs along Spring Creek approximately 400 feet upstream from bench mark 5,498 (pl. 34, E–2). Deep-green apatite crystals as much as 1 inch across are associated with the ilmenite in these veins. Quartz-magnetite veins are scarce; they have small magnetite crystals in quartz and few other minerals. Magnetite is also a minor constituent in many of the other types of veins in the mica schist unit.

Wolframite was found in two veins but probably occurs in several of the muscovite-rich veins. One of the wolframite-bearing veins is exposed in a prospect pit approximately 5,600 feet N. 9° W. of the Old Bill shaft. This discordant vein, 2–3 feet thick, is in mica schist wallrock that is impregnated with tourmaline and minor graphite adjacent to the vein. Scattered through the quartz in the vein are crystals of kyanite, andalusite, sillimanite, and muscovite. The kyanite and andalusite form crystals several inches long, and the kyanite is locally intergrown with the wolframite blades. Muscovite is largely in the outer part of the vein and is also associated with wolframite. The sillimanite occurs as small felted masses along fractures in the vein and as tiny needles associated with andalusite and kyanite.

Quartz-talc veins are small and are limited to the calcite-tremolite gneiss unit of the Vanderlehr Formation. The adjacent wallrock generally contains some talc. Traces of copper minerals are also present in most of these small veins.

AGE AND ORIGIN

The quartz veins are all believed to be Precambrian, for no veins were noted in the Paleozoic rocks. Galena near the edge of the quartz vein at the Grand Junction mine (pl. 34) is of Precambrian age, according to J. L. Kulp (written commun., 1956). That the quartz and quartz-gold veins formed before the pegmatite is well shown by pegmatites cutting quartz-gold veins at the northernmost shaft of the Saginaw mine; similar relationships were found in the Fourmile quadrangle. The large quartz vein at the Grand Junction mine and several smaller ones to the north along the Grand Junction fault presumably formed after the fault and were localized along fractures related to the fault.

Some of the quartz-rich veins have clearly formed by processes related to the metamorphism; yet others have characteristics suggesting hydrothermal processes associated with igneous activity. The veins rich in the aluminosilicates generally do not show any recognizable evidence of deformation or crystallization under stress. The feldspar in these veins has nearly the same composition as that in adjacent wallrocks, and the veins are inferred to have formed in the main period of metamorphism or in its declining phase. The veins are somewhat similar to those in Vermont described by Chapman (1950), who concluded that the veins having a mantle or envelope of dark minerals similar to those in the country rock formed by metamorphic processes. Certainly in the Berne area the veins that have the same aluminosilicate as their wallrock must have crystallized under physical conditions very similar to those that existed during the metamorphism. On the other hand, veins containing wolframite or gold and having a highly altered wallrock rich in tourmaline and graphite resemble typical deep-seated hydrothermal veins.

The beryllium-rich cordierite suggests a genetic association with the beryl-rich pegmatites and with the nearby pegmatitic granite around Harney Peak. This association seems to be confirmed by the absence of beryllium in spectrographic analyses of cordierite porphyroblasts from the mica schist unit.

Thus, some veins are probably directly associated with the magmatic activity that culminated in the emplacement of the pegmatitic rocks. Possibly there is a continuous sequence ranging from veins formed during metamorphism by recrystallization (probably assisted by fluids generated during the metamorphism) to veins formed largely by the emanation of volatiles and other chemical constituents from the granite and pegmatite. The veins containing tungsten or beryllium-bearing cordierite must be of the latter type.

PEGMATITES

Approximately 1,150 granitic pegmatites have been mapped in the Berne quadrangle (pl. 34). The Black Hills pegmatites are Precambrian, and are about 1,600 million years old according to several isotopic age determinations (Davis and others, 1955, p. 146–147; Kulp and others, 1956, p. 1557; Ahrens, 1949, p. 255). The pegmatite consists largely of plagioclase (albite-oligoclase), quartz, perthite, and muscovite. Common accessory minerals are tourmaline, apatite, and garnet. A few pegmatites contain the less common minerals beryl, lithiophilite-triphylite, amblygonite, spodumene, biotite, columbite-tantalite, cassiterite, loellingite, graftonite, allanite, uraninite, and secondary uranium minerals.

Much of the pegmatite is relatively fine grained and nearly granitic in texture. However, in all the bodies some larger crystals, generally of perthite, give the rock a pegmatitic texture. The nomenclature of pegmatitic textures used in this report is as follows:

	Inches
Very fine grained	<¼
Fine grained	¼–1
Medium grained	1–4
Coarse grained	4–12
Very coarse grained	>12

DISTRIBUTION

Most of the pegmatite bodies shown on the geologic map (pl. 34) are in the eastern and southeastern part of the quadrangle, where they are peripheral to the large body of pegmatite and granite around Harney Peak (figs. 122, 129). They decrease in abundance to the northwest, and none have been found in a belt several miles wide extending northeast from Round Mountain toward the Junction Ranger Station. Another small group of pegmatites occurs in the northwesternmost part of the quadrangle near Bear Mountain. The outer limits of the two pegmatite-bearing areas are shown in figure 130.

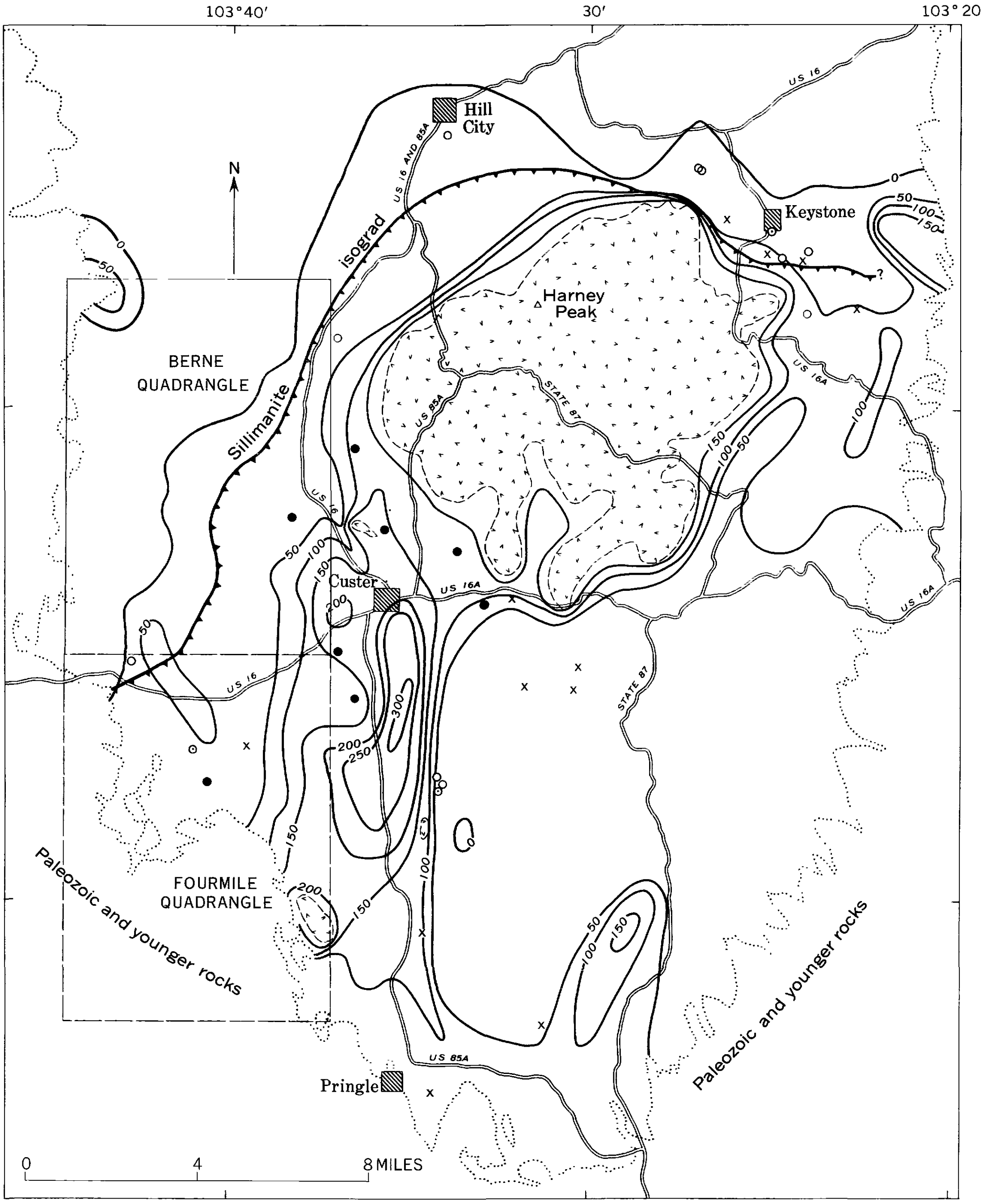
Pegmatites seem to be distributed indiscriminately in all types of wallrock except one. The Crow Formation contains only one pegmatite along a strike distance of about 12 miles through areas rich in pegmatite. This unit underlies a relatively small area, however, and thus the discrepancy may not be too significant.

SIZE, SHAPE, AND ATTITUDE

The pegmatite bodies range in size from thin, sill-like lenses 1–2 feet thick and a few tens of feet long to intrusives more than 100 feet thick and 2,000 feet long. Many of the bodies form dip slopes on ridges or hills, and the true thicknesses may be much less than the outcrop width indicated on the geologic map (pl. 34). For example, pegmatite 64 (D–8)[1] is only about 15 feet thick; yet it crops out over a 140-foot width. Pegmatite 44 (F–7) is probably only 10–15 feet thick but has a very wide outcrop.

About 95 percent of the pegmatite bodies are less than 10 feet thick, as in the Fourmile quadrangle. Only two pegmatites are believed to be more than 100 feet thick (Nos. 15, G–3; 22, G–4), and very few are more than 50 feet thick. The longest pegmatite forms a thin sill located a few hundred feet west of pegmatite 48 (G–8); the sill is less than 10 feet thick but is about 2,600 feet

[1] Pegmatite numbers refer to certain numbered pegmatite bodies shown on pl. 34, and the code D–8, given in parentheses is used to locate the pegmatite on the map. If a pegmatite body has been mined and is locally known by a name, the name is also used.



EXPLANATION

Area underlain by more than 50 percent pegmatite and granite

Contact between Paleozoic and Precambrian rocks

Modified from Darton and Paige (1925)

Isopleth showing number of pegmatite bodies per square mile

Perthite feldspar mine

Sheet-mica mine

Beryl–scrap-mica mine

Lithium mine

FIGURE 129.—Distribution of pegmatites in the southern Black Hills, S. Dak.

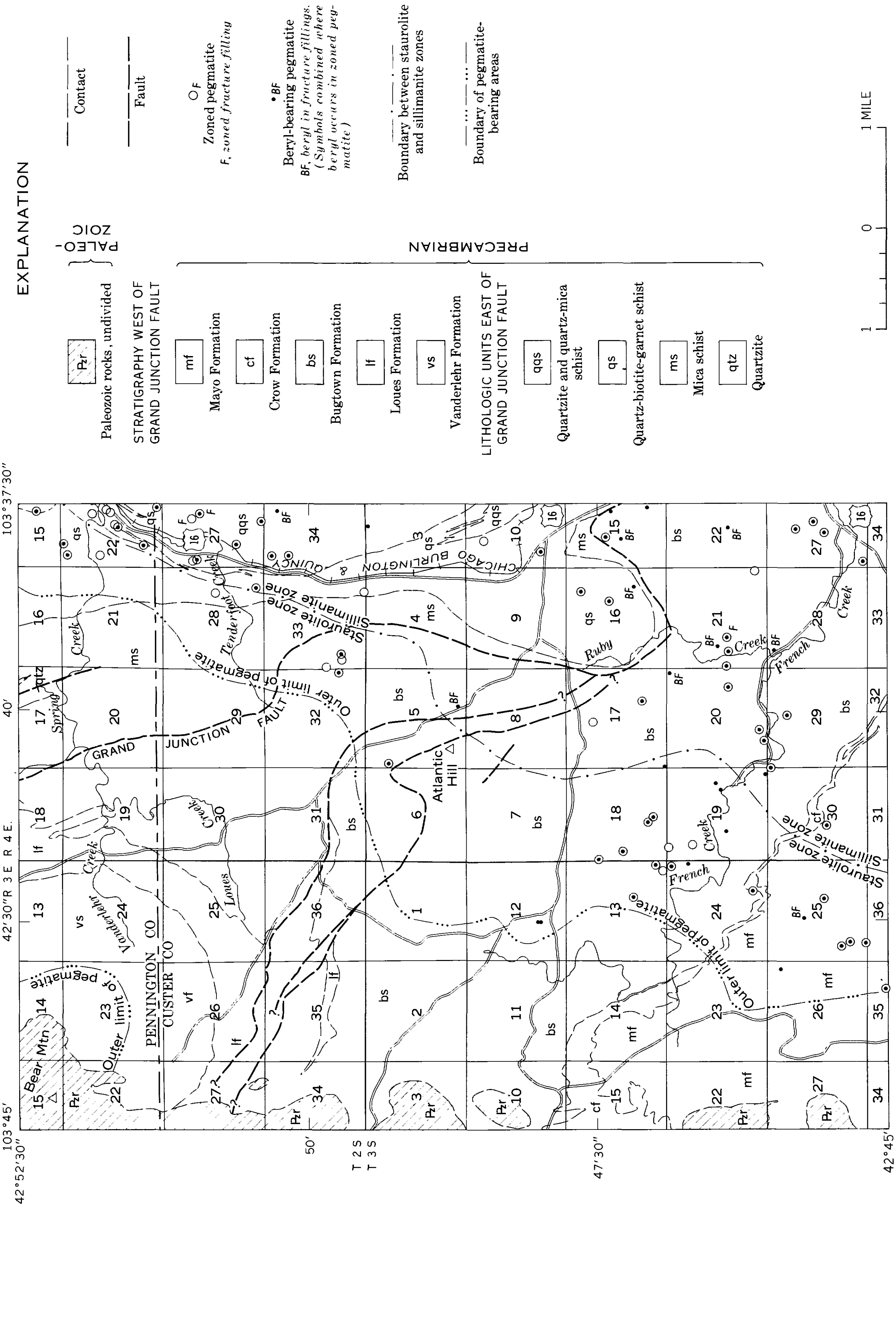


FIGURE 130.—Distribution of zoned pegmatites and beryl-bearing pegmatites in the Berne quadrangle, South Dakota.

long. Several other pegmatite bodies are more than 2,000 feet long, but most are only 200–300 feet long. Many of the larger pegmatites appear to thin slightly with depth, and probably many of the smaller bodies bottom at fairly shallow depths.

The length-thickness ratio generally ranges from 10 : 1 to 300 : 1; most of the pegmatite thus is in tabular bodies. Some pegmatites gradually thin near their ends; others maintain their full thickness and end abruptly in a semicircular nose somewhat like the blunt ends of Tertiary rhyolite dikes described by Noble (1952, p. 36–40) in the northern Black Hills. Some of the thicker bodies split into several parallel sills or dikes, which may rejoin and enclose lenses of country rock, as in pegmatite 48 (G–8). Some of the small pegmatites near pegmatite 48, though shown as single bodies on the geologic map (pl. 34), consist of two or more thin sills separated by lenticular schist septa a few inches to a few feet thick.

A few pegmatites such as the Highland Lode (D–9) and Wilhelm (E–4) are well enough exposed to show that they decrease in size at depth. In cross section both of these pegmatities have a shape that somewhat resembles an inverted tear drop.

Most pegmatities shown on plate 34 are concordant and tend to outline the structure of the country rock, particularly in the southeastern part of the quadrangle. The pegmatites follow both early folds (No. 46, F–8) and late folds (No. 37, F–6). Other pegmatites are parallel to the axial-plane foliation, where it is the strongest planar element of the rocks.

Only about 10 percent of the pegmatites in the Berne quadrangle are discordant, compared with about 30 percent in the Fourmile quadrangle. In both areas there is a large group of dikes that trend northward and dip steeply. These are most abundant in the strongly jointed Mayo Formation and commonly have right-hand jogs or foldlike curves in which the short limb parallels the bedding. In both quadrangles there is another group of dikes that strike northwestward and dip steeply southwest. Another group of northeast-trending dikes are abundant in the Fourmile quadrangle, but in the Berne quadrangle only a few vertically dipping pegmatites trend northeastward.

CLASSIFICATION AND INTERNAL STRUCTURES

The pegmatites shown on plate 34 are classified into four types according to their internal structures: (1) zoned, (2) layered, (3) homogeneous, and (4) gneissic. The layered pegmatites consist of alternating repetitive layers of differing composition and texture that are arranged generally parallel to the outer contacts of the bodies. Zoned pegmatites contain zones of different composition and texture that are in a definite concentric sequence, without repetition, around a central core (Cameron and others, 1949, p. 16–24). The homogeneous pegmatite bodies are characterized by an absence of internal structures and are similar to those described by Johnston (1945, p. 1025). Gneissic pegmatites have been sufficiently deformed that the primary structure, if any, cannot be determined with certainty.

All gradations exist between the different types of pegmatite, and parts of some pegmatites have different dominant internal structures. For example, the outer parts of some zoned pegmatites, such as the Wilhelm, are well layered. Although the correct classification of some pegmatites may be uncertain, that of most of the pegmatites is clear.

ZONED PEGMATITES

About 4 percent of the pegmatites in the Berne quadrangle are zoned. These generally correspond to the heterogeneous pegmatites of Johnston (1945, p. 1025) and to the complex pegmatites of Schaller (1925) and Landes (1933, p. 33–35). Zoned pegmatites intrude nearly all varieties of metamorphic rock in the quadrangle, and they occur in all parts of the pegmatite-bearing area except around Bear Mountain. Not readily apparent from on plate 34, is the fact that the zoned pegmatites are most abundant in areas containing relatively few pegmatites. Also, the zoned pegmatites are near the outer boundary of the pegmatite-bearing area around Harney Peak, as shown in figures 129 and 130. The distribution is similar in the Fourmile quadrangle (Redden, 1963). The percentage of zoned pegmatites relative to all pegmatites in the Berne quadrangle is approximately twice that in the Fourmile quadrangle. The higher percentage of zoned pegmatites in the Berne quadrangle is almost certainly due to the fact that more of the quadrangle is near the outer boundary of pegmatite-bearing areas.

The zoned pegmatites do not differ substantially from the other pegmatites in size, shape, attitude, or type of wallrock. However, none are as large as the largest layered pegmatites, although a few of them, such as the Rachel D (F–7), are moderately large. Most are thin sills or dikes. The Highland and Wilhelm have somewhat oval shapes. Many of the zoned pegmatites, especially in the Mayo Formation, are discordant and follow joints in the host rock.

Most zoned pegmatites in the Berne quadrangle contain only border and wall zones and an inner zone or core. About 1 in 10 have mappable intermediate zones. Sheridan (1955) mapped five zones including three intermediate zones in the High Climb pegmatite (No. 9, G–2). The outer zones of most of the pegmatites are rich in plagioclase, quartz, and muscovite (table 8);

intermediate zones have abundant perthite or, rarely, cleavelandite (table 9); and the cores are rich in quartz or perthite (table 10). The cores of lithium-rich zoned pegmatites contain spodumene, quartz and feldspar.

The bulk composition of a zoned pegmatite is difficult to determine because to do so requires a knowledge of the relative volumes of each zone as well as of the mineral proportions of the zones. The very coarse textures in some zones further complicate sampling procedures in estimating the bulk composition. Zoned pegmatites seem to be richer in quartz and plagioclase than do the layered or homogeneous pegmatites, and some zoned pegmatites are notably richer in lithium. Zoned pegmatites also contain larger amounts of the less common minerals, such as beryl, columbite-tantalite, and cassiterite, and thus must be richer in Be, Cb, Ta, and Sn.

A rough estimate of the bulk composition of the High Climb pegmatite can be obtained by calculating numerical averages from the composition of the zones as reported by Sheridan (1955, table 1). The resulting percentages are as follows: Plagioclase, 33 percent; quartz, 47 percent; perthite, 11 percent; and muscovite, 6 percent. The nearly pure quartz core, however, forms a very small percentage of the total volume of the pegmatite, and if weighted accordingly, an estimate of 40 percent each for quartz and plagioclase is probably more reasonable. This is close to the computed quartz-plagioclase content of the Peerless zoned pegmatites near Keystone, S. Dak. (Sheridan and others, 1957, table 11).

In certain small areas the zoned pegmatites are more similar to each other in composition and appear to be more closely related to one another than to more distant zoned pegmatites. For example, zoned pegmatites 74–77 are all very small, are located near one another, and contain an inner zone that is rich in perthite, quartz, cleavelandite, and muscovite. Pegmatites 25 and 26 are also identical in appearance, and they are distinctly different mineralogically from pegmatites 74–77 (C–7, D–8).

Most of the zoned pegmatites are coarser grained than either the layered or homogeneous pegmatites. Equant crystals of perthite are 6 feet across in the Rachel D pegmatite. The grain size of quartz, plagioclase, and muscovite is less striking, but clearly coarser, in nearly all the zoned pegmatites than in other kinds of pegmatites. However, the spodumeme-bearing pegmatites 2, 5, 7, 25, and 27 (the Milton pegmatite) (G–2, F–4), have outer zones that are very fine to fine grained, and their spodumene-bearing zones have an average grain size of less than 1 inch.

Mineral assemblages of the different zones in zoned pegmatites of the Berne quadrangle are listed in tables 8–10. These assemblages are fundamentally the same as those described by Page and others (1953, p. 66) for other pegmatites in the Black Hills, but the mineral sequences in individual pegmatites are generally less complex than those previously described, and none of the inner zones contain microcline-perthite or lepidolite. The outer zones of pegmatites in the Berne quadrangle are rich in quartz, plagioclase, and muscovite; inner zones are rich in perthite, quartz, or, rarely, spodumene or other minerals. Only a few of these pegmatites contain intermediate zones, and these zones are rich in quartz and perthite. For a thorough and complete discussion of the zones in a single complexly zoned pegmatite in the area, the reader is referred to the description of the High Climb pegmatite by Sheridan (1955).

Border zones.—The outermost, or border, zones of pegmatites in the area are very fine to fine grained and a fraction of an inch to 3 inches thick. Zoned fracture fillings, or pegmatites that cut other pegmatites, do not have this finer grained zone at their outer border.

Most of the border zones consist of quartz, plagioclase, and muscovite, but many are made up mainly of quartz and muscovite. Tourmaline, apatite, and minor amounts of other minerals are the common accessories. In some places these minerals are very abundant; at the Hunter-Louise mine (No. 8, G–2), for example, the border zone locally consists of nearly pure, green apatite and a small amount of muscovite and tourmaline.

Wall zones.—Wall zones are generally continuous around the entire pegmatite. The one major exception is at the Hunter-Louise pegmatite, where locally an inner spodumene-bearing zone occurs within 1 inch of the contact. The thickness of wall zones is variable, but most wall zones are about 1–10 feet thick. In pegmatites 75–77 (C–7, D–8) the wall zones are only about 6 inches to 1 foot thick, but the thickness of the individual pegmatites are less than 3 feet.

The composition of the different wall zones varies considerably; most are rich in plagioclase, quartz, and muscovite. Perthite and tourmaline may be major minerals (>5 percent). Perthite is most abundant in wall zones in the southern part of the quadrangle; many wall zones in the northeastern part of the area contain little or no perthite. The mineral assemblages in the wall zones are summarized in table 8.

The mineralogy of a few of the wall zones varies somewhat with structural position. Wall zones of gently dipping bodies tend to have more perthite on the hanging-wall side than along the footwall. That the perthite content of the wall zones tends to decrease slightly with depth is shown in pegmatite bodies exposed in areas of considerable relief. Tourmaline is also locally more abundant in the wall zones on the hanging-wall

TABLE 8.—*Mineral assemblages of some wall zones*

[Minerals are arranged in order of decreasing abundance. Data for the High Climb from Sheridan (1955); data for the Crown, Highland Lode, and Hunter-Louise pegmatites from Page and others (1953)]

Assemblage	Common accessory minerals	Pegmatites (by number) shown on plate 34
Plagioclase, quartz, muscovite.	Perthite, tourmaline, apatite, beryl, garnet, phosphate minerals.	1, 17 (Sky Lode), 19, 21, 23, 40 (Crown), 56 (Dorothy), 65, 85.
Plagioclase, quartz, perthite, muscovite.	Tourmaline, apatite, beryl, garnet, phosphate minerals.	51, 57, 59, 61, 76, 77, 87.
Plagioclase, quartz, perthite.	Muscovite, tourmaline, apatite, beryl, garnet, phosphate minerals.	5, 25, 26 (Wilhelm), 27. (Milton), 62, 86 (Inca).
Plagioclase, quartz	Perthite, muscovite, tourmaline, apatite, beryl, garnet, phosphate minerals.	8 (Hunter-Louise), 9 (High Climb), 63 (Highland Lode).
Quartz, muscovite, plagioclase.	Tourmaline, apatite, beryl, cassiterite).	3, 6.
Plagioclase, quartz, tourmaline, muscovite.	Apatite, garnet	50 (Big Spar No. 1).
Plagioclase, quartz, perthite, tourmaline, muscovite.	Beryl, apatite, garnet	74. [1]

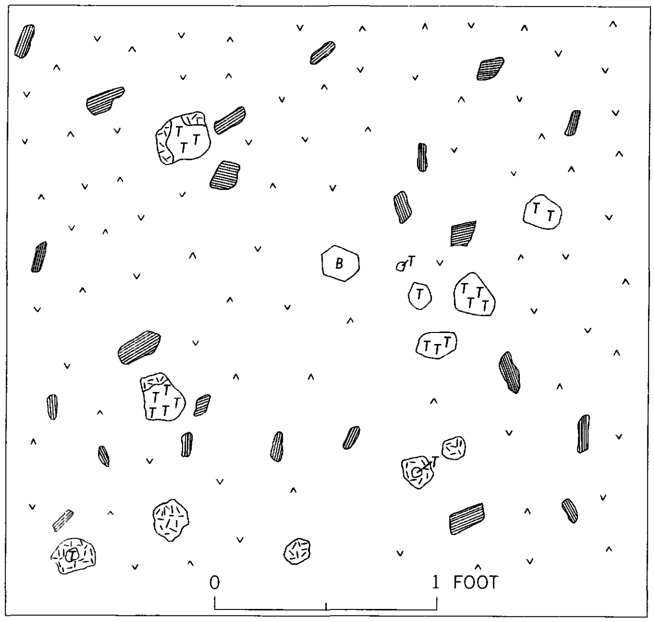
[1] Hanging-wall part of wall zone has an estimated 10 percent tourmaline, and the footwall, only 3 percent tourmaline.

side of some pegmatites, as in pegmatite 74 (C–7). Marked concentration of large tourmaline crystals occurs in the crestal part of the Highland Lode pegmatite (No. 63, D–9).

The wall zones have a variable texture but are generally fine to medium grained. Parts of the wall zones of pegmatites such as the Wilhelm and the Milton are very fine grained. Sugary-textured albite is locally concentrated in the wall zones, especially in lithium-bearing pegmatites. In the Crown mine, individual muscovite books are as much as 18 inches across in the inner part of the wall zone (Page and others, 1953, p. 95). The minerals in the wall zone are generally oriented with their long dimension normal to the contact, as in figure 131. Figure 131 also shows that tourmaline may be partly or entirely altered to muscovite in some wall zones, especially on the hanging-wall side.

Intermediate zones.—Intermediate zones of mappable size generally occur only in the larger pegmatites, although thin mineralogic equivalents of some of the same zones are commonly present in the smaller pegmatites. Intermediate zones are only a few feet thick in the smaller pegmatites and about 20 feet thick in the larger ones. Only nine pegmatites in the area contain intermediate zones that are large enough or sufficiently distinctive in mineralogy to note in table 9.

The different intermediate zones are rich in perthite, quartz, plagioclase (generally cleavelandite), and muscovite, but there is marked variation in the mineralogy and texture. The perthite-rich intermediate zones tend to pinch out downdip. The cleavelandite-bearing intermediate zones all have some accessory lithium minerals and are presumably transitional into the regular lithium-bearing intermediate zones described by Page and others (1953) elsewhere in the Black Hills.



EXPLANATION

Muscovite pseudomorphous after tourmaline

Outline of beryl crystal with *C*-axes almost perpendicular to sketch

Tourmaline crystals with *C*-axes almost perpendicular to sketch

Muscovite with cleavage almost perpendicular to sketch

Quartz and plagioclase

FIGURE 131.—Mineral orientation and tourmaline altered to muscovite in outer part of wall zone. Direction of view is perpendicular to contact. Pegmatite 30 (pl. 34, E–5).

TABLE 9.—*Mineral assemblages of some intermediate zones*

[Minerals are arranged in order of decreasing abundance. Data for the High Climb from Sheridan (1955); data for the Crown and Wilhelm from Page and others (1953)]

Assemblage	Common accessory minerals	Pegmatites (by number) shown on plate 34
Plagioclase, quartz, muscovite.	Tourmaline, apatite, beryl, perthite, garnet, phosphate minerals.	9 (High Climb 1st),[1] 26 (Wilhelm),[2] 31, 87.
Perthite, quartz, plagioclase.	Muscovite, tourmaline, beryl, and apatite, amblygonite, garnet.	9 (High Climb 2d) 63 (Highland) 57, 65, 85.
Quartz, cleavelandite	Muscovite, tourmaline, beryl, apatite, amblygonite, phosphate minerals, columbite-tantalite.	9 (High Climb 3d).
Cleavelandite, quartz, muscovite.	Tourmaline, apatite, beryl, amblygonite.	40 (Crown).

[1] High Climb 1st, 2d, and 3d refer to position, inward from the wall zone, of the intermediate zone of the High Climb pegmatite.
[2] The Wilhelm has a very thin quartz-muscovite-plagioclase intermediate zone rather than plagioclase-quartz-muscovite.

In intermediate zones the minerals are mostly medium to very coarse grained. Perthite crystals are as much as 6 feet across at the Highland Lode, and cleavelandite forms radial aggregates as much as 3 feet in diameter at the Crown pegmatite (No. 40, F–7).

Cores.—The core of a pegmatite is defined as its innermost zone (Cameron and others, 1949, p. 20). The innermost exposed zone may be only an apparent core, and the true core may be concealed. Where the core as described in table 10 is mineralogically similar to intermediate zones, the true core may not be exposed.

The quartz cores seem to be an end assemblage of the zones, as was concluded by Cameron, Jahns, McNair, and Page (1949) and by Page and others (1953). The sequence of zones found by Sheridan (1955) at the High Climb pegmatite suggests that the feldspar-quartz assemblages are intermediate stages leading to quartz cores.

Some variations in the mineral assemblages of the cores given in table 10 are a consequence of including thin intermediate zones with the cores in very narrow bodies of pegmatite. The perthite-quartz-cleavelandite-muscovite cores of pegmatites 75–77 might, in larger pegmatites, be divided into two or more units.

FRACTURE FILLINGS

Fracture fillings are abundant in pegmatites of the Berne quadrangle. They range in thickness from a few inches to at least 20 feet, and their lengths are as much as several hundred feet. These fracture fillings are confined to pegmatites; none are known to penetrate the metamorphic rocks.

Contacts of the fracture fillings and the host pegmatite are sharp and are apparently controlled by joint planes. Where several fracture fillings occur in a single large host body, they generally parallel one another. Pegmatites 16 and 17 are parallel fracture fillings in the much larger pegmatite 15 (G–3). Many fracture fillings are oriented approximately normal to the plunge of the host pegmatite; some occupy fractures between separate boudins of pegmatite (fig. 132), as described in the adjacent Fourmile quadrangle by Redden 1963, p. 257). Some of the more irregularly shaped fracture fillings clearly offset internal structural units in the host pegmatite.

Some of the fracture fillings contain thin zones. Pegmatite 17 (Sky Lode, G–3) is a fracture filling about 5 feet thick that consists of a thin wall zone of plagioclase, quartz, and mica; an intermediate zone of cleavelandite and quartz; and a core of quartz and spodumene. A nearby fracture filling (No. 16, G–3) has an inner zone of perthite and quartz and an outer zone of plagioclase, quartz, perthite, and muscovite. In host pegmatites containing several zones, the fracture fillings cross only the outer zones and contain minerals characteristic of the inner zones. For example, the Rachel D pegmatite (No. 43, F–7) has a quartz core and also has abundant quartz fracture fillings that cut across the wall zone.

In the homogeneous pegmatites (or those lacking any internal zones except a border zone), fracture fillings range from nearly pure quartz to mixtures of quartz and perthite, perhaps with small amounts of plagioclase and muscovite. Most such fracture fillings are coarser grained than the host rock. The perthite in the fracture fillings is free from quartz inclusions, whereas perthite in the host pegmatite is commonly a graphic intergrowth with quartz. Beryl is common at the edges of the fracture fillings.

The mineralogy of most of the fracture fillings suggests that they are closely related genetically to their host pegmatite. Some, such as the Sky Lode and the spodumene-bearing fracture fillings in pegmatite 23 (G–4), have no evident mineralogic relationship with

TABLE 10.—*Mineral assemblages of some cores or innermost exposed zones*

[Minerals are arranged in order of decreasing abundance. Data for the High Climb from Sheridan (1955); data for the Crown, Hunter-Louise, and Highland, Lode from Page and others (1953)]

Assemblage	Common accessory minerals	Pegmatites (by number) shown on plate 34
Quartz	None [1]	9 (High Climb), 40 (Crown), 43 (Rachel D.), 70, 74, 85, 87 (High View).
Quartz, perthite, plagioclase.	Muscovite, beryl, tourmaline, apatite, garnet, phosphate minerals.	26 (Wilhelm), 50, 51, 56 (Dorothy), 59, 61, 86 (Inca).
Quartz, perthite	Plagioclase, muscovite, beryl, apatite, garnet, phosphate minerals.	19, 57, 62, 63 (Highland Lode), 65.
Plagioclase, quartz, spodumene, perthite.	Muscovite, apatite	2, 5, 25, 27 (Milton).
Quartz, spodumene	Plagioclase, muscovite, perthite, phosphate minerals, columbite-tantalite.	8 (Hunter-Louise), 17 (Sky Lode).
Perthite, quartz, cleavelandite, muscovite.	Beryl, tourmaline, apatite, amblygonite.	76, 77.

[1] Locally perthite, spodumene, and muscovite are accessory.

FIGURE 132.—Boudinage structure in sill of pegmatite. Light area below hammer point is part of a quartz fracture filling between the boudins.

the host pegmatite and may be separately emplaced pegmatites. However, none of the fracture fillings extend into the metamorphic country rocks.

LAYERED PEGMATITES

Many pegmatites in the Berne quadrangle consist largely or partly of alternating layers having contrasting textures and mineralogies that may be repeated many times in a single exposure. The layers are parallel to the nearest contact of the pegmatite and country rock; however, in contrast to zones, the layers commonly merge along strike and fade into a more homogeneous variety of pegmatite. Layered pegmatites are especially characteristic of the areas of abundant pegmatites, as in the southeastern part of the quadrangle and in the area of granite and pegmatite around Harney Peak.

The layered pegmatites consist largely of plagioclase (An_{6-14}), quartz, perthite (generally in graphic intergrowths with quartz), and muscovite. Tourmaline, apatite, and garnet are the most common accessory minerals, although beryl or other scarce minerals may occur in fracture fillings. The estimated average composition of layered pegmatites is the same as in the Fourmile quadrangle—that is about 40 percent plagioclase, 31 percent quartz, 23 percent perthite, 4 percent muscovite, 1 percent tourmaline, and 1 percent other accessory minerals (Redden, 1963, p. 237).

The layers are of two types, having the same mineralogy and structure as described in the Fourmile quadrangle (Redden, 1963). The dominant variety consists of alternating coarse-grained and fine-grained layers that are several inches to as much as 10 feet thick. The less common type of layering is the "line rock" of Schaller (1925, p. 273). Line rock in the Berne quadrangle consists of layers that are each of uniform mineralogical composition and are generally about 1 cm thick, though they range in thickness from a few millimeters to about 10 cm. Many line-rock layers persist for tens of feet along strike and a few extend almost the entire length of the host pegmatite. Some pegmatites have both kinds of layering.

In pegmatites with alternating coarse- and fine-grained layers, the coarse-grained layers consist principally of graphic perthite, quartz, and plagioclase, whereas the fine-grained layers contain mainly plagioclase, quartz, and, commonly, muscovite. The coarse-grained layers of the Calamity Peak area, a few miles northeast of Custer (fig. 123), contain a larger percentage of potassium and a smaller percentage of sodium than do the fine-grained layers (Redden, 1963, table 8), and similar differences probably occur in the layers in the Berne quadrangle. Pegmatite 57 (F–9) and many of the pegmatites at and near Thunderhead Mountain have excellent examples of this layering.

Line rock is very fine grained and is variable in composition. Light-colored layers rich in plagioclase and quartz (less commonly microcline) alternate with darker layers that are relatively rich in tourmaline or, rarely, muscovite and garnet in addition to quartz and plagioclase. Tourmaline, muscovite, and garnet may also be present in the light-colored layers, but in small amounts. Variations in the tourmaline content of adjacent layers is generally more evident than the variations in other minerals. In pegmatite 72 (D–7), however, the contrast is between garnet-rich and garnet-poor layers, as at the Pala, Calif., locality described by Schaller (1925). Some line rock has alternating muscovite-rich and muscovite-poor layers. The line-rock layers clearly have chemical differences, chiefly in the content of alkalies, iron, boron, and water. Structurally, the layers are generally simple, but some line-rock layers are curved and crumpled and may bend around large perthite crystals. Details of line-rock structures in the Fourmile quadrangle indicate that the layers formed early and grew from the outer contacts of the host pegmatite inward toward the center. Line rock is common in many pegmatites in the Berne quadrangle; it is especially abundant in pegmatite 46 (F–8) and in some of the larger pegmatites nearby.

HOMOGENEOUS PEGMATITES

Homogeneous pegmatites, lacking any conspicious internal structure are the most numerous variety in the Berne quadrangle and are distributed throughout the pegmatite-bearing area. They have border zones, similar to those in the zoned pegmatites, and are incompletely homogeneous—that is, they have small local variations in grain size and composition. Most of these pegmatites are fairly thin and form long tabular sills. They are composed of the same minerals, in approximately the same proportions, as the layered pegmatites. Estimates of modes made on various pegmatites both in the Berne and the Fourmile quadrangles, however, suggest that the homogeneous pegmatites contain a slightly larger percentage of perthite and a correspondingly smaller percentage of plagioclase. Homogeneous pegmatites also tend to have more nongraphic perthite than the layered ones, and they are generally somewhat coarser grained, chiefly because they lack the fine-grained layers of the layered pegmatites.

GNEISSIC PEGMATITES OF THE BEAR MOUNTAIN AREA

The relatively few pegmatites in the Vanderlehr Formation near Bear Mountain are sill-like and have been deformed and partly recrystallized. Although some of these gneissic pegmatites have large outcrop areas, they are actually thin gently dipping sills exposed on dip slopes; pegmatite 89 is a good example. Some gneissic pegmatites are concordant in strike but dip more gently than the enclosing schists. In general the size and shape of the pegmatites appear to be comparable to others in the quadrangle. Many of the gneissic pegmatites seem to have been layered pegmatites prior to deformation.

The pegmatites consist of both fine-grained to very fine grained phases resembling granite and coarser grained pegmatitic phases. A cataclastic structure parallel to the foliation in the country rock is evident. The micas are poorly oriented in the finer grained phases, and examination of thin sections shows that the quartz and plagioclase are considerably broken and altered. The coarser grained phases do not appear to be very deformed, mostly because of the coarser grain size. Perthite crystals 6–8 inches in diameter are broken, sheared, and cut by sericite laminae.

Quartz fracture fillings or veins are abundant; most seem to have formed after the deformation, for their associated minerals are undeformed. Some have large rosettes of black tourmaline crystals; yet no remnants of fresh tourmaline were observed in the host pegmatite.

The gneissic pegmatite consists largely of perthite, quartz, and plagioclase. The common accessory minerals are muscovite, biotite, garnet, and apatite. Much of the very fine grained pegmatite is rich in slightly perthitic microcline, and the associated plagioclase (An_{10-15}) is commonly filled with many inclusions of sericite and apatite. Apatite is disseminated through and around all the minerals.

The composition of the gneissic pegmatites is similar to that of the other pegmatites in the quadrangle but may be somewhat richer in potassic feldspar. Tourmaline is not a notable accessory, but the black tourmaline in the quartz fracture fillings or quartz veins suggests the possibility of removal and recrystallization of earlier tourmaline.

MINERALOGY

The following minerals are found in the pegmatite: Plagioclase, quartz, perthite (also graphic perthite and microcline), muscovite, tourmaline, apatite, garnet, beryl, spodumene, amblygonite, biotite, lithiophilite-triphylite, graftonite, columbite-tantalite, cassiterite, uraninite, secondary uranium minerals, iron-manganese phosphate minerals, loellingite, allanite, chalcopyrite, chrysoberyl, and sillimanite.

Plagioclase

Plagioclase is the most common and the most abundant mineral in nearly all the pegmatites in the Berne quadrangle and adjacent areas. It is generally very fine to fine grained in the layered pegmatites and the homogeneous pegmatites and occurs as chalky-white to bluish-gray anhedral crystals. Rarely, in the zoned pegmatites, plagioclase occurs as platy cleavelandite, which forms rosettelike aggregates that are commonly 1 inch or more across. The anhedral or blocky plagioclase in the zoned pegmatites is fine to medium grained.

The composition of the plagioclase ranges from An_3 to An_{15}. The anorthite content tends to be low in areas where pegmatites are sparse: plagioclase in pegmatites near the southeast corner of the quadrangle has a composition of about An_{12-14}, according to the refractive indices of cleavage fragments, whereas that in pegmatives several miles to the north and west is mostly in the range An_8 to An_{12}. Most of the zoned pegmatite have plagioclase with slightly lower indices of refraction than nearby homogeneous or layered pegmatites. In fracture fillings the plagioclase tends to have lower indices than it does in the host pegmatite. Similar changes in the plagioclase composition were noted in the adjacent Fourmile quadrangle (Redden, 1963, p. 242).

Quartz

Quartz occurs in all the pegmatites in the area and is the dominant mineral in many outcrops. It occurs as cloudy white anhedral grains or massive aggregates or as graphic intergrowths with perthite. In some of the massive quartz-rich cores of zoned pegmatites, cleavage surfaces are clearly visible and the quartz is obviously coarse grained. In the layered and homogeneous pegmatites, most of the quartz is very fine to fine grained and is comparable in shape and size to the associated plagioclase.

Perthite, microcline, and graphic perthite

Nearly all the pegmatite contains perthite or perthite graphically intergrown with quartz, and the very fine grained to fine-grained gneissic pegmatite around Bear Mountain has abundant microcline. A small amount of fine-grained microcline occurs in the interstices between quartz and plagioclase grains in some of the fine-grained layers in the layered pegmatites.

The perthite occurs as creamy-white to tan subhedral to euhedral crystals that occur as large phenocrysts in a finer grained matrix of quartz, plagioclase, and other minerals. It is most abundant in the zoned pegmatite bodies and in fracture fillings. In the zoned pegmatites it is commonly coarse to very coarse grained; single crystals are 6 feet or more long. However, the fine-grained spodumene-bearing zoned pegmatites 5, 7, 25, and 27 (G–2, F–4), have perthite crystals only an inch or a few inches in diameter and of the same size as, or smaller than, perthite crystals in some of the layered and homogeneous pegmatites.

Graphic perthite is most abundant in layered and homogeneous pegmatite. It is similar to other perthite in appearance but is not as coarse grained.

Muscovite

Muscovite, like quartz and plagioclase, occurs in all the pegmatites and in all the internal units. In the homogeneous and layered pegmatites, it generally forms very fine to fine grained flat ruby-red to colorless crystals of subhedral to euhedral outline. In the muscovite-rich zones the crystals are generally much larger—several inches to, rarely, a foot or more across. Many of the flat muscovite crystals near the outer contact of the pegmatites are perpendicular or subperpendicular to the contact, as shown in figure 131. A less common feature, also illustrated in figure 131, is very fine grained aggregates of muscovite pseudomorphous after tourmaline. Fine-grained sericitelike muscovite occurs in some of the gneissic pegmatites near Bear Mountain and in aggregates pseudomorphous after spodumene in a few of the spodumene-bearing pegmatites.

Spodumene

At least seven of the zoned pegmatites in the Berne quadrangle contain spodumene, and another five are known to contain altered spodumene. The spodumene constitutes 15–25 percent of spodumene-bearing zones, in which it occurs as chalky-white to pale-green lathlike crystals. In the Hunter-Louise, Tenderfoot Spud (G–3), and Sky Lode (G–3) pegmatites (pl. 1, 34), the crystals are medium to coarse grained; the largest are about 4 feet long and 6 inches or more across. However, the Milton pegmatite (No. 27, F–4) and pegmatites 5, 7, and 25 have spodumene crystals that are much smaller than those typical of other pegmatites in the Black Hills. Many of the crystals are only 1 inch or more long, and the largest are only 6–8 inches long. Nearly all these smaller spodumene crystals are perpendicular or subperpendicular to the outer contact of the pegmatite. The spodumene in pegmatite 23 (G–4) occurs as a few small crystals in a thin quartz-rich fracture filling that crosses the zoned host pegmatite.

Although much of the altered spodumene retains the original grain boundaries, the boundaries of some grains have been modified considerably. Spodumene at the High Climb (G–2) and Highview (C–9) mines is altered to claylike material, but the former spodumene crystal faces are distinct. The spodumene parentage is not so easily detected in the Milton pegmatite and pegmatites 5, 7, and 25, in which spodumene in the outer parts of the pegmatites is altered to muscovite and a claylike mineral having a ϵ index of refraction of about 1.54. These micaceous minerals form a distinctive felted intergrowth; where this texture was noted, additional search generally resulted in the discovery of fresh spodumene in the interior of the pegmatite. The lathlike outline of spodumene is crudely, but noticeably, reproduced by the micaceous aggregates. Apparently all gradations occur from well-formed pseudomorphs mostly of clay (?), such as at the High Climb, to aggregates of muscovite blades in which the pseudomorphism is less readily apparent. Felted mica aggregates found in the small pegmatite just east of pegmatite 7 and in pegmatite 35 (G–6) are not accompanied by fresh spodumene in the present exposures, but spodumene may well be present at depth in these pegmatites.

Other minerals

Tourmaline was observed in nearly all the pegmatite bodies and probably occurs in all of them. However, it is sparse in the gneissic pegmatite around Bear Mountain, in which it forms radial aggregates in quartz-rich fracture fillings. In other types of pegmatite it forms black subhedral to euhedral rodlike crystals that range in size from microscopic to about 10 inches in diameter. Crystals several feet long occur in the Highland mine (D–9). Tourmaline tends to be concentrated near the hanging wall, where it is commonly oriented subperpendicular to the contact (fig. 131). Crystals along the hanging walls of some of the pegmatites are replaced in part or entirely by muscovite, the replacement beginning on the outer edges of the crystals (fig. 131). Green tourmaline occurs in the Hunter-Louise (G–2) pegmatite in association with lithium minerals. This association is common in other pegmatites in the Black Hills.

Apatite occurs as green to blue-green generally subhedral crystals that are generally less than 0.1 inch across, but in some zoned pegmatites they are several inches across. Thin sections of the gneissic pegmatite showed that tiny blebs of apatite are disseminated throughout much of the rock. Apatite is abundant in many of the border zones, especially at the Hunter-Louise mine; it is also a common accessory mineral in most of the other zones.

Garnet forms small reddish-brown crystals generally less than 2 inches across and occurs in most of the pegmatites in the southern Black Hills. In the zoned pegmatites it forms fractured and broken crystals or aggregates of crystals several inches in diameter. Most of the garnet crystals are imperfect and are commonly indented by adjacent minerals, especially muscovite.

Beryl was noted in most of the zoned pegmatites and in fracture fillings in several of the other types of pegmatite. The distribution of the pegmatites in which beryl was found is shown in figure 131. Most of the beryl occurs near the outer limit of the pegmatite area. It is generally yellowish white to pale green but some is deep green. Golden beryl, some of which is of gem quality, occurs at the Highland Lode mine. Nearly colorless beryl in fracture fillings in pegmatite 47 (F–8) has a good basal cleavage. Many crystals are euhedral and range in size from those that are only 0.1 inch in cross section to more than 2 feet across at the Highland Lode pegmatite. Beryl is most abundant along the inner part of muscovite-rich wall zones, but it is also a common accessory in many of the intermediate zones. It is abundant in fracture fillings in both layered and homogenous pegmatites but is generally not noticeable in the host pegmatite. Seldom does the beryl content of any zone exceed 1 percent, and even in a rich zone the distribution is very spotty, as the high-grade concentrations of beryl are separated by barren rock. In pegmatite 30 (E–5), which has a muscovite-rich wall zone, much of the beryl is in tapered shell crystals in which the larger end of the crystal consists of only an outer shell of beryl surrounding and intergrown with plagioclase, quartz, and muscovite. Shell beryl is also common in the Crown mica mine. Page and others (1953, p. 53) noted that the shell texture tends to occur in mica-rich pegmatites or zones.

The data on the other minor accessory minerals are summarized in table 11. These minerals occur only in a single pegmatite or in a very few pegmatites.

ORIGIN

Recent extensive work by many geologists has shown that the zoned pegmatites have crystallized from their walls inward and apparently represent crystallization and differentiation in virtually closed systems (Cameron and others, 1949; Page and others, 1953). The homogeneous and layered pegmatites also have such structures as tapered and oriented crystals which indicate that crystallization took place from the contacts inward. In the Fourmile quadrangle, evidence that layered structures formed as a result of variations in the physical conditions (perhaps chiefly in vapor pressure) in the crystallizing pegmatite led to the conclusion that the layered pegmatites are more closely related to open systems (Redden, 1963). Evidence from the Fourmile quadrangle also indicated that the zoned pegmatites represent a more advanced phase of differentiation from

TABLE 11.—*Description of minor accessory pegmatite minerals in the Berne quadrangle*

Mineral	Color	Size and habit	Occurrence
Biotite	Black to greenish black	Variable; occurs as small books and lath-shaped crystals.	Found in scattered pegmatites throughout the area; common in the perthite-rich zoned pegmatites.
Lithiophilite-triphylite.	Brownish green to nearly black; commonly altered.	Variable; forms equidimensional crystals and irregular aggregates.	All of larger zoned pegmatites such as the High Climb, Hunter-Louise, Wilhelm, and so forth. Generally found in inner part of wall zones or in intermediate zones.
Graftonite	Clove brown	Thin striated plates intergrown with triphylite. Individual intergrowths are as much as 1 foot in diameter.	Found only in Highland Lode pegmatite.
Amblygonite	Chalky white	Equant subhedral crystals 1 inch–2 feet across.	Intermediate zones in High Climb, and as isolated crystals in inner zones of the Wilhelm, Highland Lode, Highview, and Inca pegmatites.
Columbite-tantalite	Dull to shiny black	Thin plates, or blades, an inch or several inches across.	Same as amblygonite.
Cassiterite	Brownish black	Equant crystals less than 0.5 inch across.	Wall zones of pegmatites 2, 3, 4.
Uraninite	Dull black	Irregular to blocky aggregates	Highland Lode pegmatite (inner zone?).
Secondary uranium minerals.	Yellow; orange	Occurs as thin crust and veinlets filling fractures and coating minerals.	Highland Lode pegmatite (wall zone).
Loellingite	Dull silvery gray	Spear-shaped euhedral crystals	High Climb, Highland Lode, and especially abundant at pegmatite 33 (pl. 34, E–5).
Allanite	Shiny black	Rounded aggregates	Inner wall zone of pegmatite 83 (pl. 34, B–9).
Chalcopyrite			High Climb.[1]
Chrysoberyl		Subhedral microscopic grains	Tiny grains in thin section. Border zone of 10-inch-thick pegmatite stringer just east of pegmatite 34 (pl. 34, G–5).
Sillimanite		In needles replacing(?) muscovite along fractures.	Same as chrysoberyl.

[1] Described by Sheridan (1955).

the parent mass than do the other types. This factor may have been as important as the closed nature of the system in influencing crystallization and differentiation.

The pegmatites of the Berne quadrangle were apparently emplaced along such planes of weakness as bedding, foliation, and, rarely, joints. The pegmatite distribution indicates that they were derived from the same source as the pegmatite and granite of the Harney Peak area. Although the pegmatite distribution generally parallels the metamorphic intensity, the author has shown in the adjacent areas that the distribution of pegmatite is at least in part independent of the metamorphism (Redden, 1963).

Sillimanite instead of muscovite is the high-alumina mineral of many pegmatites in the Harney Peak area, east of the quadrangle. This fact implies that water-vapor pressure was too low for muscovite to form from the pegmatite fluid. In the eastern part of the Berne quadrangle, sillimanite is present instead of muscovite along fractures in some of the pegmatite, also suggesting that the water-vapor pressure in the surrounding rocks was low.

PALEOZOIC AND YOUNGER ROCKS

DEADWOOD FORMATION

The Deadwood Formation, consisting largely of sandstone but containing minor amounts of conglomerate and shale, unconformably overlies the Precambrian crystalline rocks along most of the extreme west side of the Berne quadrangle (pl. 34). The formation crops out around the entire Black Hills. Its type locality is at Deadwood in the northern Black Hills. On the basis of fossils, Darton and Page (1925, p. 7) concluded that the Deadwood is Upper Cambrian.

Outcrops of the Deadwood Formation are sparse. The formation forms fairly steep slopes, but these are masked by soil and talus and some colluvium derived from cliffs of overlying limestones. The Precambrian rocks are deeply weathered below the unconformity in the adjacent Fourmile quadrangle and are probably similarly weathered in the Berne quadrangle. The lower contact of the Deadwood Formation is concealed throughout the quadrangle, but its position can readily be identified by a pronounced steepening of the slope that is in contrast to the gentle slopes on the Precambrian rocks. This steep slope continues upward to the top of the formation, where a small bench marks the contact with the overlying thin-bedded limestones.

The Deadwood ranges in thickness from about 160 feet to 220 feet. The formation gradually thickens from the southern Black Hills to its type locality, but most of its thickness variation in the Berne quadrangle is local and is apparently due to the relief on the underlying unconformable surface. The base of the Deadwood outlier in the southern part of sec. 27 (pl. 34, A–3), is at least 60 feet lower in altitude than is to be expected from the regional dip. The basal conglomerate of the Deadwood exposed in this outlier contains fragments from the quartz-tourmaline veins of the Vanderlehr Formation these fragments were probably deposited in a topographic low on the unconformable surface.

The lowermost part of the Deadwood Formation is a basal conglomerate, generally less than 5 feet thick, that commonly contains locally derived fragments. Above this conglomerate is a fairly thick section of pale-maroon sandstones, thin conglomeratic lenses, greenish-red glauconitic sandstones, and a few greenish-gray shale lenses. The upper part of the formation consists of at least 10–20 feet of sandstone, composed of coarse very well rounded grains, overlain by lenses of shale.

ENGLEWOOD FORMATION

Above the Deadwood Formation is the thin-bedded Englewood Formation of Devonian and Early Mississippian age. The Englewood is better exposed than the Deadwood Formation, but it, too, is largely covered by blocks and soil from the overlying massive limestone. It crops out on prominent ridges or steep slopes. The thickness is about 45 feet and probably ranges from 40 to 55 feet. No systematic variations in thickness have been detected, and no apparent discordance was noted between the Englewood and the Deadwood Formation, despite the great stratigraphic break. The Englewood consists of relatively uniform thin-bedded pebbly to slabby highly fossiliferous limestone interbedded with some dolomite. The lavender to mauve color and the nodular structures caused by the abundant fossils are especially noticeable characteristics of this limestone. Some beds contain many small limonite-filled concretions.

PAHASAPA LIMESTONE

The Pahasapa Limestone is Early Mississippian in age (Darton and Paige, 1925). Its thick massive beds dip about 2½° W. The more massive parts of the formation form cuestas that have nearly vertical cliffs as much as 80 feet high. The limestone also caps Paleozoic outliers, such as Round Mountain and the oval hill in sec. 27 (plate 34, A–9). The contact between the Pahasapa and Englewood Limestones is not exposed but can be located within a few feet. It is apparently conformable but is marked by an abrupt change from lavender thin-bedded Englewood to massive thick-bedded Pahasapa Limestone.

The upper contact of the Pahasapa is absent in the Berne quadrangle. The maximum exposed thickness of the formation is 200 feet on Bear Mountain, but to the west, in the Signal Hill quadrangle, it is 400–500 feet.

The Pahaspa Limestone is pale buff to light gray and forms beds 1–8 feet thick. Fossils, consisting mainly of branchiopods and crinoid stems, are abundant in a few beds but are not nearly so abundant as are the fossils in the Englewood Formation.

QUATERNARY AND RECENT DEPOSITS

Along Spring Creek there are a few small deposits of terrace gravel at altitudes ranging from a few tens of feet to as much as 270 feet above the present alluvium-covered floor of Spring Creek Valley. Many of these deposits are a plaster on the noses of small ridges and have a vertical extent of as much as 70 feet; most of these deposits are probably only a few feet thick or few tens of feet thick. The gravels consist of well-rounded boulders, generally ranging from a few inches to 10 inches across, in a matrix of coarse sand. Many of the boulders consist of rock derived from the Vanderlehr Formation on Bear Mountain and surely were deposited by a stream ancestral to Spring Creek.

The age of the terrace gravels cannot be accurately determined from fossils or other evidence, but presumably the gravels are Quaternary. These terrace gravels occur at so many different altitudes that they do not appear to be related to a major period of terrace formation.

Several feet of soil and alluvium cover coarse gravel and sand along the major creeks, such as Spring Creek and French Creek. These deposits, presumably of Recent age, have a total thickness of probably not more than 10–15 feet.

STRUCTURE

The structure of the Precambrian rocks in the Berne quadrangle is generally complex, both in detail and in the overall relationships. The details of some of the structures are not clearly understood and may therefore, have been misinterpreted. Three periods of deformation have been recognized, though no conclusive evidence has been found that these periods were greatly separated in time.

The major structures in the Berne quadrangle, shown on plate 36, are named after prominent geographic features. The Grand Junction fault trends northwestward across the quadrangle, dividing it into two major structural and stratigraphic parts. The Western Star syncline and Atlantic Hill anticline are two major early folds west of the Grand Junction fault. Several smaller early folds, also shown marked on plate 36, occur both west and east of the fault. The Bear Mountain dome and the Park dome are small features that disrupt the major early folds. The Park dome also deforms the Grand Junction fault and complicates the structure around and between the domes. The Harney Peak dome is a large structure east of the Berne quadrangle that modified early structures in the easternmost part of the area.

Structure of the Paleozoic rocks is uniform, dipping 2°–3° W. away from the Precambrian core of the Black Hills.

MAJOR FOLDS AND DOMES

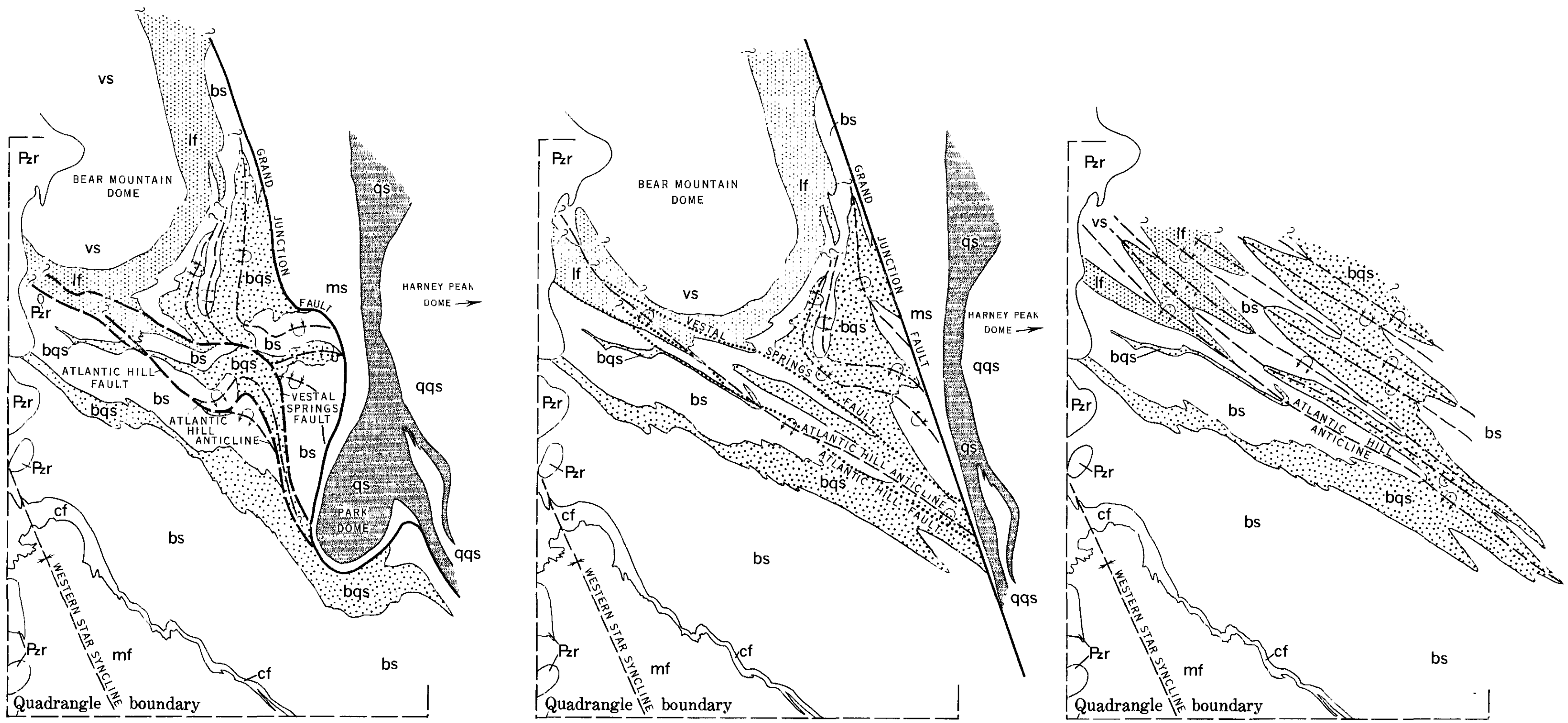
WESTERN STAR SYNCLINE

The Western Star syncline, named for the gold mine near its axis, was first mapped in the Fourmile quadrangle, where the trace of its axial plane extends to the south until it is concealed by Paleozoic rocks (fig. 122). The fold seems to be relatively simple and open (pl. 34, cross section D–D′). It has a vertical to steep west-dipping axial plane that strikes about N. 20° W. The plunge of tight minor folds near the axial plane is about 60° S. and is probably also the plunge of the main fold. To the northeast, away from the trace of the axial plane, the plunge of the minor folds gradually decreases and the attitude of the axial-plane foliation strikes more to the northwest and has a more gentle dip. The pattern of the foliation and of the minor folds on the east limb of the syncline resembles the pattern of a fan fold.

This relatively simple fold was produced by the earliest known deformation in the area. The associated minor folds and lineations and the attitudes of folds suggest a simple structure that was not greatly modified by any later deformation.

ATLANTIC HILL ANTICLINE

The northwest-trending Atlantic Hill anticline adjacent to the Western Star syncline is isoclinal, is overturned to the northeast, and has been considerably deformed (pl. 36; pl. 34, cross sections *B–B′*, *C–C′*). The axial plane is repeated by the Atlantic Hill fault and is warped, both in the area north of Atlantic Hill and in the Park dome area. The southeastern part is presumably cut off by the Grand Junction fault. Graded bedding and the pattern of minor fields in outcrops on the east side of Atlantic Hill, but west of the Atlantic Hill fault, indicate an overturned structure for that part of the fold. The repetition, or largest part, of the trace of the fold axis north and east of the Atlantic Hill fault is inferred mainly on the basis of the interpretation that the Atlantic Hill part of the fold axis has moved southeastward along the Atlantic Hill fault. This fault is believed to have cut across the axial plane at a low angle, and its movement must have been largely in a plane subparallel to the axial plane of the fold. Figure 133 illustrates this interpretation, which is only



A
Present rock distribution and structural interpretation

B
Structure prior to the formation of the Park dome and cross folds in the Atlantic Hill area. Although the fault displacements are restored in the area northeast of Atlantic Hill, the actual displacements may have occurred at many different times. Similarly, deformation around the Harney Peak and Bear Mountain domes may have been contemporaneous with that around the Park dome. Heavy dotted lines indicate locations of later faults. The vertical movement of the faults is unknown and hence is ignored in the diagram

C
Possible structure prior to the doming and formation of the Grand Junction fault. The quadrangle boundaries shown here have no geographic significance because presumably all of the rock units and folds were at some unknown distance to the west prior to faulting

EXPLANATION

Pzr — Paleozoic rocks

mf — Mayo Formation

cf — Crow Formation

bs, bqs — Bugtown Formation
bs, *quartz-mica-feldspar schist*
bqs, *quartz-mica schist*

lf — Loues Formation

vs — Vanderlehr Formation

qqs — Quartzite and quartz-mica schist

qs — Quartz-biotite-garnet schist

ms — Mica schist

Contact

Fault
Dashed where approximately located

Syncline

Overturned anticline

Overturned syncline

FIGURE 133.—Schematic restoration of prefaulting and predoming structure in Berne quadrangle, South Dakota.

one of the many possible ways of interpreting the structure of the Atlantic Hill anticline prior to faulting, doming, and cross folding.

The axial plane of the anticline has a northwesterly strike and a southwesterly dip over most of the area. The bedding attitudes and plunges of minor folds indicate that the main fold plunges 30°–40° S. Northwest of Atlantic Hill the trace of the axial plane is largely inferred, and only the generalized geology of the overturned limb is known. The axial plane of the Atlantic Hill anticline cannot be traced within the Loues Formation nor within any of the units in the Bear Mountain dome.

Most of the small folds on the limbs of the Atlantic Hill anticline have a left-hand pattern on the southwest limb and a right-hand pattern on the northeast limb. There are several uncertainties, however, about some of the larger folds. The syncline and anticline in secs. 15 and 22, 1 mile southeast of the Park dome, are here regarded as left-hand folds on the southwest limb of the Atlantic Hill anticline; however, the anticlinal part of these two folds may be the site of the main Atlantic Hill axis. The outcrop patterns in the general area about 1 mile northwest of Atlantic Hill and south of the Vestal Springs fault indicate the presence of very tight folds that are interpreted as right-hand folds on the northeast limb of the Atlantic Hill anticline (fig. 133). However, some of the elements of the structure of this area, especially the inferred faults, cannot be fully substantiated. Very possibly there are additional faults and either stratigraphic or structural pinchouts that further complicate the structural pattern.

FOLDS BETWEEN THE VESTAL SPRINGS AND GRAND JUNCTION FAULTS

Three large folds—two synclines and an anticline—are indicated by the distribution of lithologic units in the Bugtown Formation between the Vestal Springs and Grand Junction Faults. Near Graveyard Gulch the axial planes have an easterly strike and southerly dip; also, the south limbs of the synclines are overturned. The early minor folds have a moderate plunge in a S. 5°–10° W. direction. The traces of the axial planes of the two south-trending folds are covered by part of the Vestal Springs fault; but, the folds visible to the northwest are interpreted to be a continuation of the folds between the two faults. All three folds are warped about south-plunging axes west of Graveyard Gulch. In the area to the north, the axial planes dip east, and most minor folds plunge moderately south-southeast.

The pattern brought out by these three folds is consistent with the pattern of the Atlantic Hill anticline, and though interpretation of these folds is based on fewer data and involves more inference, the main elements of structure of these folds are less subject to doubt than are the details. Figure 133*A* illustrates how the folds may have fitted together. Axial planes trend mainly to the northwest, but all of them are deformed by cross folds near Atlantic Hill and are disrupted still more by the Bear Mountain dome. The axis of the Atlantic Hill anticline trends mainly northwest on the south flank of the dome, but the axes of the folds north of the Vestal Springs fault trend northward along the east side of the dome. The strikes of the axial planes are deflected to the north by the dome, and the dips are eastward off the east flank of the dome.

PARK DOME

The Park dome is a small elliptical structure in the east-central part of the Berne quadrangle. The name is taken from Park School, which is near the axis of the dome (pl. 34, F–6). A detailed map showing the structural data available is shown on plate 36. The long axis of the dome trends about N. 15° W. The dome is outlined on both the west and the south by the Grand Junction fault, which is deformed by the dome. Along the east side of the dome is a shallow synform that gradually dies out on the north-northeast side of the dome but has a pronounced curvature on the southeast side of the dome. There, its southerly plunge is generally parallel to the major fold structures.

The rocks exposed in the Park dome belong mainly to the quartz-biotite-garnet schist unit and to a thin wedge of the mica schist unit. The Bugtown Formation lies on the west and south sides of the dome. East of the dome, quartz-biotite-garnet schist and the quartzite and quartz-mica schist unit are deformed by early (predoming) folds that plunge into the east side of the dome, probably in the manner shown in the cross sections of plates 34 and 35. A possible restoration of the predoming structure of the Park dome area is shown in figure 133*B*). If the quartzite and quartz-mica schist unit is younger than the quartz-biotite-garnet schist unit (as the previously discussed stratigraphy suggests) some beds on the dome have been overturned more than 180°.

Late minor folds on the north side of the dome plunge in a general N. 30° W. direction. The plunge value decreases to zero near the central part of the dome and then reverses in the southeastern part, where the plunge direction is S. 25° E. These plunge directions are only slightly different from the N. 15° W. direction of the dome axis outlined by the pattern of stratigraphic units.

A belt of many late recumbent isoclinal folds is east of the crest of the dome, particularly southeast of the Crown mine. These folds have amplitudes of only a few feet and are piled one above the other; their axial planes dip gently southeastward, and their limbs are of nearly equal length. Early minor folds in this area plunge northeastward or southwestward and have been deformed by later folds and doming.

BEAR MOUNTAIN DOME

The Bear Mountain dome, in the northwestern part of the quadrangle (pl. 36), is very similar in trend and attitude to the Park dome but is several times larger—at least as now exposed. Only the south half of the dome is exposed in the Berne quadrangle, but reconnaissance indicates that the north half is almost a mirror image (fig. 122).

Rock units exposed in the central part of the dome outline a very simple structural pattern similar to that of the Park dome; but the simplicity is probably deceiving, for small and extremely tight isoclinal folds are present in some parts of the Vanderlehr Formation. The lower and middle parts of the Loues Formation also have a simple pattern, but the upper part has a structurally complex arrangement corresponding to that of the lower part of Bugtown Formation. On the south side of the dome, the upper part of the Loues Formation and the lower part of the Bugtown are inferred to have been displaced by three left-hand faults (pl. 34, cross section *B–B′*) in much the same manner as that illustrated in figure 133*B*. Some vertical displacement along these faults has presumably concealed much of the structure and prevented the tracing of the Atlantic Hill anticline into the area of the dome.

HARNEY PEAK DOME

The Harney Peak dome influences the structural geology of only a small part of the Berne quadrangle—in general, the area east of U.S. Highway 16. The contact of the quartz-mica-garnet schist and the quartzite and quartz-mica schist unit curves gently from a northwesterly strike east of the Park dome (pl. 36) to a northeasterly strike in the northeastern part of the quadrangle. Detailed mapping in the Berne quadrangle and reconnaissance in adjacent areas indicates that this curvature reflects the structure of the large dome around the Harney Peak area, as shown in figure 122, and that the easternmost part of the quadrangle lies on the west flank of the dome. Bedding and foliation dip outward on all sides of this dome and are nearly flat in many areas near its center, where remnants of the country rock are exposed among the many sills and dikes of pegmatitic granite. Mapping by R. G. Wayland (written commun.) in the Hill City quadrangle indicates that axial planes of large isoclinal folds strike north-northwest in areas a few miles from the dome but are warped or cross folded to a northeasterly strike closer to the dome (fig. 122). These folds are overturned toward the dome, are low plunging, and locally reverse their plunge direction where they apparently bow over or are bent by the dome. Some of the digitations in the contact between the quartz-mica-garnet schist and the quartzite and quartz-mica schist unit on plates 34 and 36 may be isoclinal folds, but some of them have so little resemblance to observed smaller folds that they are interpreted to be stratigraphic tongues rather than folds.

SPRING CREEK AREA

The stratigraphic sequence is in doubt in the area near Spring Creek east of the Grand Junction fault, and as a consequence the structure is not well understood. The quartz-biotite-garnet schist is greatly thickened in the northeast corner of the quadrangle. During reconnaissance in the Medicine Mountain quadrangle, to the north, and the work by R. G. Wayland in the Hill City quadrangle, to the northeast, several folds were mapped that seem to be the cause of this great apparent thickness. However, the axial plans of these folds cannot be followed in the Berne quadrangle.

The massive quartzite unit north of Spring Creek dips fairly steeply and strikes a few degrees west of north. The south end of the quartzite seems to be a fold nose, and some lithologically distinctive parts of the mica schist which are traceable for short distances also apparently close to the south around this nose. Minor folds all plunge south at moderate angles. The anticlinal structure shown in cross section *A–A′* of plate 34 is consistent with such a plunge. Nevertheless, if the quartzite is in a fold of any kind, whether anticline or syncline, it creates difficulties in correlation (discussed in the section on stratigraphy) that can be resolved only by additional work in other parts of the southern Black Hills.

MINOR STRUCTURES

Minor structures in the Precambrian rocks of the Berne quadrangle comprise at least three ages of minor folds and their associated planar and linear structures. The minor structures are useful in interpreting major structures in most parts of the quadrangle, but minor structures of different ages are superimposed on major structures in some areas and tend to obscure the pattern. The age of minor structures could not be distinguished in several areas.

EARLY STRUCTURES

The early minor structures include folds, foliation, and various linear elements, such as elongate mineral aggregates, elongate pebbles, and the calc-silicate ellipsoids. Early minor folds are abundant, and the plunge and attitude of many such folds were measured. These folds are generally isoclinal, or at least very tight, and most of their axial planes trend northwest to north. Plunge of the folds is generally moderate to the south, except where it is affected by younger folds and domes. Commonly, a well-developed axial-plane schistosity is readily visible near the noses of the folds but is less distinct on the limbs where it is nearly parallel to bedding. The intersection of this schistosity with bedding produces a lineation in which the mineral grains are conspicuously elongate parallel to the fold axes. In some thin-bedded rock where there is considerable difference in competency of adjacent beds, the schistosity is parallel to bedding in the nose of the fold as well as the limbs. In some varieties of schist both bedding-plane and axial-plane foliations are common. The long axes of calc-silicate ellipsoids parallel the fold axes; also, the ellipsoids are generally oriented with their short axes perpendicular to bedding so that their plane of flattening coincides with bedding planes. However, in some of the minor folds in thick-bedded schist of the Bugtown where the axial-plane schistosity is well developed, the short axis of the ellipsoids is perpendicular to schistosity; hence, the plane of flattening does not coincide with bedding.

The best examples of the early minor folds and early structures are found in the southwestern part of the quadrangle, where they have been unaffected by subsequent deformation (fig. 126). Folds shown by the contacts of the Crow Formation are of this type, and the many diverse bedding attitudes in the upper part of the Bugtown Formation are on parts of these folds. Plunge data for these folds are shown on diagrams *i* and *l* of plate 36.[2] Diagram *j* of plate 36 shows the poles to bedding and also the attitudes of discordant foliation in this same area. The poles to foliation are concentrated along the great circle described by poles to bedding. Bedding shows two maximums: the dominant one represents the long limbs of the folds, which have a northwesterly strike, and the other is caused by the more northerly striking, short limbs of the folds. The maximum for foliation lies between the two maximums for bedding and, thus, clearly represents axial-plane foliation. The folds are left handed, as they should be on the east side of the Western Star syncline. The pole (β) of the great circle for bedding coincides almost exactly with the plunge maximum of each of the diagrams for linear elements. This excellent correlation of plunge, bedding, and foliation indicates that these early folds are virtually cylindroidal, and that in the area covered by these diagrams the structural elements have not been greatly affected by later deformation.

The effects of younger deformation on the early minor structures are more evident in other parts of the quadrangle. The plunge of early folds and linear elements gradually shifts to the southwest, closer to the Park dome, as shown by diagrams *f* and *h* on plate 36. Furthermore, the early folds are clearly deformed by younger folds (figs. 134, 135). The early folds are nearly at right angles to the younger folds, and some (fig. 134) give rise to a drawn-out S-shape in the distribution of plunge determinations of early folds. The early folds are extremely tight (fig. 135), and many thin quartzite beds are deformed into isolated rods or are piled up in the noses of folds, as shown in figure 127. The small folds in figure 127 cross the limbs of much larger, younger folds having wave lengths and amplitudes of about 30 feet.

In the structural trough southeast of the Park dome, the early folds and linear elements produce the b_0 maximum of diagram *k*, plate 36. The bedding attitudes in diagram *k* have a spread that results largely from the effects of the dome. Nevertheless, β from the bedding attitudes nearly coincides with b_0, and the deformation of the older structures near the dome was around the same axis as that of the early folds.

Near Atlantic Hill the general trend of the early minor folds and their axial-plane schistosity is north-northwest. Early minor folds along the east side of the Atlantic Hill anticline are right handed, and those on the west side are left handed. In general, the plunge of the early linear elements here is to the south. However, the bedding attitudes in this area and in areas to the

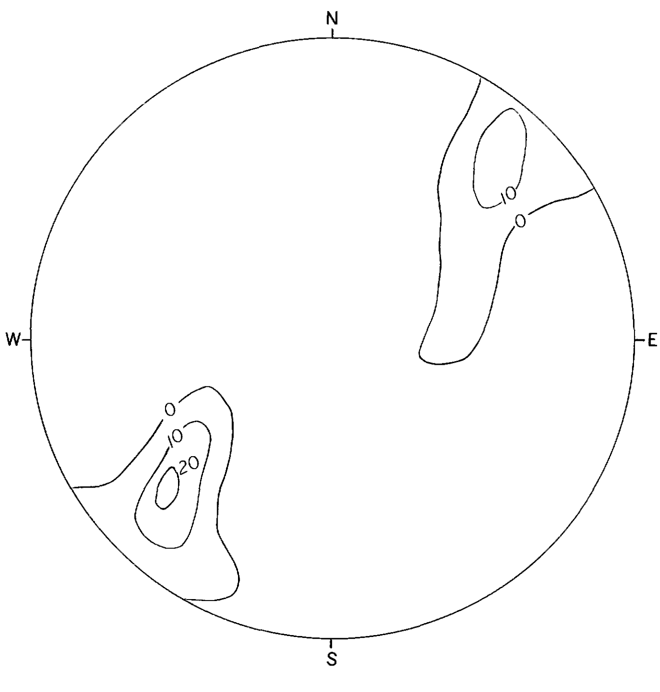


FIGURE 134.—Contour diagram of older linear elements in the Park dome near bench mark 5,752. Most of the elements are small fold axes.

[2] The diagrams on plate 36 also include small numbers of linear elements other than fold axes, but the position of the maximums of other linears do not affect those caused by folds alone.

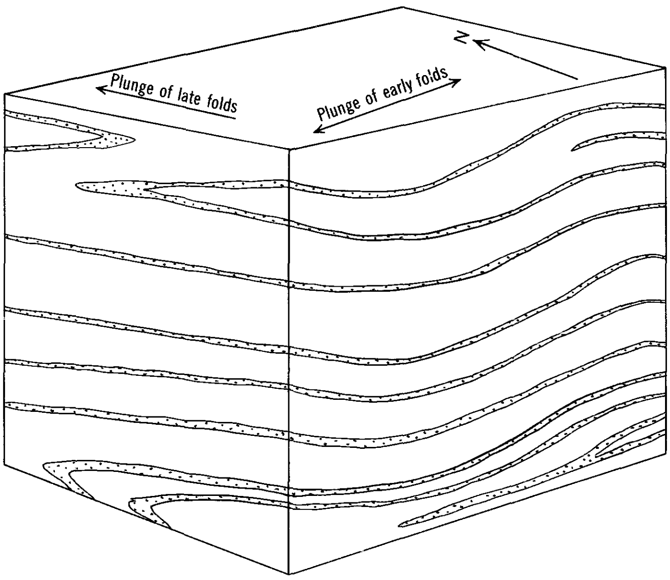


FIGURE 135.—Refolded early isoclinal folds near the Park dome. Thin beds are biotite-garnet schist that is interbedded with quartz-biotite-muscovite schist. Apparent lenses of thin beds on right side of block are noses of folds. Diagram shows generalization of exposure in railroad cut 900 feet S. 85° E. from bench mark 5,752. Height of block approximately 6 feet.

east and to the north (pl. 36, diagrams *g*, *f*, *e*) do not fall on a simple great circle, indicating that some later deformation of early minor structures has occurred.

In the area north and northwest of Atlantic Hill and around the Bear Mountain dome, the early minor folds and related structures are difficult to recognize. The axial planes of probable early isoclinal folds strike east and dip gently to moderately south on the south end of the Bear Mountain dome. On the east and southeast sides of the dome, the folds strike northeastward and dip to the southeast; they are extremely tight and seldom visible. Plunge data on all the folds and other linear elements are summarized on plate 36 in diagram *a*, which represents the area from Bear Mountain eastward to the Grand Junction fault. The submaximums indicating a due south plunge are caused by the early folds and linear elements. The submaximums to the east may be partly from the same type of fold but from the east side of the dome, where plunges are southeastward. However, most of the linear elements contributing to these submaximums are from younger folds that have very low dipping axial planes and limbs of nearly equal length, as do some younger folds on the Park dome. The a_0 maximums in diagram *a* of plate 36 are a few determinations on the plunge of boudins and elongate amygdules(?) in the amphibole schist of the Vanderlehr on opposite sides of the Bear Mountain dome. The trend of these is almost perpendicular to that of the plunges of early folds.

In the northeastern part of the quadrangle, the early minor folds are tight to isoclinal and plunge south-southwest. Their axial planes tend to strike more toward the northeast where the bedding curves to the northeast around the Harney Peak dome. The pattern of the east contact of the quartz-biotite-garnet schist indicates either intertonguing of lithologies or very tight left-hand folds, which would probably be early folds.

Within the mica schist in the northeastern part of the quadrangle, the early minor folds are isoclinal, and their axial planes generally strike north-northwest, parallel to the major trend of the structure. Diagram *b* of plate 36 gives a summary of the data on linear elements (chiefly fold plunges) in this area; part of the structure contributing to the b_0 maximum in the diagram may be of intermediate age. In general, the early folds plunge about 30° S. in the northern part of the quadrangle, but locally they plunge slightly southeastward. Figure 136 shows an early fold in this area, one that was warped at a later time.

INTERMEDIATE STRUCTURES

The intermediate minor structures include folds, foliation, and elongate minerals or mineral aggregates. Where these intermediate structures have been recognized, the plunge of the linear elements is more toward the west than that in the early structures. In general

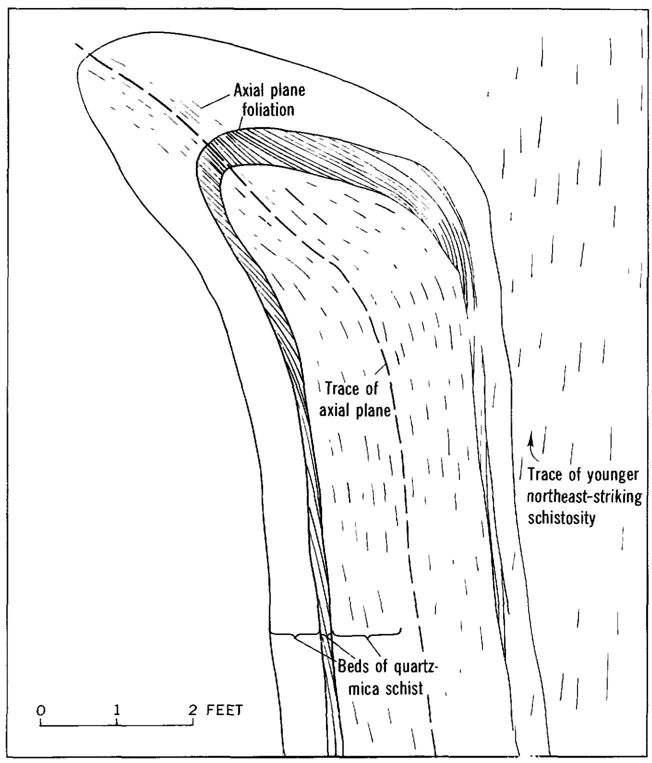


FIGURE 136.—Fold with deformed axial plane and two schistosities. One schistosity is parallel to the axial plane, and the other is a younger, northeast-striking schistosity. Fold is in the mica schist unit 1 mile south of Spring Creek.

these structures seem to have little effect on the major structural framework of the area.

The intermediate structures are associated with a foliation that strikes N. 15° E. to N. 45° E., dips steeply, and cuts across the early structures. This foliation, where identified, was mapped separately as the "younger foliation" on plate 34. It is best formed in the massive mica schist in the northeastern part of the quadrangle and in the Bugtown Formation northeast of Atlantic Hill. It can be detected in scattered areas southwest of Atlantic Hill, but it apparently dies out farther to the southwest. A similarly oriented foliation noted locally in the Fourmile quadrangle has sillimanite aggregates flattened in the plane of the schistosity. The foliation ranges from a distinct schistosity in some of the micaceous rocks to an indistinct cleavage in the more quartzose zones.

Minor folds have this crosscutting foliation as an axial-plane foliation. These folds are most abundant and are most easily recognized in the mica schist unit in the northeastern part of the area. Where the folds are assymetric, they have a left-hand pattern; but more commonly the folds are symmetrical. Many of the folds are open, but others are moderately tight to isoclinal, especially where the foliation is well formed. The plunge of the folds is moderate in a southwesterly direction.

The intersection of the crosscutting foliation and earlier folded beds, which trend predominantly northwest to north and dip moderately to the southwest and west, produces a mineral elongation and lineation in many outcrops that is parallel to the axes of the folds of intermediate age. In some outcrops the calc-silicate ellipsoids are elongate in this same direction. Some of the ellipsoids contain hornblende needles that are oriented parallel to the intermediate lineation, but the main elongation of the ellipsoid is that of the early fold axes. The more westerly plunge of the intermediate linear elements can be seen on diagrams *i*, *l*, and *k* of plate 36. Probably the intermediate linear elements are included in the b_0 maximum in diagram *h* (pl. 36) on the Park dome.

Farther north the effect of the intermediate structures is more evident. In diagrams *c*, *e*, *f*, and *g* (pl. 36), the early folding gives rise to the most prominent great circle of bedding planes whose pole is β, but the intermediate folding causes a less prominent great circle whose pole is β'. The β' pole lies west of the β pole and corresponds to the b_i maximum in these areas. Also, poles to the crosscutting foliation are shown in diagrams *c* and *g*. The b_i maximum lies on the intersection of this foliation with bedding attitudes from the early folds, mostly because the intermediate linear elements are more distinctly formed, and hence have been measured where the earlier bedding attitudes are nearly at right angles to the crosscutting foliation.

The intermediate structures cannot be readily recognized on the Park or Bear Mountain domes—either because they are lacking, or, more likely, because of a similarity in orientation to early and late structures. The crosscutting foliation dips to the southeast on the southeast side of the Bear Mountain dome and apparently flattens nearer the dome. The moderate dips of the late crosscutting foliation in figure 137, which summarizes the structural data for a small area northwest of Atlantic Hill, nearly coincide with the foliation maximum of diagram *d* (pl. 36), which is closer to the Bear Mountain dome. The crosscutting foliations shown in diagram *d* may be either intermediate or early, for in this area so many structural elements trend northeast that it is impossible to distinguish early from intermediate structures. The convergence of the attitude of the early foliation and intermediate foliation is brought out by figure 138, in which all crosscutting foliations in the Graveyard Gulch area are plotted. The steeper dipping foliations near the edge of the diagram are clearly intermediate structures, and the more gently dipping maximum is from east-trending lower dipping early foliations. The two foliations merge on the left side of

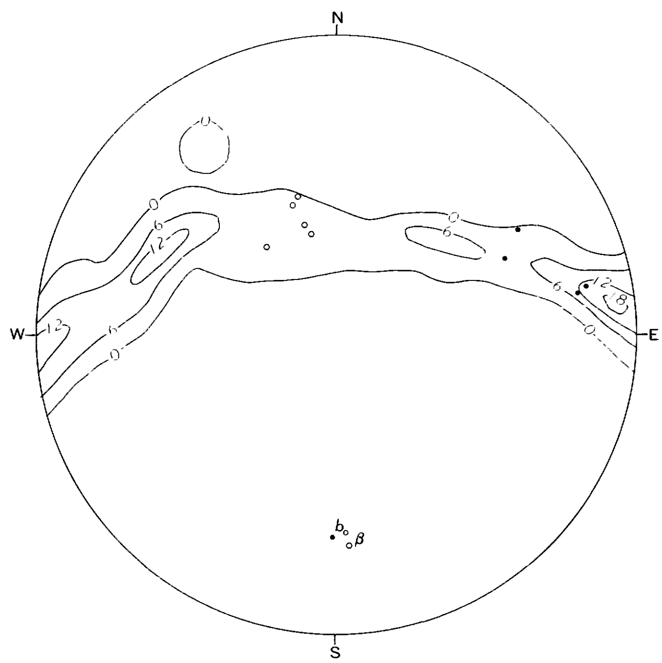


FIGURE 137.—Contours of poles to bedding and late foliation northwest of Atlantic Hill in NE¼ sec. 6. Solid-line contours are on 32 poles to bedding attitudes; circles are poles to late foliation. β is pole to great circle through bedding attitudes. Dots are poles to early foliation; b_0 is earliest linear maximum from diagram *i*, plate 36. Lower hemisphere plot, contoured in percent of poles per 1 percent area.

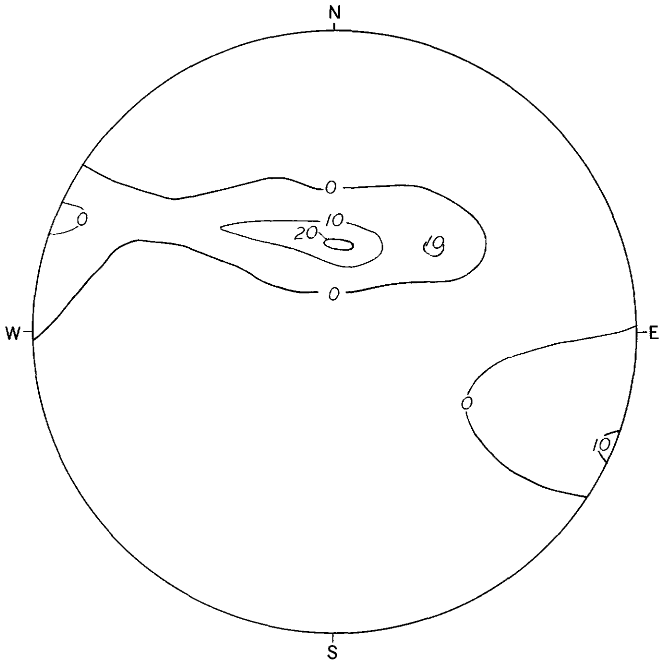


FIGURE 138.—Contours of poles to crosscutting foliation in the Graveyard Gulch area. Compare with diagram *e* on plate 36.

figure 138; in areas farther to the northwest the two types may coincide.

Near Graveyard Gulch and the Grand Junction mine, the existence of northeast-trending cross folds is indicated by the distribution of lithologic units and by the bedding attitudes. These are apparently cut off by the Grand Junction fault. They are probably of intermediate age.

The gentle cross folds or warps in the Atlantic Hill area may also be part of the intermediate structures. For reasons discussed in the following section, however, these seem best regarded as late structures.

LATE STRUCTURES

The latest known structures include minor folds, cleavage, crenulations, and elongate mineral aggregates. These structures are best formed in the Park Dome area, and many of them have been briefly described in the section on the Park dome. Many of the late folds deformed earlier folds (fig. 135) and thus tended to obliterate earlier lineations. These folds range in size from microscopic crinkles to gentle folds several hundred feet across (pl. 35). Some of the smaller folds are isoclinal, with axial planes dipping to the east, and commonly have a slip cleavage (shown by a separate symbol on pl. 34 and in fig. 125) parallel to the axial plane. Metamorphic minerals, such as garnet and biotite, are locally broken, deformed, and altered along this slip cleavage, and commonly a mineral lineation is parallel to the latest fold axes.

The late folds tend to decrease in amplitude and abundance away from the center of the Park dome and are virtually absent at the contact between quartz-biotite-garnet schist and mica schist (pl. 36). Near the crest of the dome and in the saddle to the east, the axes of the late folds are horizontal, or nearly so, and trend about N. 25° W. to S. 25° E. The plunges of both the late fold axes and the mineral lineations increase northwestward and southeastward, so that the average plunge on the north side is about 15° N. 22° W. and that on the south side is 20° S. 27° E., as shown by the b_1 maximums in diagram *h* of plate 36. The north-plunging folds have a right-hand map pattern, and the south-plunging folds, a left-hand pattern.

In the railroad cuts northeast of the Park School the folds are in a low-dipping trough that separates the dome from steeper dipping rocks to the east. Here, parts of the quartz-biotite-garnet schist are repeated several times by early isoclinal folds, which are themselves deformed by gentle symmetrical late folds (fig. 135). Minor crenulations are common in the thinly bedded low-dipping quartz-biotite-garnet schist, but in areas of steeper dip farther from the center of the dome these crenulations disappear.

Near the Crown mine, southeast of the crest of the Park dome, are many small very tight isoclinal folds whose axial planes are almost flat or dip gently to the south or southeast. These folds generally do not have conspicuous long limbs, and they appear to be "piled up," as if they were part of the nose of a larger fold. The plunge directions of these folds correspond to those of other late folds on the southern part of the dome, and they probably are contemporaneous.

The Bear Mountain dome does not have clearcut late folds like those of the Park dome, but it does have other evidence of late structure. The area represented by diagram *d* of plate 36 has very tight folds whose axial planes strike northeast and dip gently to moderately to the southeast. The general trend of map units at this locality is to the north-northeast, yet diagram *d* indicates that the strike of bedding is about N. 60° E., which no doubt reflects the influence of the small northeast-trending folds. Deformation of minerals in both the metamorphic rocks and the pegmatites suggests late deformation.

The change of strike and the flattening of dip of the Grand Junction fault across Tenderfoot Creek can also be viewed as evidence of late deformation. The dip of the fault is only about 28° at Tenderfoot Creek, but it steepens considerably in a distance of about 0.5 mile in each direction along strike. The bend in trace of the fault is identical with that around the Park dome, where the dip of the fault is also flatter on the southwest side of the dome. Possibly, a concealed dome may be present near Tenderfoot Creek. The open cross fold north of Atlantic Hill, which is an anomalous structure at best, may be associated with a dome, and the flattening in attitude of the intermediate foliation in this area is consistant with this same possibility.

In several outcrops along Bugtown Gulch and in exposures of the Bugtown Formation along French Creek, the axial-plane foliation of the earliest folds is itself folded and warped into foliation "folds." The wave length of the "folds" is directly related to the thickness of the beds, as shown in figure 139. These folds are the prominent structural feature of a few cliff exposures. The folds were generally recognized only in areas of steep dips on the overturned limbs of folds, where they formed as a result of late shearing and slippage between thick competent beds. Like the youngest structures of the Park dome area, they have a "west over east" movement pattern.

BOUDINAGE

Boudinage is most common in the amphibole schists of the Vanderlehr Formation. Its existence in these schists seems extraordinary, for the rock seems to be uniform in both composition and competency. Yet small boudins—of the type sketched in figure 140, are common on the flanks of the Bear Mountain dome. Slight differences in grain size probably control the relative competency of different layers. Thus, the thin bed of fine-

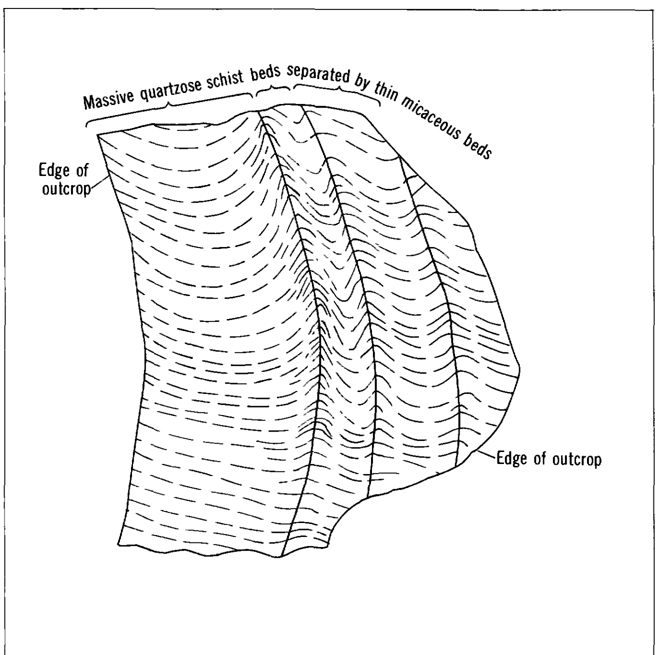


FIGURE 139.—Foliation folds. The amplitude of these folds is a function of both bedding thickness and competency. Bugtown Formation, 2,000 feet northeast of Lucky Bird mine.

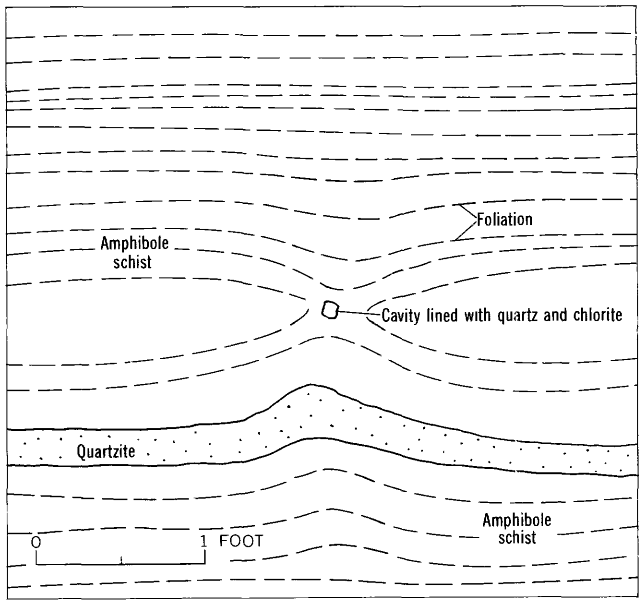


FIGURE 140.—Boudinage in the lower amphibole schist unit of the Vanderlehr Formation. Fine-grained quartzite is less competent than the coarser grained amphibole schist. Small differences in amphibole content cause the main foliation in the amphibole schist.

grained quartzite shown in figure 140 is less competent than the coarser grained amphibole schist of the boudin.

The boudins are elongate in the same direction as the amygdulelike aggregates and are at nearly right angles to the long dimension of the Bear Mountain dome. The boudins plunge in opposite directions on each side of the dome, thus producing the A_0 maximums in diagram *a*, plate 36. That the data yield two maximums, rather than a great or small circle, is somewhat surprising. The reason probably is that most of the measurements are of rocks that have uniform strike and dip on the east and west flanks of the dome. No attitudes could be measured on the south side of the dome.

The orientation of the boudins suggests that tension was parallel to the dome axis, but the elongation of the supposed amygdules cannot be so readily explained. The amygdules may represent an *a* lineation, or they may not be amygdules at all, but tiny tension cracks in the amphibole schist along which adjacent minerals have recrystallized. The latter may be the better explanation because in some places the foliation bends somewhat inward at the cavity or at the edge of the mineral aggregate, thus resulting in structures that appear to be transitional to miniature boudins.

The age of the boudinage is not clear. The boudinage orientation is virtually perpendicular to the long axis of the Bear Mountain dome and suggests that the boudins formed at the same time as the dome. Lower rank metamorphic minerals at the ends of some boudins also suggest a late origin. However, the boudins are also virtually perpendicular to the axes of the earliest folds, and may in part be old structures.

A noteworthy boudinage is associated with the pegmatites in which a thin pegmatite serves as the more competent layer and is separated into two boudins by septa of the schist country rock, as illustrated in figure 14. In thicker pegmatites, the schist does not penetrate as far proportionately, and a quartz fracture-filling occupies most of the space between the boudins. The long axes of the boudins are generally oriented approximately normal to the plunge of the pegmatite.

FAULTS

Six faults have been mapped in the Berne quadrangle: two that have been observed in outcrops, and four that are inferred from geologic relations. Other faults may possible be present but unexposed. The largest of the exposed faults is the Grand Junction fault, which is so large that stratigraphic correlations between the rocks on the two sides cannot be made. The other exposed fault, east of the Park dome, has been observed in only one outcrop, and its displacement is believed to be small. The four inferred faults are (1) the Vestal Springs fault, (2) the Atlantic Hill fault, (3) a smaller fault which joins the Atlantic Hill fault, and (4) a small fault along the east side of the quartzite unit in the northeastern part of the quardangle.

In addition to the faults shown on plate 34, breccia zones, commonly parallel to bedding, occur at several localities. One such zone is represented by scattered blocks of breccia in a field along the north side of Middle Fork Creek about 0.3 mile west of the Saginaw mine (pl. 34, B–5). The breccia blocks contain angular to subangular fragments of schist, up to several inches long, cemented by limonite and albite. These breccias probably resulted from small faults, but no significant displacement is apparent.

The Grand Junction fault is exposed on the west side of Tenderfoot Creek and in a prospect pit about 500 feet to the north. The inclined shaft in the Old Bill mine (pl. 34, F–4), is apparently along the fault plane. The fault is marked by a broken and somewhat altered zone about 1 foot thick, and graphite is abundant locally in the rocks for several tens of feet on either side. Quartz veins are common on the hanging-wall side of the fault near Grand Junction mine; the main quartz vein at the Grand Junction mine is about 100 feet thick. The steep dip of the vein suggests that it probably intersects the fault plane at a shallow depth. The Grand Junction fault is generally parallel to the structures in the country rocks; the greatest discordance occurs near Tenderfoot Creek, where the fault parallels the beds east of the fault but is at a fairly wide angle to the beds west of the fault. Along the Park dome the beds of the Bugtown Formation are at a 10° angle or less to the trace of the fault.

The curved trace and the variable dip of the Grand Junction fault indicate that the fault plane was deformed, presumbly at the time the Parl- dome was formed. The fault dips 45° W. at the old Bill mine and flattens southward where its trace curves around the Park dome. At Tenderfoot Creek the fault trace forms a curved pattern that is a small replica of the curve around the Park dome. Furthermore, the dip is only 28° W. on the west side of this curve, which corresponds to the low dip on the southwest side of the Park dome. North of Spring Creek and east of Bear Mountain the trace of the fault across the steep topography indicates a steep dip. A possible restoration of the trace of the fault plane prior to doming is shown in figure 133*B*.

The forming of this fault was one of the oldest structural events in the quadrangle. It preceded intrusion of the gold-bearing quartz veins, for they are localized by the fault. Because the quartz veins are older than the pegmatites, the fault is clearly older than the pegmatites. The fault transects the northeast-trending cross folds in the Graveyard Gulch area, thus indicating fault movement prior to the deformation of inter-nediate age. The main movement may have occurred much earlier, for linear features on both sides of the fault are concordant, suggesting that the fault formed during the earliest folding and deformation.

A reverse fault that dips 45° W. is exposed in a railroad cut east of the Park School. The rocks in the fault zone are much broken, and the wallrocks are drag folded. The fault occurred after metamorphism and is contemporaneous with, or slightly later than, the youngest folding. Several hundred feet of apparent dip-slip displacement is inferred in section *A–A′* (pl. 35). A 10-foot-wide gouge zone in a long adit about 1 mile north of Berne siding (pl. 34, F–4) is probably on this same fault and is so shown on the geologic map.

The Atlantic Hill fault, the Vestal Springs fault, and the minor branch fault extending from the Atlantic Hill fault are inferred from the lack of continuity in the distribution of units in the lower part of the Bugtown Formation. The Atlantic Hill fault is inferred to have displaced much of the upright limb and, locally, part of the overturned limb of the Atlantic Hill anticline a considerable distance to the southeast. The displacement possibly occurred largely along a plane approximately parallel to the axial plane of the fold and the axial-plane schistosity. To the northwest, near the Bear Mountain dome, the movement of the main fault and of the associated branch fault may have been largely along the axial plane. The lateral displacement, combined with an unknown amount of vertical displacement, tended to conceal the northwest extension of the axial plane of the Atlantic Hill anticline. The Vestal Springs fault explains the offset of approximately 1 mile in the quartz-mica-feldspar schist unit northwest of Vestal Springs. The inferred displacement is the same as that on the Atlantic Hill fault but is confined largely to the axial-plane part of a smaller fold on the overturned limb of the major fold. The inferred restoration of the units in plan view (ignoring vertical movement) is shown in figure 133. Each major fault is shown as a moderately low angle reverse fault in cross sections *B–B′* and *C–C′*, plate 34.

The Vestal Springs fault and its companion fault to the south are inferred from the lack of continuity in the distribution of units in the lower part of the Bugtown Formation. The lower quartz-mica-feldspar schist unit northwest of Vestal Springs seems to be horizontally offset by the Vestal Springs fault for approximately 1 mile. The same schist unit is similarly offset, though not as far, by the companion fault to the south. The inferred restoration of the schist in plan view (ignoring vertical movement) is shown on plate 36, diagram *b*. In most places the trace of both faults is apparently subparallel to the structure of the Bugtown Formation. Near the trace of the Vestal Springs fault, as shown on plate 34, a few schist outcrops contain bent minor-fold axes and crumpled foliation planes; however, no other direct evidence of the location of the fault plane was observed. Bedding strikes into the Atlantic Hill fault at a high angle just northwest of Atlantic Hill, but this is the only locality where this much discordance was noted.

The map pattern shows that these faults formed before the domes. The general similarity of these faults to the Grand Junction fault suggests that they are of the same approximate age as the Grand Junction fault. Their occurrence in the overturned limb of the Atlantic Hill anticline suggests that they formed concurrent with the early folding.

The inferred fault along the east side of the quartzite unit north of Spring Creek is based on the distribution of minor subunits in the mica schist unit. The available data are consistent with the interpretation that the east side of the fault has been downthrown generally parallel to the axial plane of the inferred fold nose, as shown in cross section *A–A′*, plate 34.

EVOLUTION AND SUMMARY OF STRUCTURES

The descriptions of the different geologic structures indicate that the three periods of deformation were (1) a major folding, (2) a shear deformation, and (3) a late doming. The time separating these events cannot be determined, but was not necessarily great. The chronologic sequence is summarized as follows:

1. Formation of the early major structures (including the Western Star syncline, Atlantic Hill anticline, and other large isoclinal folds) and the associated early minor structures (such as minor folds, axial-plane foliation, elongate calc-silicate ellipsoids, and possibly, boudinage structures). The faults in the Vestal Springs and Atlantic Hill areas were formed subparallel to axial planes probably during this period of folding, and the Grand Junction fault may have formed at the same time. The early major folds west of the Grand Junction fault presumably had nearly straight axial planes that trended north-northwest. East of the Grand Junction fault, the early major structure is either a north-northwest-trending fold that contains the quartzite north of Spring Creek in its core, or the entire sequence of lithologic units is in the limb of a very large fold that also has a general northerly trend. Regional metamorphism affected the rocks during at least the last part of this period of folding and deformation.
2. Formation of a northeast-trending nearly vertical foliation, probably caused largely by shearing forces. Chevronlike minor folds and minerals lineations associated with the foliation produced the b_i maximums (pl. 36), which plunge more to the west than do the earlier b_0 linear elements. The deformation was not extensive and decreased in intensity to the southwest. Northeast-trending cross folds in the Graveyard Gulch area may have formed at this time, but the folds of this age did not otherwise greatly deform earlier structures. Part of the movement along the Grand Junction fault occurred after this shear deformation, although the main displacement was probably earlier. The shear deformation may have been an early phase in the formation of the Harney Peak dome, for the shear foliation and related minor folds tend to die out to the southwest away from this dome. Also, in the northeastern part of the quadrangle, the foliation has the same strike as northeast-trending cross folds in the Hill City area (fig. 122), which seem to be molded on the northwest flank of the Harney Peak dome. The locally distinct crosscutting relationship of the foliation and the lack of cataclastically deformed metamorphic minerals along the foliation indicate that temperatures and pressures were moderately high.
3. Probably chiefly a doming phenomena related to the emplacement of the granite and pegmatite masses. This late deformation may not have differed

greatly in age from the intermediate deformation. The Park, Bear Mountain, and Harney Peak domes formed at this time. Preexisting structures influenced their localization in ways illustrated by the Park dome, where south-plunging folds around the dome are much more conspicuous than north-plunging ones, and the dome is clearly elongate to the north-northwest. The correlation between the domal structure and the earlier fold axes can be seen in diagram *K* of plate 36, where the bedding attitudes in the late-formed trough on the southeast side of the dome fall on a great circle whose pole is virtually the same as that of the early b_0 linear elements.

In the southern and southeastern parts of the Bear Mountain dome, the doming has warped the earlier major and minor structures around south-plunging axes, which also coincide with those of earlier structures. The earlier folds have been distended, and the central part of the dome has an apparently simple structure much like that in the central part of the Park dome. However, this simplicity is only apparent and probably results from much plastic flowage of the nearby rocks.

METAMORPHISM

The Precambrian rocks of the Berne quadrangle have undergone medium- to high-grade metamorphism ranging from that of the staurolite zone to that of the sillimanite zone. This relatively high grade metamorphism is characteristic of the entire region surrounding the Harney Peak Granite and contrasts with the much lower grade metamorphism of the schists to the north. The largely thermal metamorphism is associated with the intrusion of large masses of pegmatitic granite in the Harney Peak dome, as well as the many satellitic pegmatites and domes nearby, and it affected the rocks so greatly that earlier metamorphic events cannot be clearly distinguished. Nevertheless, such effects as the formation of foliation involving micas and other minerals in early structures leaves little doubt that there were at least two ages of metamorphism, or two peaks in a single age of metamorphism. Some rocks show evidence of retrograde metamorphic reactions, which were possibly associated with one or both of the major events. Certain of the schists also show evidence of metasomatism, which was apparently related primarily to the intrusive periods.

SILLIMANITE ISOGRAD

The sillimanite isograd shown on plate 34 trends northeast across the Berne quadrangle and, in general, does not follow geologic structure. Sillimanite also occurs near the center of the Bear Mountain dome but was noted in too few places to permit drawing an isograd. Southeast of the isograd shown on plate 34, the more aluminous schists in the Mayo and Bugtown Formations contain sillimanite, as do similar beds in the quartzite and quartz-mica schist unit. Sillimanite also occurs in part of the mica schist unit, in the Crow Formation, and in a few places in the quartz-biotite-garnet schist. Northwest of the isograd, the conspicuous metamorphic minerals are staurolite, garnet, and andalusite.

Near the isograd, sillimanite occurs only in the highly aluminous rocks, and its distribution is very spotty even in them. This effect of composition on the first growths of sillimanite is very marked. The westward bulge of the isograd in the Atlantic Hill area may be due entirely to the greater abundance of alumina-rich schists in the middle part of the Bugtown Formation. Sillimanite was noted as an accessory mineral in two thin sections of very aluminous rocks from the Crow Formation at least 0.6 mile west of the isograd, as shown on plate 34. Nevertheless, the high aluminum content did not necessarily lead to an early formation of sillimanite. The mica schist and quartz-biotite-garnet schist units, which are both much richer in aluminum than most units of the Mayo and Bugtown Formations, are also rich in iron and magnesium; the aluminum was taken up in stable minerals such as garnet and staurolite, so that the formation of sillimanite was delayed. In the richest aluminum-bearing schists of the mica schist unit, andalusite occurs rather than sillimanite. The andalusite is stable once it has formed far into the sillimanite zone, and thus sillimanite cannot form in the most aluminous rocks. In the less aluminous (or more alkalic) rocks, destruction of some minerals, such as chlorite, and probable changes in composition of others, such as the micas, permits the formation of sillimanite. Each composition obviously has its own isograd for the first growth of sillimanite; the one shown on plate 34 is for aluminous quartz-mica schists characteristic of the Bugtown and Mayo Formations. The variability in the rate of growth of sillimanite is substantiated by the experimental findings of Winkler (1957) regarding sillimanite in carbonate-free clays of various composition, and by the observations of Harker (1939) and of many other geologists.

Sillimanite is more widely distributed in quartz veins than in country rock. It occurs in some of the quartz veins in the mica schist unit and in the Loues Formation throughout the northern part of the quadrangle, far beyond the area bordered by the isograd. Sillimanite also occurs rarely in some veins in aluminous beds in the Bugtown and Mayo Formations at a considerable distance west of the isograd. The occurrence of sillimanite in veins beyond the sillimanite isograd was also noted by A. E. J. Engle (oral commun., 1957) in the Adirondack Mountains of New York. Probably the vein processes associated with the main metamorphism, and the presence of a fluid phase in the vein, permitted the sillimanite to nucleate and grow in areas where the nucleation rate or growth rate was inadequate in the dryer country rock.

Andalusite is also present in many quartz veins on both sides of the sillimanite isograd. In many veins it coexists with sillimanite, apparently in equilibrium, but in a few veins the textures indicate that andalusite is replaced by sillimanite.

MINERALOGIC AND TEXTURAL CHANGES

Except for the presence or absence of sillimanite, there are no marked mineralogic and textural changes in the rocks immediately above and below the sillimanite isograd. The grain size of the matrix of some of the schists does increase slightly from the northern, central, and western parts of the area toward the southern and eastern parts, and the porphyroblasts also increase in size. The differences in textures among rocks having different compositions in a single metamorphic zone are much greater than those among rocks of the same kind above and below the sillimanite isograd.

Staurolite occurs throughout the quadrangle in beds rich in iron and aluminum. Garnet, biotite, and muscovite are universally associated with the staurolite, and andalusite or sillimanite may also be present. No difference was noted in the appearance of unaltered staurolite above and below the sillimanite isograd, but above the isograd more of the staurolite is replaced by muscovite and biotite aggregates (fig. 125*E*, *F*). Staurolite in areas west and north of Atlantic Hill tends to be finer grained than that in rocks to the southeast, but the differences are not great. North of the quadrangle, however, the staurolite decreases in size, and farther north it is absent.

Garnet textures and grain sizes vary considerably throughout the quadrangle but show no distinct relation to the sillimanite isograd. The garnet ranges from irregular skeletal grains—some of which have been rotated, as shown in figure 124—to euhedral crystals containing few inclusions. Some of the garnet grains in the mica schist and quartz-biotite-garnet schist units appear to be greatly elongated, as shown in figure 141. This elongation was not necessarily caused by deformation of previously equant garnet crystals; for in rocks of lower metamorphic grade in the same units near Hill City, similar elongate garnets are clearly formed by selective replacement of thin beds (fig. 142) and ap-

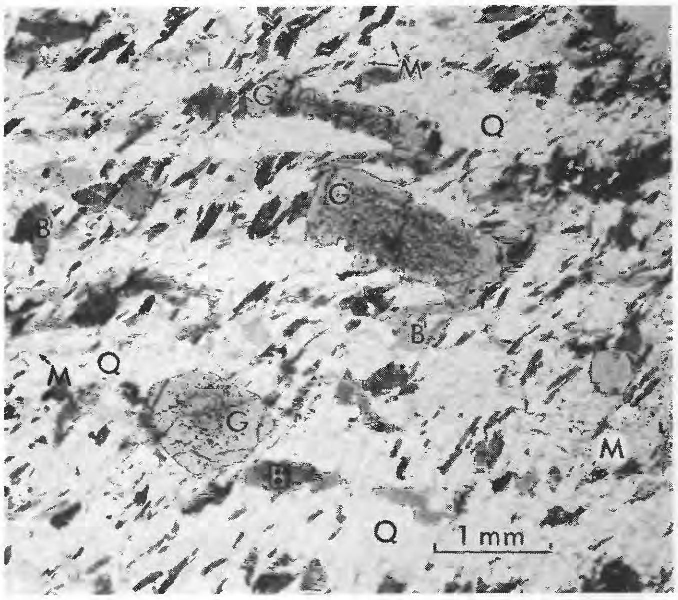


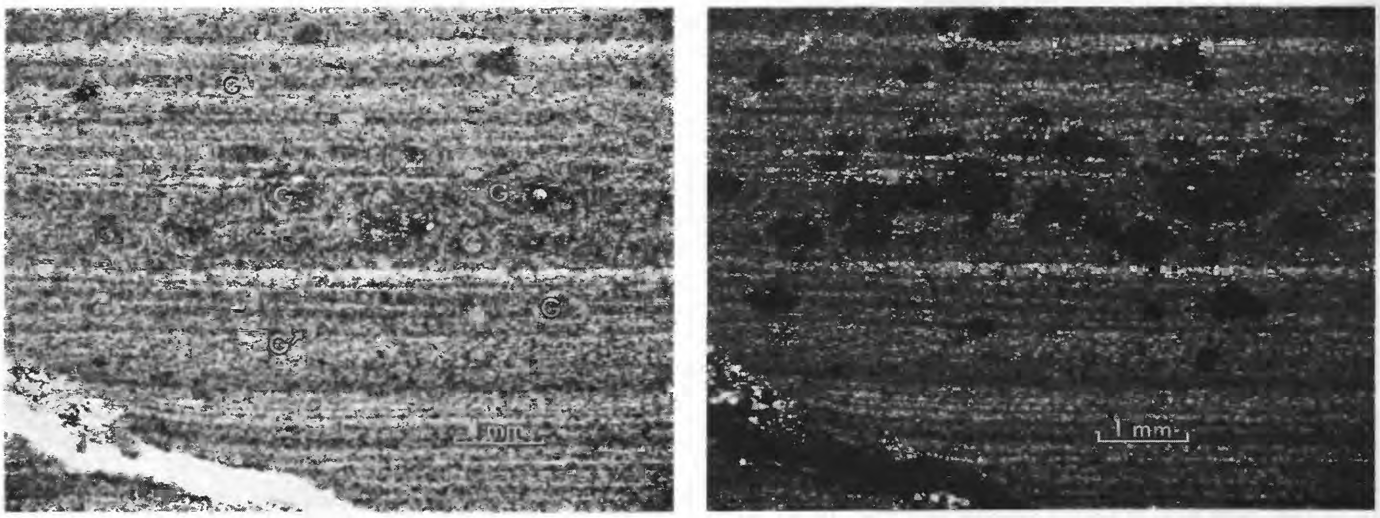
FIGURE 141.—Elongate garnets and discordant foliation in the mica schist unit. Horizontal banding is probably the original bedding. Foliation of biotite (B) cuts bedding. Long garnet (G) crystals probably were parallel to bedding but have been rotated somewhat. Matrix is quartz (Q) and muscovite (M).

parently consist of a group of separate twinned(?) crystals.

Kyanite is associated with staurolite and garnet in one outcrop of the Mayo Formation below the sillimanite isograd, and is also present with sillimanite in the quartz-biotite-garnet schist unit about 0.5 mile northwest of the Newark mine (pl. 34, D–9). Kyanite is not known to be abundant in any of the metamorphic rocks of the southern Black Hills.

Andalusite is present in most of the mica schist unit—both above and below the sillimanite isograd—and in at least two subunits of the quartzite and quartz-mica schist unit along the east side of the quadrangle. A few aluminum-rich schists in the Bugtown Formation contain a small amount of andalusite, but none was found in the Mayo Formation. Andalusite would probably also occur in any other part of the quadrangle if the rocks had had an appropriate initial composition. The andalusite clearly is restricted to the muscovite-rich rocks of the mica schist unit; it characteristically occurs as large knots or flattened lenses consisting of aggregates of smaller poikiloblastic crystals, many of which are replaced by muscovite and biotite. Fresh, unreplaced andalusite is less common (except in veins) in the eastern part of the quadrangle—above the sillimanite isograd—than elsewhere.

Sillimanite characteristically forms aggregates of tiny needles (fig. 125*E*). The aggregates are commonly elongated parallel to one another and produce a promi-



A *B*

FIGURE 142.—Elongate garnets in the mica schist unit. Garnets (G) have grown in the darker beds richer in biotite. Matrix is fine-grained biotite, sericite, quartz, and chlorite. Sample from north of Hill City. *A*, Plane-polarized light. *B*, Crossed nicols. (White area in *A* is a hole in section.)

nent lineation, unlike the unoriented crystals of staurolite and andalusite. This lineation is clearly related to differences in grain size rather than to differences in intensity of deformation. In the eastern part of the quadrangle near the Harney Peak dome, the sillimanite occurs in flat disks or lenses smaller than, but similar in shape to, the andalusite aggregates. Sillimanite is seldom replaced by mica, whereas andalusite and staurolite are commonly replaced (fig. 125*E*).

In most thin sections the sillimanite is intergrown with, and seems to have replaced, biotite. In the eastern and southwestern parts of the quadrangle the muscovite content is notably low in the sillimanite-rich schists, and very probably some of the sillimanite formed from muscovite.

Cordierite is abundant in a few beds in the mica schist and the quartz-biotite-garnet schist units, in the Vanderlehr and Crow Formations and in the upper part of the Loues Formation, both above and below the sillimanite isograd. In the quartz-biotite-garnet schist unit and in the Vanderlehr and Crow Formations, the cordierite occurs only as small grains, which were noted in thin sections. The cordierite is sufficiently abundant in some beds in the Crow Formation, however, to give a slightly purple cast to the rock. This fine-grained cordierite is accompanied by microcline (possibly replacing the cordierite), biotite, muscovite, and plagioclase. In the mica schist unit and the Loues Formation, the cordierite forms large black to dark-blue porphyroblasts that are generally rodlike and are as much as 2 inches long. Many of these exhibit various stages of replacement by micas and chlorite, but unaltered cordierite has been observed from the north edge of the quadrangle to as far south as the Old Bill mine, which is stratigraphically well above the sillimanite isograd. Biotite, muscovite, plagioclase, quartz, garnet, chlorite, and magnetite or ilmenomagnetite are generally present in the groundmass of the cordierite-rich rocks, and staurolite and andalusite may form porphyroblasts in the same rock. Locally the cordierite crystals, like the andalusite crystals, are somewhat deformed—apparently by the northeast-trending younger axial-plane foliation. The cordierite in the quartz-biotite-garnet schist unit is present both above and below the sillimanite isograd; that occurring in areas of younger folds, as near the Park School, is deformed and is in part replaced by microcline.

In the meta-iron-formation, grunerite appears to be stable under the existing metamorphic conditions throughout the quadrangle. A small amount of green hornblende occurs where aluminum and the alkalies are too abundant to be contained by the grunerite. Some of the associated garnet is filled with needles of grunerite, as if the garnet were being replaced (fig. 123 *C*, *D*). In several thin sections showing this replacement, the garnet is rimmed by green hornblende, which takes up the excess aluminum from the replaced garnet. The amphiboles and garnet in the meta-iron-formation are believed to have formed by reaction of iron-rich carbonate beds and quartz-rich beds. Iron-rich carbonate minerals are now sparse in these rocks, and were found only below the sillimanite isograd.

Pyroxene is limited to the meta-iron-formation and the calc-silicate rocks throughout the quadrangle. In the meta-iron formation the pyroxene is largely diopside or augite. It is very coarse grained; crystals an inch or more long across the rock fabric as if the pyroxene formed late. Diopside is common in some of the calc-silicate ellipsoids, and bronzite was noted in a few ellipsoids in rocks above the sillimanite isograd.

In the aluminous mica schists in the mica schist unit, the relationships of magnetite and magnetite-ilmenite intergrowths are sufficiently complex to warrant additional study in areas of different metamorphic intensity. The small plates containing 10–20 percent of ilmenite laths in magnetite were initially believed to have formed from original solid solutions of ilmenite in magnetite. The much larger magnetite crystals that coexist with the small mixed crystals contain only 0.025–0.050 percent titanium (analyzed by J. C. Hamilton, U.S. Geological Survey), or much less titanium than in the mixtures. (The 10–20 percent is an estimate made during petrographic examination, for a pure concentrate of this fine-grained material could not be obtained for analysis.) The initial assumption then led to the conclusion that the differences in titanium content could only have resulted from two distinct growth periods under different physical or chemical conditions. However, more recent work by Lindsley (1962, p. 100–106) indicated that ilmenite-rich magnetite mixtures probably result from oxidation of magnetite-ulvöspinel solid solutions. A complete solid solution of magnetite ulvöspinel exists above about 600°C (Vincent and others, 1957), but at lower temperatures there should be a relatively pure magnetite and a solid solution of magnetite-ulvöspinel in equilibrium. Oxidation of the latter mineral would lead to the magnetite-ilmenite integrowths. If the solvus is relatively flat on the ulvöspinel side, as in the published tentative curve (Vincent and others, 1957, p. 636), the composition of the intergrowths can eventually be used as a geothermometer for metamorphic rocks.

RETROGRADE METAMORPHISM

The main retrograde effect noted in the Berne quadrangle is the common replacement of cordierite by aggregates of chlorite, muscovite, and biotite, and of staurolite and andalusite by muscovite and biotite. In partial replacements of staurolite, biotite forms a distinct halo around a core of fresh staurolite (fig. 125*F*); rarely, a few blades of chlorite occur with the biotite. In complete pseudomorphs the biotite rim is absent, and the pseudomorph is wholly muscovite. Potassium and water have migrated inward, and iron outward, but the source and the distance traveled are not known.

In areas of late deformation, such as in the Park and Bear Mountain domes, garnets have been replaced by aggregates of sericite and chlorite (fig. 125*B*.) The rock is generally sheared and altered, and the replacement is not always a typical retrograde effect; the replacement may be the result of hydrothermal alteration or of alteration associated with local metasomatism.

A more puzzling phenomenon, which may not be a retrograde effect at all, is the spotted character of some schists in the Bugtown Formation. The spots have dark halos, rich in biotite, and light-colored interiors, rich in quartz (fig. 123*E*, *F*). Similar spots occur in lower grade rocks near Hill City and Keystone. Widespread distribution of these spots suggests that they are altered relicts of any eariy metamorphic mineral, but there are no clues in the chemistry of the replacement(?) or in lower grade rocks of this area as to the identity of this mineral.

METASOMATISM

Metasomatism involving boron, potassium, iron, and water has had a small but significant effect on some of the rocks of the Berne quadrangle. Probably the best examples are the altered rocks of the Bugtown Formation in the area near pegmatite 72 (D–7). These rocks now, instead of being normal Bugtown quartz-mica-feldspar schist contain quartz and about 10 percent microcline and 20 percent tourmaline. The rock has lost its typical schistosity and resembles a sugary-textured granular rock or greisen. The altered rock may be several tens of feet thick and is generally cut by a few pegmatite stringers, which are clearly associated with the alteration. Much of the potassium in the rock probably came from the micas originally present in the schist, for mica is replaced by the tourmaline and microcline (fig. 125*A*). Boron, necessary for the formation of the tourmaline, has been added.

In the southeastern part of the quadrangle, especially near pegmatite 52 (G–9), many outcrops of quartz-mica schist contain large microcline porphyroblasts, some more than 1 inch in diameter. The porphyroblasts occur only where pegmatite are abundant. The porphyroblasts are very poikilitic, containing about 30 percent of the other minerals of the schist. Comparison of analyses of similar rock in the Fourmile quadrangle with an analysis of unaltered schist a few feet away shows that potassium has been added and iron removed during alteration (Redden, 1963, p. 262).

Large lathlike microcline crystals in the quartz-biotite-garnet schists on the Park dome also indicate potassium metasomatism. The microcline crystals enclose small chevron folds that are related to a younger slip cleavage (fig. 125*D*); thus, the microcline is younger than the slip cleavage. These relationships suggest that some potassium metasomatism occurred after the formation of the dome. Many of the rocks in the Bear Mountain dome are also rich in microcline, and some of the potassium may be of the metasomatic origin.

Local metasomatic effects can be observed around many of the pegmatites. Tourmaline, microcline, apatite, and albite have been added to a relatively thin zone near the pegmatite contacts. The altered rock is thickest along discordant contacts, apparently where the country rocks were most permeable.

The pegmatites and fluids associated with them are the most obvious cause of metasomatic effects of late age involving potassium, boron, and water. Nevertheless, metamorphic processes may also have been major factors. Little or no microcline is present in the higher grade schists, where, apparently, muscovite and some biotite are replaced by sillimanite. The generally supposed reaction for the replacement of muscovite is:

$$\text{Muscovite} + \text{quartz} \rightarrow \text{sillimanite} + \text{microcline} + \text{water}.$$

But if no microcline is present at the site of the reaction, then potassium as well as water are likely to migrate—presumably upward to areas of lower metamorphic intensity, where they can cause metasomatic reactions and retrograde metamorphic effects. The fact that potassium and water are the most abundant metasomatic materials in the area adds to the plausibility of this supposition.

ECONOMIC GEOLOGY

Pegmatite mines in the Berne quadrangle have produced sheet and scrap mica, feldspar, beryl, spodumene, amblygonite, and columbite-tantalite. There has also been minor gold production from several small mines and extensive prospecting for tin. The Crown mine has produced mica valued at more than $200,000 and possibly as much as $400,000 (Page and others, 1953, p. 95–96). This value far exceeds that of all other pegmatite minerals obtained in the Berne quadrangle.

The mineral deposits and geology of nearly all the larger pegmatite mines in the quadrangle have been described since World War II. The Crown (F–7), Highlands Lode (D–9), and Hunter-Louise (G–2), were described by Page and others (1953), and the Big Spar No. 1 (F–8), Dorothy (G–9), and pegmatites 58–62 and 65 were described by Lang and Redden (1953). An excellent detailed study of the High Climb (G–2) and its mineral deposits was made by Sheridan (1955).

PEGMATITE DEPOSITS

The zoned pegmatites contain the only deposits that are minable under current conditions. However, most of the zoned pegmatites shown on the geologic map (pl. 34) are too small to be profitably mined. The many homogeneous and layered pegmatites contain large reserves of feldspar and scrap mica, but the grade of the minerals is too low for mining.

Sheet mica and beryl are produced largely from the outer zones of the zoned pegmatites, and feldspar and lithium minerals from the inner zones. Columbite-tantalite is also recovered from inner zones, but very little has been produced. Table 12 summarizes data about the more valuable mines and indicates the zonal position of the different economic minerals.

MICA

Sheet mica from the Crown mine (pl. 34, F–7) has been the most valuable mineral product from the Berne quadrangle, though deposits of scrap and sheet mica also occur in several other pegmatites. The Crown mine produced a recorded 70,000 pounds of sheet mica and probably an equal quantity that is unrecorded, plus 390,000 pounds of punch mica (Page and others, 1953, p. 96). The value of the known mica production exceeds $200,000, indicating that this is one of the largest sheet

TABLE 12.—*Summary of mineral production from the major pegmatite mines in the Berne quadrangle*

[Zone: C, core; IZ, intermediate; WZ, wall. Production: O, minor; X, major]

Mine (pl. 34)	Minerals and zones from which mined					Comments on size and ore grade of deposits
	Potassic feldspar	Sheet mica	Scrap mica	Beryl	Spodumene	
Big Spar No. 1 (F-8)	X(IZ)		O(WZ)	O(WZ)		Size diminishes rapidly at depth.
Crown (F-7)		X(WZ)	O(WZ)	O(WZ)		Mostly mined.
High Climb (G-2)	O(2d IZ)		X(1st IZ)	X(1st IZ)		In production in 1958. Reserves stated by Sheridan (1955) may be depleted.
Highland Lode (D-9)	X(IZ)		O(WZ)	X(WZ, IZ)		Mostly mined.
Highview (C-9)	O(IZ)		O(WZ)	O(WZ, C)		Small. Locally the inner part of wall zone is rich in beryl.
Hunter-Louise (G-2)			O(WZ)	O(WZ)	X(C)	Spodumene zone thin; zone increases in thickness at north end.
Milton (F-4)					X(C)	Spodumene core small; spodumene crystals too small for efficient hand cobbing. Good milling ore.
Rachel D (F-7)	X(IZ)			O(WZ)		Grade of feldspar rather low. Pegmatite flat lying and not as large as suggested by outcrop.
Sky Lode (G-3)			O(WZ)	O(WZ)	X(C)	Pegmatite thin, but probably would be mined if more easily accessible.
Tenderfoot Spud (G-3)					O(C)	Pegmatite very thin, and spodumene zone is discontinuous.
Wilhelm (E-4)	X(C)			O(WZ)		Core largely removed. Moderate reserves of beryl in wall zone and dump.
Willdon (C-8)	O(C)			O(WZ)		Pegmatite small for feldspar mine. Considerable beryl.

mica mines in the Black Hills. The deposit, as mapped by Page and others (1953, p. 11), is in the wall zone of a highly irregular but concordant pegmatite that was emplaced in the tightly folded rocks just east of the crest of the Park dome. The wall zone is thickened where it follows tight folds, and the pipelike sheets thus formed were very rich in mica. Previous estimates indicate a mica content of about 25 percent.

Small amounts of sheet mica occur at the Highland Lode mine, especially in the lower part of the pegmatite, where the wall zone thickens. The only other pegmatite known to contain sheet-mica deposits of any size is pegmatite 38 (G–6), which is only 4–5 feet thick and about 250 feet long. The pegmatite is mainly concordant with the enclosing quartz-biotite-garnet schist, but locally it penetrates the country rock to form rolls in which the wall zone is thickened and contains minable deposits. Clarence Chedel, of Custer, produced several tons of marketable handcobbed sheet mica from these rolls in 1957. The rolls pinch out along plunge. The richest part of the rolls contains about 15 percent mica, but for the most part the deposit is of marginal grade. Mica books are small and seldom are more than 6 inches across.

Scrap mica is recoverable from all the feldspar mines in the quadrangle. None of the mines are exceptionally rich in scrap mica, and production is as a byproduct. The High Climb pegmatite has about 20 percent mica, mainly of scrap quality, in its first intermediate zone, and thus constitutes a sizable deposit of scrap mica (Sheridan, 1955).

FELDSPAR

The Highland Lode, Wilhelm, Big Spar No. 1, Dorothy, and Rachel D are the chief feldspar mines in the quadrangle. The richest feldspar-bearing zones of all these except the Rachel D have been largely mined out; the Rachel D was being mined in 1958. The inner feldspar-rich zones of these pegmatites contained about 50 percent perthite; only the leaner, downward extensions of these zones and the outer parts of the zones have not been mined. A feldspar deposit of moderate size at the High Climb pegmatite was described by Sheridan (1955).

Smaller feldspar deposits occur in the Willdon pegmatite (No. 78, C–8) and pegmatite 79 (D–9). At the Willdon, the inner perthite-rich zone is about 10 feet thick and 300 feet long. The exposed inner zone of pegmatite 79 is 15–20 feet thick and is very rich in perthite intergrown with quartz. At shallow depth the amount of quartz may decrease, and the perthite may be salable.

Other small feldspar deposits were described in the southeastern part of the quadrangle by Lang and Redden (1953), but none are commercially important sources.

Several pegmatites that appear to be poorly zoned may contain concealed feldspar deposits. Pegmatites 32 and 33 (E–5) contain exposed quartz-free perthite crystals that are several feet in diameter. The coarser grained perthite-rich pods of pegmatite 33 transect and offset finely banded line rock, but the feldspar deposit is of low grade. Pegmatite 32 is virtually unexplored, but a perthite- rich segregation near the middle of the pegmatite appears to warrant exploration. Moderately large feldspar deposits may also be present in pegmatite 10 (G–3).

BERYL

Beryl was found in most of the zoned pegmatites and probably occurs in all of them. Some of the layered and homogeneous pegmatites also contain beryl, but only as sparse scattered crystals, most of which occur in small fracture fillings.

In the zoned pegmatites, the richest beryl deposits are commonly along or in the inner part of the wall zone. A typical example of this occurrence is the first intermediate zone of the High Climb (pl. 34) pegmatite, described by Sheridan (1955). This zone contains an estimated 1 percent beryl (half of which is recoverable by means of hand sorting). It is one of the richest beryl-bearing zones in the quadrangle. At the Highland Lode mine (pl. 34, D–9), the beryl was found largely on the inner side of the wall zone, where some occurred as green very coarse grained crystals as much as 2 feet across. Production of beryl from this mine has totaled about 100 tons, but little remains that could easily be mined.

Another large beryl deposit occurs in the Wilhelm pegmatite (pl. 34, E–4) where the inner feldspar-rich zone of the pegmatite has been removed and the beryl-bearing wall and first intermediate zones are largely untouched. The beryl-rich zone was left when the feldspar zone was removed, either because the beryl is white and difficult to identify or because the low price of the beryl at the time (in the 1930's) discouraged mining. Along the east side of this oval-shaped pegmatite, a 3-foot-thick unit along the inner part of the wall zone was estimated to contain 0.5 percent beryl in crystals 1–8 inches in diameter. The extent of this beryl deposit is not known, but the deposit apparently pinches out on the west side.

The large Rachel D pegmatite (pl. 34, F–7) contains some beryl in the wall zone, but in general the beryl content is low, probably only about 0.15 percent. The observed beryl crystals are small; most of them only about 2 inches across. At the nearby Crown mine, Page and others (1953, p. 97) reported that possibly 75 tons of beryl occurs in the wall zone. Much of this beryl is shell beryl that is intergrown with other minerals and not easily recovered.

The lithium-bearing pegmatites, such as the Hunter-Louise (pl. 34, G–2), Tenderfoot Spud (pl. 34, G–3), and Milton (pl. 34, F–4), and pegmatites 2, 5, 25, 27, and 35, are either poor in beryl or the beryl is too fine grained to be recovered by hand.

Most of the other zoned pegmatites include small deposits of beryl a few feet wide and containing 0.2–1.0 percent beryl. Beryl deposits of this type in sections 20, 21, 29, and 28 were described by A. J. Lang (Lang and Redden, 1953). Other pegmatites in this category are Nos. 11, 19, 20, 21, 23, 30, 38, 42, 53, 54, 55, 66, 67, 69, 70, 74–80, and 84–88 (pl. 34). Of these pegmatites 19 and 70 appear to be the richest in beryl in the present exposures, and probably several tons could be recovered by small mining operations. Several tons of beryl was obtained from pegmatite 38 (G–6) during mining for sheet mica. Pegmatites 79 and 85–87 are somewhat larger than the others, and each is a potential source of several tons of beryl.

LITHIUM MINERALS

Spodumen and amblygonite are the only lithium minerals of commercial interest in the Berne quadrangle. Amblygonite occurs in quantity only in the High Climb (pl. 34, G–2) pegmatite, where it is an accessory mineral in the third intermediate zone (Sheridan, 1955, p. 94). A few crystals of amblygonite have been found at the Highland Lode (pl. 34, D–9) and Highview (pl. 34, C–9) mines.

Spodumene deposits occur in the Hunter-Louise, Sky, Tenderfoot Spud and Milton pegmatites and in pegmatites 5 and 25. Other deposits may exist in pegmatites 2, 7, and 35. The largest known deposits in the quadrangle are at the Hunter-Louise (pl. 34, G–2) described by Page and others (1953, p. 136–137). The spodumene-rich core in the Hunter-Louise is narrow but is about 600 feet long. The spodumene zone has a maximum thickness of about 15 feet, and it contains 15–20 percent spodumene. The crystals are mostly medium grained and can be recovered fairly easily by hand methods.

The spodumene deposits of the Tenderfoot Spud (pl. 34, G–3) are considerably smaller than those of the Hunter-Louise. Generally the spodumene-rich zone is only about 3 feet thick and is discontinuous along strike. The crystals are similar in size to those in the Hunter-Louise.

In the Sky Lode pegmatite the spodumene-bearing core is 3–4 feet thick and contains 25–30 percent spodumene in crystals as much as 3 feet long. The spodumene deposit may be as much as 200 feet long, for that is the exposed length of the fracture filling in which the spodumene occurs. It cannot be expected to extend to a depth greater than the thickness of the host pegmatite (pl. 34, No. 15). The deposit crops out high along the hanging wall of the host pegmatite on an approximately 45° slope, and development of the deposit would be expensive.

At the Milton pegmatite (pl. 34, F–4) spodumene is exposed for a length of at least 100 feet in a trench on the south end of a much longer pegmatite. The thickness of the spodumene-bearing zone ranges from 3 to 6 feet, and the grade ranges from 25 to 30 percent spodumene. The average grain size of the spodumene is only 1 inch, and thus the ore would have to be milled. Further exploration might expose considerable reserves of milling ore.

The spodumene deposits in pegmatites 25 (F–4) and 5 (G–2) are almost duplicates of that in the Milton pegmatite. In pegmatite 25 the zone is about 5 feet thick, as exposed, and the spodumene crystals also average only 1 inch long. The length of the deposit cannot be determined at present, but may be several hundred feet. In pegmatite 5, the spodumene zone is about 5 feet wide and not more than 120 feet long. The plunge of the keel of the pegmatite suggests that the bottom of the deposit is not at shallow depth.

OTHER MINERALS

Small quantities of columbite-tantalite and cassiterite have been found in a few of the zoned pegmatites, but production has not been more than a few hundred pounds. Columbite-tantalite has been mined at the Highland Lode and High Climb pegmatites, and also occurs at the Hunter-Louise mine.

Spodumene deposits probably also occur in some of the pegmatites containing pseudomorphs of spodumene. Where the pseudomorphs are sparse and consist entirely of clay, as in the High Climb and Highview pegmatites, there are probably no sizable deposits of fresh spodumene. In pegmatites 2, 7, and 35, however, the altered spodumene forms small fibrous aggregates of muscovite and clay similar to aggregates in the upper parts of the spodumene deposits in the Milton pegmatite and in pegmatites 5 and 25. It is believed that a deposit of an unaltered spodumene is probably present wherever these aggregates are noted. Although pegmatites 2, 7, and 35 are relatively small, they may well contain deposits of fine-grained spodumene like that in the Milton.

These fibrous mica aggregates, which are crude pseudomorphs after spodumene, appear to be a valuable tool in searching for concealed deposits of fine-grained spodumene. This is especially true in the deposits like the Milton and pegmatite 25. All the minerals in these pegmatites are very fine-grained to fine-grained, and even the perthite crystals are only a few inches in diameter. The natural surface exposures of the outer parts of these pegmatites do not appear to be zoned or to be of any value whatsoever. The author overlooked the spodumene at pegmatite 25 on first inspection, and it was not until pieces of float containing the fibrous irregular pseudomorphs were found that a second, more detailed inspection revealed the zone rich in fine-grained spodumene.

Cassiterite is an accessory in several pegmatites in the northeastern part of the quadrangle and probably occurs in several of the small zoned pegmatites. Cassiterite has been noted at the Hunter-Louise (Hess, 1909, p. 138), in the small pegmatite about 100 feet to the southeast of the Hunter-Louise (Page and others, 1953, p. 136), and in pegmatites 1, 2, and 3. Pegmatite 1 (G-1) is probably the Naiad tin claim described by Hess (1909, p. 139). The cassiterite at the Hunter-Louise described by Hess was probably in the wall zone. In pegmatites 1 and 2, traces of cassiterite occur in wall zones that are very rich in mica and also contain quartz, plagioclase, and beryl. In pegmatite 3, rock of this type forms the entire 1.5-foot thickness of the pegmatite. Plagioclase is very sparce; hence, the pegmatite consists mainly of quartz and muscovite. Pegmatites 4, 6, 13, 20, and 24 have similar quartz- and mica-rich wall zones and probably contain cassiterite. The mine workings about 0.5 mile northeast of the Old Bill mine are probably in small cassiterite-bearing pegmatites exposed near the contact of the quartz-biotite-garnet schist and the mica schist units. Dump specimens indicate that these workings developed thin mica-rich pegmatites similar to pegmatites 1, 2, and 3. All the tin deposits are small and of low grade, and the possibility that they will be developed is very remote.

GOLD

Several lode and placer gold deposits were mined in the Berne quadrangle at various times since about 1879. None of the mines were operated longer than a few years and the total value of the gold produced probably did not exceed $200,000. All the gold mines were inactive in 1958, and the gold was last produced in the late 1930's. However, a small mill was being erected at the Pine Tree mine in 1957–58. During the active mining period in the late 1930's, P. T. Allsman, U.S. Bureau of Mines, prepared brief accounts of the gold mines in the Black Hills (Allsman, 1940) and described most of the gold mines in the Berne quadrangle. Because nearly all the mines have become inaccessible since then, Allsman's report contains the best available data on most of these mines. Much of the following data on the extent of workings is from Allsman's report. Some of the old shafts and prospects cannot now be readily identified with the mine or claim to which they originally belonged. Many claims were unpatented, and the claim titles have long since lapsed.

Lode deposits

Most of the lode deposits are along nearly vertical quartz veins that strike northeastward or northwestward. The veins range in thickness from a few inches to as much as 6 feet, and they are as much as 500 feet long. Commonly the contacts of the veins are sharp, and the wallrock may be enriched in graphite or tourmaline. Along many of the veins the wallrock is bent and contorted against the quartz. Little is known about the mineralization of most of the veins, but free gold and such sulfides as arsenopyrite and pyrrhotite are present in some of them. Galena and sphalerite have been found at the Grand Junction mine, and Allsman (1940, p. 129) stated that tetradymite and sylvanite were reported at the Rough Rider mine. Sulfide minerals are not abundant in dump material from most of the veins.

The Western Star and Grand Junction mines are somewhat different from the other mines in that they do not follow individual quartz veins. The Western Star mine is on small quartz veinlets and sulfide-impregnated biotite and amphibole schist in the upper Crow Formation, very near the axis of the Western Star syncline. Locally, massive quartz veins crop out in the area west of the shaft, but most of the area around, and west of, the mine is covered by soil and debris from the nearby Paleozoic rocks. Mineralization may be much more extensive than is now apparent. However, past prospecting of this locality may have been thorough, and physical evidence of it may have since been obliterated.

The Grand Junction mine was originally named the Grand Junction and Hartford and was first operated in 1879. A very large quartz vein dominates the geology at the mine, but the quartz in the vein is nearly barren of gold. Apparently the ore is associated with graphitic schists and grunerite-rich meta-iron-formation adjacent to the quartz vein. The Grand Junction fault projects below the mine workings and probably influenced the mineralization and the localization of the large quartz vein. Similar, but smaller, veins are exposed along the fault for a considerable distance to the north. The old mine workings consist of adits, several shafts, and two glory holes. In 1950 George Campbell and Glen Ventling, of Custer, drove an adit from the west side of the quartz vein and penetrated the meta-iron-formation along the footwall (east side) of the vein. A tape-and-compass map of this adit is shown in figure 143. Near the fork in the adit are small stringers of galena that are very rich in silver. Apparently all analyses of quartz

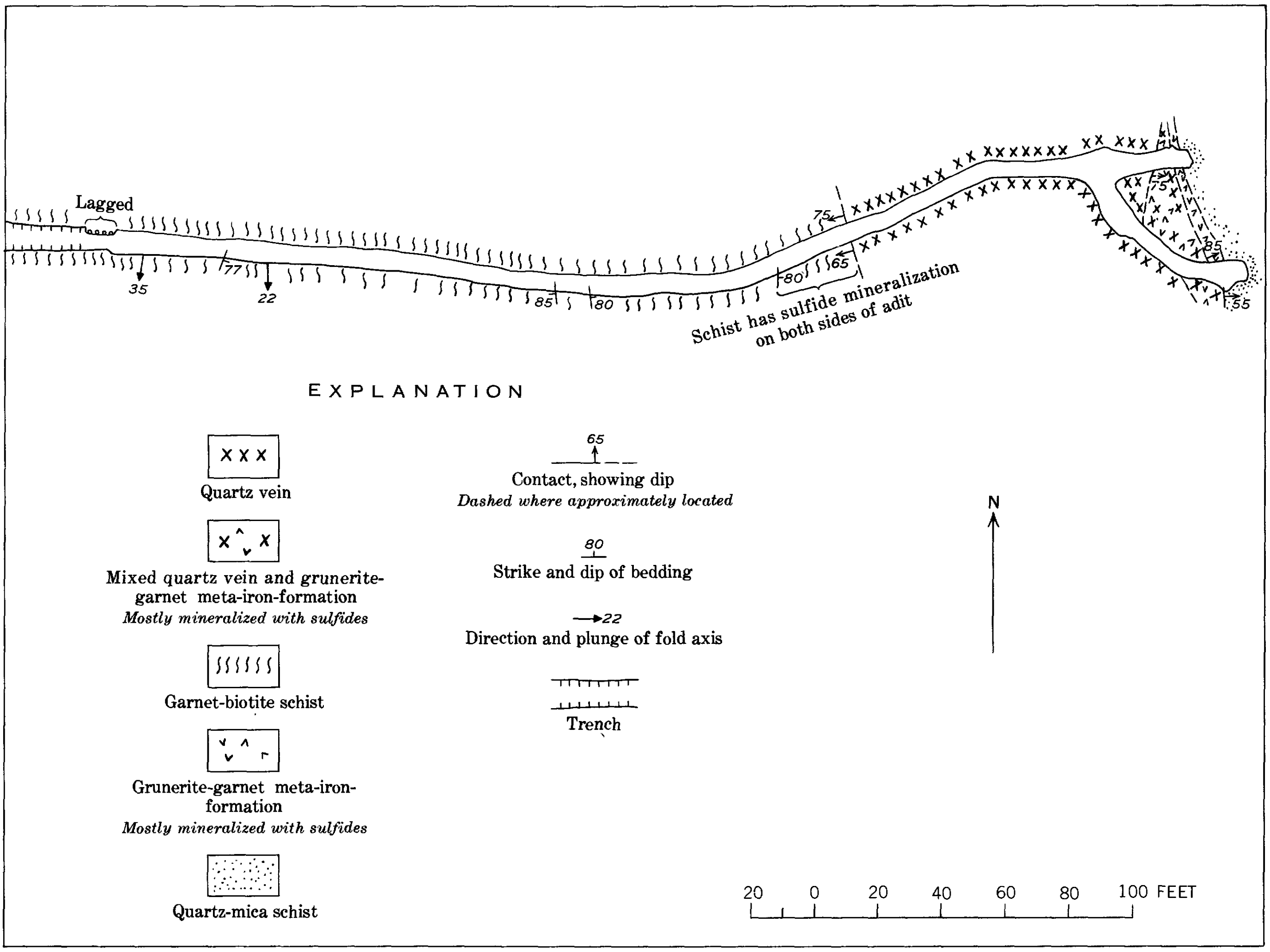


FIGURE 143.—Tape-and-compass map of the Campbell-Ventling adit, Grand Junction mine.

in the adit from the footwall part of the vein were low in gold. The meta-iron-formation and some of the adjacent schist contains abundant arsenopyrite. The meta-iron-formation at the end of the crosscut is sufficiently rich in magnetite to cause noticeable compass deviations.

The Grand Junction deposit is in tightly folded rocks above the Grand Junction fault. The intersection of the meta-iron-formation and the fault should be a favorable place for exploration, but early mining may have reached this intersection before being discontinued. The geology of the deposit is so complex that detailed plane-table mapping and exploratory drilling would be needed to ascertain the ore controls satisfactorily. Northeast-trending cross folds may have been a major factor in the localization of the ore.

Available data on other gold mines is summarized in table 13. Most or all of the mines are on quartz veins. The Old Bill mine, like the Grand Junction, is on a quartz vein along the Grand Junction fault. Most of the gold-quartz veins in the various mines are too low grade and too small to warrant development. The Western Star and Grand Junction mines appear to be the only mines that possibly contain sizable unrecognized deposits.

Placer deposits

French Creek has been extensively placered at least as far west as the Highland Lode mine. The extent of the placer operations is not accurately known, for some of the placer leases required that the ground be releveled and the soil returned so that the land could be farmed. Allsman (1940, p. 138–139) described one dragline operation in the southeastern part of the quadrangle. The gravels and the upper 1 foot of the bedrock apparently

TABLE 13.—*Description of gold mines, Berne quadrangle*

Mine	Location	Mine workings [1]	Wallrock	Description of deposit
Echo	NE¼ sec. 14, T. 3 S., R. 3 E.	120-foot shaft	Quartz-mica schist	Quartz veins: strike N. 65° W.; dip 85° SW; N. 15° E; dip 20° NW.
Grand Junction	W½ sec. 29, T. 2 S., R. 4 E.	Numerous adits and shafts and 2 open pits.	Grunerite-garnet meta-iron-formation; biotite-schist.	Replacement deposit adjacent to large quartz vein.
Gold Fish	SE¼ sec. 3, T. 3S., R. 4 E.	Caved 68-foot(?) shaft	Biotite-garnet schists	Quartz vein: strikes N. 30° W.; dips 60° SW.
Hard Scrabble	SE¼ sec. 2, T. 3 S., R. 3 E.	Caved 318-foot shaft; 1,500 feet of drifts.	Quartz-mica-garnet schist	Quartz veins: strike N. 5° W.; N. 45° E.; dip vertical.
Lucky Bird	NE¼ sec. 14, T. 3 S., R. 3 E.	Shallow shaft	Quartz-mica schist	Quartz veins: strike N. 45° W.; dip 85° SW.; N. 50° E.; dip 90°.
Minnie May	NW¼ sec. 21, T. 3 S., R. 4 E.	50-foot shaft	Biotite-garnet schist	Quartz veins: strike N. 40° W.; dips 50° NE.
Newark	NW¼ sec. 29, T. 3 S., R. 4 E.	Caved 110-foot shaft	Biotite amphibole schist; cummingtonite schist.	Quartz vein(?).
Old Bill	SW¼ sec. 33, T. 2 S., R. 4 E.	140-foot inclined shaft	Mica schist; quartz-mica schist	Quartz vein: strike N.; dip 55°.
Oneonta	NW¼ sec. 21, T. 3 S., R. 4 E.	Caved 65-foot shaft; 40-foot shaft; 60-foot shaft.	Quartz-mica-garnet schist	Quartz vein.
Penobscot	NE¼ sec. 11, T. 3 S., R. 3 E.	Adit and pit about 90×50×75 feet	Quartz-mica-garnet schist; grunerite schist.	Quartz veins, northwest strike.
Pine Tree	NW¼ sec. 5, T. 3 S., R. 4 E.	90-foot(?) inclined shaft	Garnet-graphite schist; grunerite-garnet meta-iron-formation.	Quartz vein: strikes N. 5° W.; dip 60° W.
Rough Rider	NE¼ sec. 11, T. 3 S., R. 3 E.	Caved 175-foot shaft and 3 small shafts.	Quartz-mica schist	Quartz vein: strike E.; dip 65° N.
Saginaw	NW¼ sec. 1, T. 3 S., R. 3 E.	384-foot shaft; 70-foot shaft; other drifts and crosscuts.	Quartz-mica schist	Quartz veins: strike N. 46° E.; dip 85° SE.
Wabash	SE¼ sec. 23, T. 3 S., R. 3 E.	Caved shaft and underground workings of unknown extent.	Quartz-mica schist	Unknown.
Western Star	NE¼ sec. 15, T. 3 S., R. 3 E.	Caved 100-foot shaft	Biotite-amphibole schist	Northwest-trending quartz veinlets in mineralized country rock.

[1] Data largely from Allsman (1940).

contained about $1 of gold per yard, but the soil and the overburden lowered the average grade to about 20 cents per yard.

Limited placering has been done along Spring Creek—apparently all long ago—and evidence of the extent of placering has been destroyed.

QUARTZ DEPOSITS

Relatively large reserves of apparently pure quartz occur in two large veins. One vein is in the Grand Junction gold mine, which has already been described. The other vein underlies a small hill that crosses the section line between secs. 27 and 34 in the southwesternmost part of the quadrangle. This vein, though poorly exposed, apparently has a thickness of at least 50 feet at its thickest part and a length of approximately 400 feet. A small caved shaft is on the east side of the vein. The Grand Junction quartz vein is about 100 feet thick at the surface and 80 feet thick in the adit (fig. 143); it is nearly 500 feet long. The quartz exposed in the adit is massive and uniformly white and contains no readily apparent impurities. The quartz in the other vein should be of similar purity but may have more iron staining near the surface because of its proximity to the Cambrian-Precambrian boundary.

The veins should extend to a depth of at least several hundred feet, and the reserves may therefore be large. Although the deposits are far from most industrial areas, their size may encourage eventual exploitation.

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