Geology and Pegmatites of the Fourmile Quadrangle, Black Hills South Dakota

GEOLOGICAL SURVEY PROFESSIONAL PAPER 297-D

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

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By JACK A. REDDEN

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By Jack A. Redden

ABSTRACT

The Fourmile quadrangle, Custer County, S. Dak., is on the southwest side of the Black Hills uplift. About half of the quadrangle has schist and granitic pegmatite of Precambrian age; the other half has Paleozoic and Tertiary sedimentary rocks. Pegmatite mines in the area have produced potash feldspar, sheet and scrap mica, beryl, lithium minerals, and minor quantities of rarer pegmatite minerals.

Three Precambrian metamorphic rock formations, aggregating at least 15,000 feet in thickness, were mapped. These are, in ascending order: (a) the Bugtown formation, consisting almost entirely of quartz-mica schist; (b) the Crow formation, a thin unit having various kinds of calcareous and ferromagnesian gneiss and schist; and (c) the Mayo formation, consisting largely of quartz-mica-feldspar schist but also containing quartz-mica schist rich in garnet, staurolite, and sillimanite, as well as beds of calc-silicate gneiss, metagrit, and metaconglomerate. The Crow formation, although very much thinner than the other two formations, contains many different rock types; the main ones are amphibole schist, calcite-hornblende gneiss, calc-silicate gneiss, cordierite-biotite schist, microcline-biotite schist, and quartzite.

The Bugtown formation was derived largely from graywacke, subgraywacke, and impure sand. The Crow formation was probably in part originally mafic volcanic rock, but some of its units may have been derived from impure carbonate rock and from shale. The Mayo formation consisted of graywacke and silty shale prior to metamorphism. Mafic sills and dikes intruded during deformation had the composition of diabase but are now amphibolite.

Most of the metamorphic rocks are in the sillimanite zone but some in the northwest part of the quadrangle are in the staurolite zone. Southeast of the sillimanite isograd a progressive increase in metamorphic intensity is marked by a slight increase in grain size and a decrease in muscovite content.

The major structural feature of the Precambrian rocks in this part of the Black Hills is a large open syncline that plunges about 40° S. The exposed Precambrian rocks in the Fourmile quadrangle are largely on the eastern limb of this fold. The principal minor structural features visible in the Precambrian rocks are an axial plane schistosity and a bedding plane schistosity. A later northeast-trending schistosity that locally has destroyed the earlier foliation appears to have had little effect on the distribution and gross structure of the rock units. It probably was formed during the emplacement of the large masses of granite and pegmatite northeast of the quadrangle.

The Precambrian rocks are overlain by Paleozoic sedimentary rocks that dip gently southwest away from the Black Hills uplift. The Paleozoic rocks are the Deadwood formation of Cambrian age, the Englewood and Pahasapa limestones of Mississippian age, and the Minnelusa sandstone of Pennsylvanian age. The unconformity between the Deadwood and the Englewood is flat and undeformed, but the one separating the Pahasapa from the Minnelusa has numerous irregularities resulting both from erosion and postburial solution of the upper part of the limestone. The solution may have taken place during the Tertiary uplift and erosion of the Black Hills. Sand, gravel, and volcanic ash of the Oligocene White River group overlie both Precambrian and Paleozoic rocks. The distribution of these unconsolidated deposits suggests little erosion of the Precambrian rocks since Oligocene time.

About 2,300 separate pegmatite bodies are exposed in the quadrangle, chiefly in the eastern and southeastern parts. The pegmatites are classified on the basis of their dominant internal structure as layered, homogeneous, or zoned. The homogeneous pegmatites generally lack internal structure. The layered pegmatites consist of alternating plagioclase-rich and perthite-rich layers that contrast in grain size, mineral composition, or both. The zoned pegmatites consist of concentric zones of different mineralogy and texture in an unrepeated sequence.

The pegmatites consist chiefly of plagioclase, quartz, perthite, and muscovite. The most common accessory minerals are tourmaline, apatite, and garnet. Less abundant accessory minerals include biotite, beryl, lithiophilite-triphylite, amblygonite, spodumene, lepidolite, columbite-tantalite, microlite, pollucite, sphalerite, and some alteration products of the phosphate minerals.

All the pegmatite is related in age and origin to the larger masses of granite and pegmatite to the northeast around Harney Peak. Abundant evidence indicates that the pegmatites were intruded as fluid bodies. This fluid was essentially granitic in composition but was rich in boron, fluorine, and water. The composition of the more abundant layered and homogeneous pegmatites is on the feldspar side of the low-temperature part of the quartz-albite-orthoclase-water system, whereas the composition of zoned pegmatites probably is in the quartz field. Temperatures inferred from various data suggest that most of the pegmatite crystallized below 600°C and possibly in the range 500°–600°C. The inner zones of zoned pegmatites may have crystallized as low as 300°C.

The available evidence indicates that the zoned pegmatites crystallized inward from outer zones to the cores without sig-

nificant addition or subtraction of material after intrusion. Layers also formed inward from the outer contacts of individual bodies, but the layered structure is probably an effect of changes in the volatile pressures of the fluid part of the crystallizing pegmatite. Alteration of wallrock suggests that some of the volatile material escaped into the surrounding country rock where volatile pressures were presumably lower. Homogeneous pegmatites crystallized under conditions intermediate between the closed system of the zoned pegmatites and an open system in layered pegmatites in which material periodically escaped.

The abundance of pegmatite is as much as 20 percent of the enclosing rocks, and differences in abundance are systematic. Layered bodies are most abundant in areas containing much pegmatite; homogeneous pegmatites are generally in areas of intermediate abundance; and zoned ones are largely in areas containing little pegmatite. Most of the zoned pegmatites found in areas containing much pegmatite are sheet-mica deposits; those in areas of intermediate abundance are mostly feldspar deposits; and those in areas containing little pegmatite are largely beryl-scrap mica deposits and lithium deposits. The rarer pegmatite minerals are also most abundant in areas containing little pegmatite.

This distribution pattern probably reflects the relation of chemical and physical forces resulting from differentiation of the source material and is analogous to the crystallization-differentiation in zoned pegmatites. The latest differentiates, which have the lowest temperatures of crystallization, move farther outward and upward from their source, and a zonal pattern results. Local conditions during crystallization of individual bodies modify the generalized regional distribution.

Quartz-rich veins in the metamorphic rocks contain minerals such as feldspar, sillimanite, kyanite, and staurolite; they probably formed during the late stages of the metamorphism. Gold-bearing quartz veins follow fractures and are presumably younger than the other veins but are still prepegmatite in age.

The value of the past production of sheet mica exceeds that of all other pegmatite minerals. Most of the sheet mica, however, came from the large New York mine, and no other exposed mica deposit is of significance for this commodity. The other larger mines are the Tin Mountain (feldspar, beryl, scrap mica, lithium minerals), Helen Beryl (beryl, feldspar, scrap mica), and Tip Top (feldspar). Smaller mines contain deposits of feldspar, beryl, and scrap mica.

Most of the mica and beryl is produced from outer zones, and feldspar and lithium minerals are in the inner zones of zoned pegmatites. In some parts of the quadrangle, zoned pegmatites containing these minerals have distinctive structural characteristics that are useful in prospecting for new deposits.

INTRODUCTION

The Fourmile quadrangle in the southern Black Hills, S. Dak. (fig. 73), is underlain almost equally by sedimentary and metamorphic rocks. Metamorphic rocks in the northeastern part of the quadrangle are part of a Precambrian metamorphic terrane that forms the elongate core of the Black Hills. An area around Harney Peak in the southern part of this core (fig. 75) is underlain by granitic and pegmatitic rocks, and in the surrounding area there are thousands of granitic pegmatites. A few of these, such as the Etta, Peerless, and Ingersoll pegmatites have become well known because of the unusually large size of their component

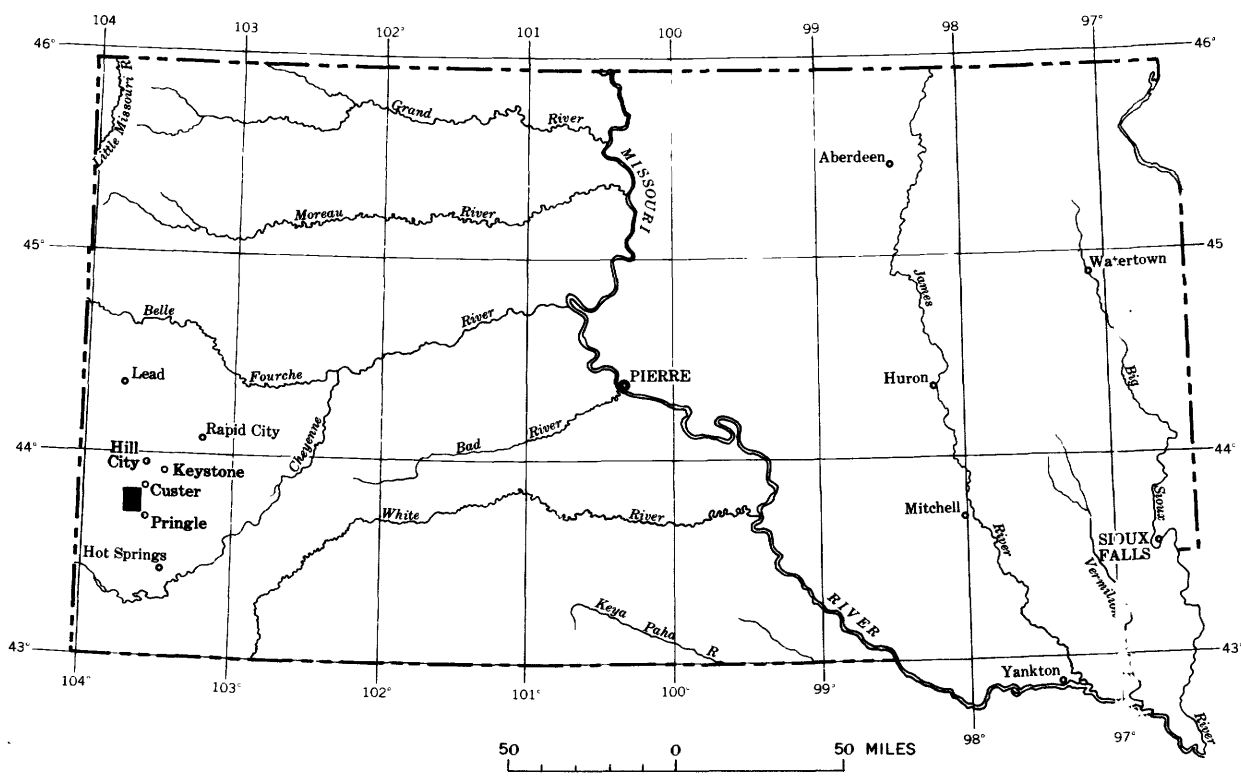

FIGURE 73.—Index map showing location of the Fourmile quadrangle, S. Dak.

minerals or because of economically minable concentrations of industrial minerals. The mineralogy and texture of less well-known individual pegmatite bodies in the Fourmile quadrangle range from those approaching granite to those having a much more complex mineralogy and pegmatitic texture. Pegmatites increase in abundance from the northwestern to the eastern part of the quadrangle.

The Fourmile 7½-minute quadrangle is bounded on the north by lat 43°45′ and on the east by long 103°37′ 30″ (pl. 21). The northeast corner of the quadrangle is about 2 miles from Custer, S. Dak.

U.S. Highway 16 and several good secondary roads cross the area. Logging and mining trails are so numerous that, during the summer months, it is possible to drive within a quarter of a mile of almost any place in the area. A freight line branch of the Chicago, Burlington & Quincy Railroad passes through Custer. The small settlement of Fourmile is in the north-central part of the quadrangle on U.S. Highway 16. The quadrangle is sparsely populated by ranchers and farmers.

The area has moderately rugged topography. The altitude ranges from about 4,750 feet at the southwest corner of the quadrangle to 5,889 on top of the Twin Sisters knobs (pl. 21). Paleozoic rocks form a pronounced cuesta that extends northwestward across the central part of the quadrangle and rises 200 to 300 feet above the Precambrian metamorphic rocks to the northeast. Outliers of these sedimentary rocks form flat-topped knolls or ridges northeast of the cuesta. The area of metamorphic rocks generally has a relatively subdued relief, and there is no evidence that the surface existing prior to Cambrian sedimentation has been greatly dissected. Pegmatites are very resistant to erosion; where steeply dipping they form nearly vertical cliffs that in some places rise more than 100 feet above the surrounding metamorphic rocks (fig. 74).

The drainage in most of the quadrangle is to the south. Southward-flowing Fourmile and Lightning Creeks join to form Pleasant Valley Creek, which empties into the Cheyenne River west of Hot Springs, S. Dak. The drainage of the southeastern part of the quadrangle is southeasterly toward the Cheyenne River. Approximately a square mile in the extreme northeastern part of the quadrangle is drained by tributaries of French Creek which flows east to the Cheyenne.

The climate is moderate and the annual rainfall is 15 to 20 inches. Short periods of very cold weather may be expected in winter, but daytime temperatures here are generally higher than those on the surrounding plains. The growing season is short; the main crops of the few farms are hay and oats. Most of the area is heavily timbered by ponderosa pine.

FIGURE 74.—Resistant pegmatite body (Castle Rock, Pegmatite 55). Top of pegmatite outcrop is approximately 120 feet above quartz-mica schists in foreground.

The chief industries of the area are ranching, lumbering, and pegmatite mining.

PREVIOUS WORK

The general geology of the Black Hills has been described by Newton and Jenney (1880), Van Hise (1890), Hosted and Wright (1923), Paige (1924), and Darton and Paige (1925). Connolly and O'Harra (1929) described the geology of the various ore deposits. Balk (1931) studied the relation of inclusions to foliation in the granite of Harney Peak northeast of the Fourmile area. The area east of Custer was mapped by Thomas[1] and the Keystone area was investigated by Hamilton.[2] Runner (1928, 1934, 1943) studied the structure and origin of the Precambrian granite domes and described the geology of part of the northeastern Black Hills. Noble and Harder (1948), and Noble, Harder, and Slaughter (1949) made detailed studies in the vicinity of the Homestake gold mine in the northern Black Hills.

Detailed studies of the pegmatites in the southern Black Hills were made during World War II by the South Dakota and the United States Geological Surveys. Fisher (1942, 1945) studied several pegmatite mines near Custer and the Tip Top pegmatite in the Fourmile quadrangle. Gwynne (1944) mapped the distribution of pegmatites in the Beecher Rock topographic basin, which is mostly east of the Fourmile quadrangle but includes approximately 2 square miles in the eastern part of the quadrangle. Page and others

[1] Thomas, L. C., 1932, Stratigraphy and structure of the Precambrian rocks of the southeastern Black Hills, South Dakota: Iowa Univ., Ph. D. thesis.

[2] Hamilton, R. G., 1934, Pre-Cambrian geology of the Keystone district, Black Hills, South Dakota: Iowa Univ., Ph. D. thesis.

(1953) described many pegmatite deposits in the southern Black Hills and most of the larger pegmatite mines in the Fourmile quadrangle.

Lang and Redden (1953) described the geology of a 7-square-mile area north of Fourmile. This area is largely in the Berne quadrangle, but about 1 square mile is in the northeast part of the Fourmile quadrangle. The geology of the overlapping area is somewhat modified on plate 21.

FIELDWORK AND ACKNOWLEDGMENTS

The present work began in 1948 when A. J. Lang, Jr., mapped an area just north of the Fourmile quadrangle and D. B. Stewart mapped a small area southwest of Fourmile. In 1949 and 1950 Redden and Lang mapped the part of the quadrangle that is north of U.S. Highway 16. The rest of the quadrangle was mapped by Redden, with the assistance at various times of R. E. Roadifer, D. B. Tatlock, Robert Lawthers, and J. E. Roadifer.

Approximately 3 square miles of the quadrangle were mapped in 1948–49 as part of the beryllium program of the U.S. Geological Survey, which was carried out on behalf of the Division of Raw Materials of the U.S. Atomic Energy Commission.

Mapping was first done on aerial photographs enlarged to a scale of approximately 1 : 12,000. In 1954 a topographic map with a 40-foot contour interval became available, and geology was mapped on this base at a scale of 1 : 10,000. Previously mapped geology was transferred to the new base at a compilation scale of 1 : 20,000. Detailed maps of mines and pegmatites were prepared with a telescopic alidade or with tape and compass.

The help and assistance of the ranchers and miners in the area is gratefully acknowledged. Mr. Montgomery Heumphreus supplied considerable background information on some of the less known deposits. Thanks and acknowledgements are extended to those geologists who participated in the mapping. However, mistakes and improper conclusions are the writer's responsibility.

REGIONAL GEOLOGIC SETTING

The Black Hills uplift is a north- to northwest-trending oval dome consisting largely of Precambrian crystalline rocks flanked by younger sedimentary rocks. The crystalline rocks are mostly metamorphosed graywackes except in the southern Black Hills where there are many sills and dikes of granite, pegmatitic granite, and pegmatite. It is generally supposed that sedimentary rocks covered the entire dome during Paleozoic and Mesozoic time, and that erosion exposed the core of Precambrian rocks during the Tertiary. Prior to the erosion, a few bodies of Tertiary igneous rocks were emplaced in the northern Black Hills.

The granite and pegmatitic granite are predominantly in a large group of intrusions in the vicinity of Harney Peak (fig. 75). These intrusions form a subsidiary dome near the center of the southern Black Hills. The outer contact of the granitic rocks as shown in figure 75 is gradational in the sense that it marks the place where these rocks become less abundant than the rock they intrude. The adjacent metamorphic rocks contain many thousands of pegmatite bodies, which in general tend to decrease in abundance away from the granitic rocks.

The major structural feature in the Fourmile quadrangle is a large syncline (fig. 75). East of the quadrangle there is a northwest-trending overturned isoclinal anticline. Most of the east limb has been removed by a major fault (fig. 75). Dips are mostly southwest and the plunge is generally south-southeast. The fault that has removed the east limb of the anticline is largely a strike fault. The displacement is not known, but it must be moderately large because rocks cannot be correlated from one side to the other even though the fault has been traced for about 20 miles. The axial plane of the anticline and the fault plane are deformed by a small dome about 4 miles northwest of Custer. The fault plane has a moderate to gentle dip around this dome but steepens away from it.

South of Custer and west of the fault, the regional trend is nearly north. The intersection of this northerly trend with the more northwesterly trend to the west of Custer results in a warp or flexure whose axis trends northeast across the Fourmile quadrangle toward Calamity Peak.

The structure is less well known east of the main fault. The gross pattern indicates that the metamorphic rocks diverge around the granitic rocks of the Harney Peak area. Near Hill City, structural trends are northeast and at Keystone they are northwest. Though it appears that the intrusive granitic rocks have distended the metamorphic rocks, as concluded by Darton and Paige (1925, p. 4–5), by Runner (1943), and by Balk (1931), additional detailed mapping is needed to verify this relation. It is possible that east of Custer another major fault (Runner, 1943, p. 437) separates the quartz-mica schist from the quartzites, and if this is so, the amount of distension of the metamorphic rocks around the intrusive rocks at Harney Peak may be much less than is suggested in figure 75.

The Precambrian metamorphic rocks of the Fourmile quadrangle have not been directly correlated with

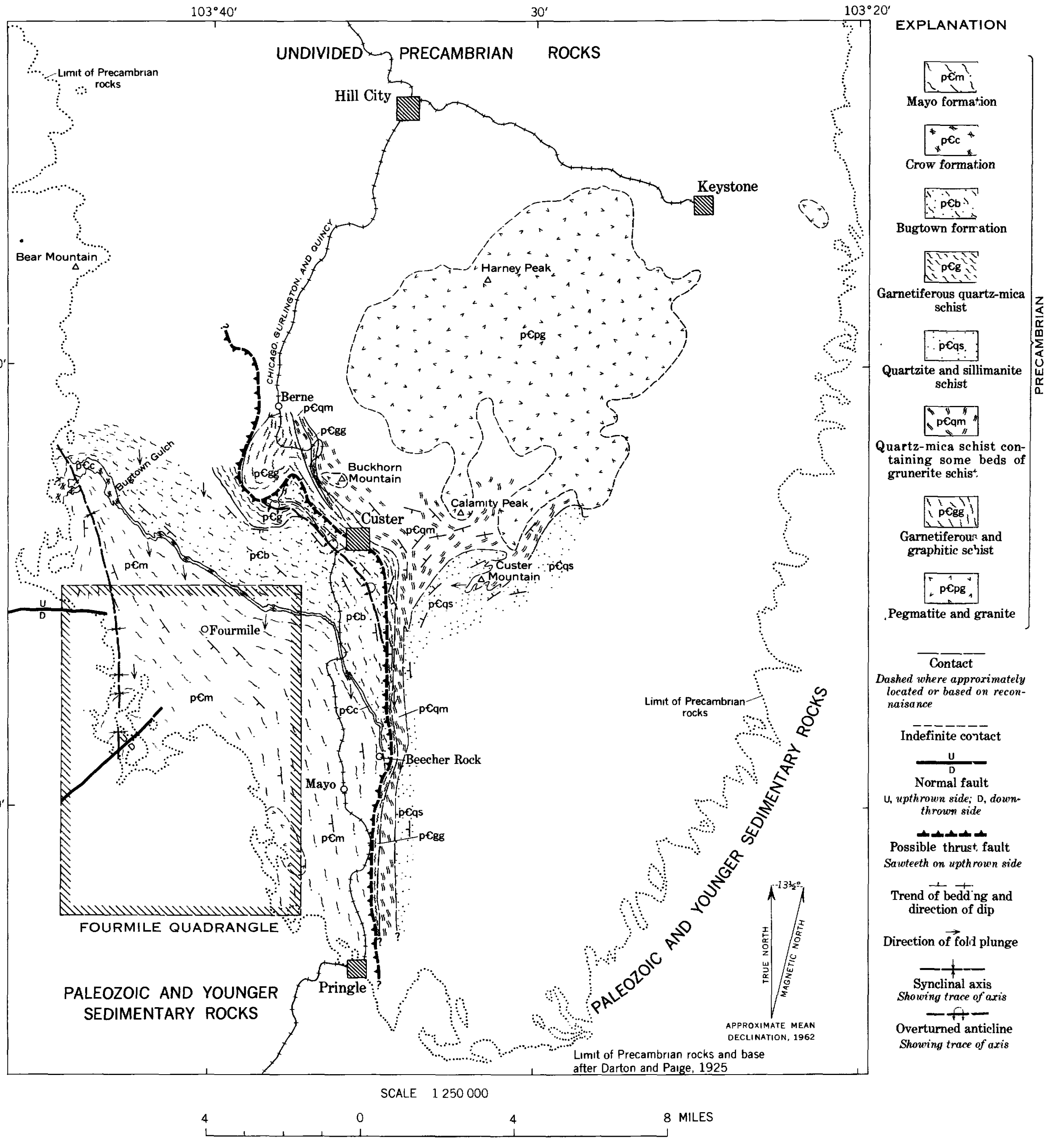

FIGURE 75.—Generalized geologic map of part of the southern Black Hills.

those in other parts of the southern Black Hills. They continue northward into the Berne quadrangle, but on the east they are transected by the fault. They are most similar to rock units near Keystone that presumably overlie the metamorphic rocks intruded by granite rocks near Harney Peak. The metamorphic rocks just east of Bear Mountain (fig. 75) are believed to underlie those in the Fourmile quadrangle. These underlying rocks resemble those that are intruded by granite and pegmatite in the Harney Peak area.

The rock units of the Fourmile quadrangle have not been definitely recognized elsewhere in the Black Hills. However, six metasedimentary formations in the vicinity of the Homestake mine at Lead, S. Dak. (Noble and

Harder, 1948), are similar in lithology to rocks found in the Bear Mountain area.

PRECAMBRIAN ROCKS

METAMORPHIC ROCKS

The metamorphic rocks of the Fourmile quadrangle (pl. 21) are divided into the Bugtown, the Crow, and the Mayo formations and mafic sills and dikes, now amphibolite. The Bugtown formation is the oldest and consists predominantly of quartz-mica schist, derived from graywacke and impure sandstone. The Crow formation, of intermediate age, is a thin unit in which the most distinctive rocks contain amphibole, calcite, and microcline. The Mayo formation is youngest. It is lithologically similar to the Bugtown formation but also contains some garnet- and staurolite-bearing schist and quartz-feldspar gneiss derived from shale and conglomerate respectively.

In general the Precambrian rocks become progressively older toward the northeastern part of the quadrangle. The total thickness of the rocks approximates 15,000 feet, but it is difficult to make an accurate estimate because of folding, plastic deformation, and the scarcity of marker horizons.

These rocks are of medium to high metamorphic grade. Most of the area is in the sillimanite zone and only the northwest part is in the staurolite zone.

BUGTOWN FORMATION

A thick unit of quartz-mica schist, part of which is found in the extreme northeastern part of the Fourmile quadrangle (pl. 21), is here named the Bugtown formation. The name is derived from Bugtown Gulch (fig. 75) in the Berne quadrangle, which adjoins the Fourmile quadrangle on the north. Bugtown Gulch crosses a well-exposed section of the formation. The formation is moderately resistant and forms large outcrops as typified by those along U.S. Highway 16, southwest of Custer.

The lower contact of the Bugtown formation is not exposed within the limits of the Fourmile quadrangle. This contact on figure 75 has been tentatively placed between massive quartz-mica schist typical of the Bugtown and an unnamed unit of garnetiferous schist that lies to the northeast.

The upper contact is where quartz-mica schist is succeeded by rocks rich in the ferromagnesian minerals characteristic of the Crow formation. This contact is tightly folded and poorly exposed. In many places, however, it is not so tightly folded as the beds in the upper part of the Bugtown formation, mainly, no doubt, because of the differences in competency and slippage of beds along the contact during folding. One of the results, however, is a strong local discordance that suggests the upper contact of the Bugtown is an angular unconformity. In the absence of marker beds in the Bugtown, the presence or absence of such an unconformity in this area cannot be demonstrated.

The Bugtown formation consists of an estimated 4,000 feet of quartz-mica schist, but only about 1,500 feet are exposed in the Fourmile quadrangle. These estimates may be in considerable error because of unrecognized folds, faults, and effects of plastic deformation. Mapping of the Berne quadrangle to the north, when completed, may show that another 2,000 feet of garnet-bearing and quartz-mica schists should be included with the Bugtown.

SCHIST

The formation consists almost entirely of grayish medium-grained quartz-mica schist in medium to thick beds. Thin-bedded dark micaceous schist, which rarely contains garnet or sillimanite, is interlayered with the thick-bedded, more quartzose schist. The schist is remarkably uniform vertically and laterally, and single beds can be traced only short distances. Modes of typical samples in table 1 indicate that the schist consists largely of quartz, mica, and feldspar. Muscovite is commonly more abundant than biotite and plagioclase more than microcline. Table 2 contains a chemical analysis of the massive quartz-mica schist (JR–1–54) that is the dominant rock in the unit, and two analyses of mica-rich schist (JR–2–54 and JR–1–52) that is sparsely distributed through the formation.

TABLE 1.—*Estimated modes of the Bugtown formation*

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Quartz	80	70	70	68	67	65	61	60	58	58	55	40	5	4	18
Muscovite	10	12	13	10	5	7	18	30	15	18	10	32	55	11	55
Biotite	9	10	11	10	15	7	15	10	15	17	15	25	35	-------	17
Plagioclase	-------	7	2	8	10	20	4	-------	10	5	15	-------	4	6	-------
Microcline	-------	Tr.	-------	Tr.	3	-------	-------	-------	-------	-------	-------	-------	-------	30	-------
Garnet	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	Tr.	-------	-------
Tourmaline	-------	-------	2	3	-------	Tr.	-------	Tr.	Tr.	-------	Tr.	1	Tr.	43	1
Apatite	Tr.	Tr.	1	Tr.	Tr.	Tr.	1	Tr.	Tr.	Tr.	Tr.	1	Tr.	-------	Tr.
Zircon	Tr.	Tr.	Tr.	-------	Tr.	Tr.	Tr.	Tr.	Tr.	-------	Tr.	-------	-------	-------	-------
Sphene	-------	Tr.	Tr.	-------	-------	-------	Tr.	-------	Tr.	Tr.	-------	Tr.	-------	-------	Tr.
Rutile	-------	-------	-------	Tr.	-------	-------	-------	-------	-------	-------	-------	-------	-------	Tr.	-------
Graphite	-------	Tr.	-------	-------	-------	-------	Tr.	-------	Tr.	Tr.	-------	-------	-------	5	-------
Opaque	-------	-------	-------	-------	Tr.	-------	-------	Tr.	-------	-------	Tr.	-------	-------	-------	Tr.
Chlorite	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	8
Sillimanite	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	5	-------	-------	-------	-------

1–11. Quartz-mica schist.
12. Quartz-mica-sillimanite schist.
13. Muscovite-biotite schist.
14. Tourmaline-microcline granulite near quartz vein.
15. Muscovite-biotite schist near a large pegmatite.

TABLE 2.—*Chemical analyses of metamorphic rocks from the Fourmile and adjacent quadrangles, Black Hills, S. Dak., and comparative analyses selected from the literature*

Constituent [1]	Group 1		Group 2			Group 3		Group 4			Group 5		Group 6	Mayo formation quartz-feldspar schist [7]	Group 7	
	Field sample JR-1-54	Average subgraywacke [2]	Field samples JR-2-54	Field samples JR-1-52	Average graywacke [3]	Field sample JR-3-54	Glenwood shale [4]	Field samples JR-4-54	Field samples JR-8-54	Garnet rock [5]	Field sample JR-6-54	Average shale [6]	Field sample JR-7-54		Field sample JR-5-54	Diabase [8]
SiO_2	79.8	77.8	73.3	72.2	64.2	53.1	56.3	41.5	49.2	41.01	58.5	58.10	72.9	72.31	50 8	51.9
Al_2O_3	10.2	9.5	13.3	14.3	14.1	19.7	19.2	21.5	5.4	18.50	18.0	15.40	12.6	10.48	13 8	15.1
Fe_2O_3	.7	.9	1.4	1.0	1.0	1.8	4.39	.7	1.3	6.57	1.2	4.02	.7	1.03	1.4	1.3
FeO	2.1	2.6	2.4	3.3	4.2	6.1		6.1	7.6	11.06	6.7	2.45	3.2	2.83	11.6	9.0
MgO	.94	1.6	1.3	1.3	2.9	1.8	1.65	6.6	21.4	11.02	3.3	2.44	1.5	2.23	6 3	6.6
CaO	1.4	1.2	.42	.9	3.5	.66	.09	7.6	9.3	10.31	.65	3.11	1.7	2.94	10 2	10.0
Na_2O	3.2	2.0	.95	2.9	3.4	.33	.19	.74	.19	.48	1.4	1.30	2.8	3.00	2.4	2.1
K_2O	1.2	1.5	4.0	3.0	2.0	9.9	10.85	6.3	.04	.31	5.5	3.24	2.3	3.05	.28	.9
TiO_2	.42	.6	.58	.53	.5	2.4	.6	2.3	.58	--------	.76	.65	.50	.51	.78	1.3
P_2O_5	.28	.2	.29	.14	.1	.30	--------	.42	.22	--------	.28	.17	.27	.22	.21	.2
MnO	.04	.2	.04	.07	.1	.06	--------	.06	.14	--------	.06	--------	.06	.07	.22	.2
H_2O	.48	1.7	1.5	1.1	2.1	2.7	5.6	4.4	3.9	--------	2.4	5.17	.72	.18	1.5	1.4
CO_2	.05	.5	.05	--------	.5	.05	--------	.05	.05	--------	.05	2.63	.05	.75	.05	--------
Total (rounded)	101	100	99	101	99	99	99	98	99	99	99	99	99	100	99	100

Description of samples and locations:
JR-1-54 Massive quartz-mica schist from the Bugtown formation 1,800 ft due E. of SW. corner, sec. 20, T. 3 S., R. 4 E., Berne quadrangle.
JR-2-54 Mica-rich schist from the Bugtown formation at same locality as JR-1-54.
JR-1-52 Bugtown formation from SE¼SW¼ sec. 35, T. 3 S., R. 4 E., Custer quadrangle.
JR-3-54 Microcline schist from the Crow formation in the SW¼SW¼NE¼ sec. 30, T. 3 S., R. 4 E., Berne quadrangle.
JR-4-54 Biotite-feldspar gneiss from the Crow formation in the SW¼NE¼SE¼ sec. 30, T. 3 S., R. 4 E., Berne quadrangle.
JR-8-54 Tremolite schist from the Crow formation in the NE¼NW¼ sec. 14, T. 3 S., R. 4 E., Berne quadrangle.
JR-6-54 Quartz-biotite-muscovite schist from the Mayo formation in the SW¼NE¼NE¼ sec. 23, T. 4 S., R. 3 E., Fourmile quadrangle (pl. 21).
JR-7-54 Quartz-mica-feldspar schist from the Mayo formation in the NE¼SW¼ sec. 17, T. 4 S., R. 4 E., Fourmile quadrangle (pl. 21).
JR-5-54 Amphibolite from intrusive body in SW¼NE¼SW¼ sec. 36, T. 3 S., R. 3 E., Fourmile quadrangle (pl. 21).

1 Rapid rock analyses by U.S. Geol. Survey. JR-1-52 was by Harry F. Phillips, Joseph M. Dowd, and Katrine E. White. Others were by Harry F. Phillips, Paul L. D. Elmore, and Katrine E. White.
2 Average subgraywacke, by Pettijohn (1949, p. 256).
3 Average graywacke, by Pettijohn (1949, p. 250).
4 Glenwood shale at Minneapolis, Minn. (Gruner and Thiel, 1937, p. 844).
5 Garnet rock from the Black Forest. Described by Rosenbusch as probably derived from rock consisting of a mixture of 48 percent carbonates and 52 percent silicates (Clarke, 1924, p. 635).
6 Average shale compiled by Clarke (1924, p. 34).
7 Specimen of Mayo formation approximately 2 miles NNW of Pringle, S. Dak. Analysis obtained from W. C. Ackerman.
8 Diabase, average of numerous analyses. (Turner and Verhoogen, 1951, p. 187).

The minerals of the schist in the Bugtown are virtually entirely metamorphic products, but a few possible relicts of the original sedimentary rocks can be recognized, though they are now recrystallized. Some of the larger aggregates of quartz or feldspar, nearly 1 mm in diameter, may have been sedimentary grains or rock fragments in graywacke. Single grains of quartz or feldspar, outlined by tiny biotite and opaque grains, also may be residual grains. Most of the quartz grains are about 0.2 mm in diameter. Parallel oriented mica plates may be as much as 1 mm long, especially in the more micaceous beds, but are commonly less than 0.5 mm. Locally a few late porphyroblasts of coarser grained muscovite and biotite cut across the foliation planes. Some diamond-shaped biotite aggregates contain laths of biotite oriented in an S-pattern that apparently resulted from rotation during growth.

The plagioclase (An_{12-30}) is untwinned and unaltered. Norms of schist in the Bugtown in table 3 have considerably more plagioclase than the estimated modes of the same specimens, and it may be that in estimating the modes some of the plagioclase was mistaken for quartz.

TABLE 3.—*Norms of field samples listed in table 2*

Mineral	JR-1-54	JR-2-54	JR-3-54	JR-4-54	JR-5-54	JR-6-54	JR-7-54	JR-8-54	JR-1-52
Quartz	53.9	52.3	13.2	0	.8	26.2	45.3	0	41.3
Albite	27.3	7.9	2.6	6.3	20.4	12.1	23.6	1.6	24.6
Anorthite	5.3	0	0	27.0	26.7	2.0	6.4	12.8	3.6
Microcline	0	0	26.7	5.0	1.7	0	0	2.2	0
Muscovite	5.0	29.0	34.0	14.1	0	31.5	9.9	0	17.6
Biotite	5.4	7.4	15.4	38.2	0	19.2	11.4	0	9.1
Diopside	0	0	0	0	19.7	0	0	25.9	0
Hypersthene	0	0	0	0	24.9	0	0	45.4	0
Olivine	0	0	0	0	0	0	0	5.3	0
Sillimanite	.6	.3	0	0	0	0	0	0	0
Garnet	0	0	0	0	0	5.5	0	0	0
Sphene	0	0	.6	5.7	0	0	0	0	0
Magnetite	1.	2.1	2.6	.9	2.1	1.9	.9	1.9	1.4
Ilmenite	.8	1.2	3.8	0	1.5	1.5	.9	1.2	.9
Apatite	.7	.7	.7	1.0	.3	.7	.7	.3	.3

Method of calculation of:

JR-1, 2, 7, 8-54, JR-1-52: Make apatite, ilmenite, and magnetite using all the P_2O_5, TiO_2, and Fe_2O_3. Use the remainder of FeO and MgO for biotite, and the K_2O remaining after biotite for muscovite. Make plagioclase from the remaining CaO plus all the Na_2O. Calculate sillimanite from remaining Al_2O_3, and quartz from remaining SiO_2.

JR-3-54: Make apatite from P_2O_5 and sphene from the remaining CaO (the original rock contains sphene). Convert the remaining TiO_2 to ilmenite, and make magnetite, biotite, and plagioclase as in JR-1-54. Divide the balance of Al_2O_3 between microcline and muscovite so that there is no remaining Al_2O_3 or K_2O.

Excess SiO_2 is calculated as quartz.

JR-4-54: Same as JR-3-54 except there is no ilmenite or quartz.

JR-5-54: Calculated by the CIPW method as described by Johannsen (1932, p. 88-99).

JR-6-54: Make ores, accessories, and plagioclase as in JR-1-54. Distribute the K_2O, Al_2O_3, FeO and MgO among muscovite, biotite, and garnet so there is no excess. The remaining SiO_2 is calculated as quartz.

Garnet forms porphyroblasts, 1 to 3 mm in diameter, in the micaceous beds. It is pink to amber and commonly euhedral. The garnet in some samples of schist cuts across the foliation, but in others the mica plates bend around it. The index of refraction of the garnet is about 1.815, and the average size of the unit cell of 3 garnet specimens as determined by X-ray powder photographs, is 11.47 A. Based on the size of the unit cell and the refractive index, the garnet is about 85 percent almandite, 5 to 10 percent spessartite, and 5 to 15 percent pyrope (Fleischer, 1937, p. 754–57).

CALC-SILICATE ROCK

The Bugtown formation contains many ellipsoidal masses and a few small lenticular beds of calc-silicate rock. Similar structural features elsewhere in the southern Black Hills were described by Runner and Hamilton (1934, p. 51–53). The long axis of the ellipsoids ranges from 2 to 40 feet in length, the intermediate axis from 3 to 20 inches, and the short axis from 0.5 to 10 inches. Others in the Keystone area are nearly perfect spheres. Most of the ellipsoids from the Fourmile quadrangle are decidely oval in cross section, but a few are almost circular. The long and intermediate axes in the ellipsoids are generally parallel to the bedding plane schistosity, but in a few they are parallel to an axial plane schistosity at a distant angle to bedding. The ellipsoids are found chiefly in the middle parts of quartz-rich beds of schist; in some thick massive beds they form 5 percent of the entire bed. Where they are this abundant, they are of use in tracing the beds. Several hundred feet of diamond-drill core from the Bugtown formation east of the Fourmile quadrangle contained from 1 to 3 percent calc-silicate rock, presumably from the ellipsoids. Rarely, the calc-silicate rock occurs as lenticular beds that are a few inches thick and extend for several feet to as much as 10 feet along the strike of enclosing beds.

The modes of the calc-silicate rocks have a wide range. The minerals commonly are distributed in such a zonal fashion that biotite, quartz, and some hornblende are in an outer rim; hornblende, epidote, and a minor amount of garnet make up the next zone; and garnet, diopside, calcite, and plagioclase are in the interior. Clinozoisite, sphene, and opaque minerals may also be present in the innermost zone and microcline in the outer zones. Some plagioclase and quartz appear in all the zones. Although some minerals are characteristic of a particular zone, locally they may not be present. A few ellipsoids contain hornblende as their major dark mineral and many are without diopside in their centers. One sampled ellipsoid consists of about 95 percent quartz and 5 percent biotite. Calcite may be absent, or may form as much as 30 percent of an ellipsoid. The centers of calcite-rich ellipsoids weather to produce a cavity bordered by many coarse garnet crystals.

The zones also differ in texture. Only the outer border of biotite-rich ellipsoids has an easily recognizable foliation, and even there it is not as well developed as in the schist. The hornblende-rich zones have a much poorer foliation, and the centers are massive and coarse grained. Crystals of garnet, hornblende, and diopside are 3 to 4 mm in diameter and hornblende crystals are as much as 20 mm long. These minerals are highly poikilitic; half of the volume of individual skeletal garnet crystals may consist of inclusions of other minerals.

The composition of the garnet and the plagioclase varies within individual ellipsoids. The index of refraction of the garnet ranges from 1.775 to 1.795; garnets with the higher index are closer to the center of the ellipsoids. X-ray determinations show that the unit cells of 2 garnets average 11.69 A. The unit cell size and index of refraction indicate a content of about 25 to 40 percent grossularite, 40 to 50 percent almandite, 5 to 10 percent pyrope, and probably some andradite (Fleischer, 1937, p. 754–757). The plagioclase from the interiors of some of the more calcite-rich ellipsoids is richer in anorthite than that from the outer zones. The composition of the plagioclase from all ellipsoids ranges from An_{40} to An_{95} and averaged An_{70}.

ALTERED ROCK TYPES

The quartz-mica schist of the Bugtown formation is altered to a dense nonschistose rock adjacent to many of the large quartz veins. This altered schist is generally only a few inches thick. It consists of various proportions of plagioclase (An_{15-22}), microcline, tourmaline, biotite, graphite, and chlorite. Specimen 14 (table 1) is an example of such a schist.

Altered quartz-mica schist next to some of the pegmatites contains abundant tourmaline, apatite, or microcline or combinations of these materials. Near some of the larger pegmatites it contains coarse chlorite as an alteration product of biotite (specimen 15, table 1). Where the alteration is most intense, the original foliation in the schist disappears and the rock resembles a granulite. It differs from the rock adjacent to quartz veins in that it does not have graphite, and it generally contains more tourmaline.

The schist in the Bugtown also contains large porphyroblasts of microcline in a few exposures in the extreme northeastern part of the quadrangle. The porphyroblasts consist of individual microcline crystals ordinarily 0.1 to 1.0 inch in diameter, with many inclusions of mica, quartz, and other minerals; they are especially conspicuous on weathered surfaces. Approximately 1 mile northeast of the quadrangle, the porphyroblasts are as much as 2 inches in diameter and form nearly 25 percent of the rock. The centers of the larger porphyroblasts are almost free from inclusions, but the distribution and orientation of the inclusions in the outer parts suggest that the porphyroblasts have formed after the development of the fabric of the schist. Exposures of the "knobby" schist commonly cover only a few tens of square feet. All observed occurrences were in areas of abundant pegmatite and most were above the hanging wall of bodies of pegmatite.

ORIGIN

The composition of the rocks in the Bugtown formation indicate that they were derived from impure sandstone, graywacke, and sandy siltstone. Table 2 contains chemical analyses of 3 specimens of the quartz-mica schist from the upper unit of the Bugtown formation as well as analyses of some average sedimentary rocks. Specimen JR–1–54 is typical of the relatively massive quartz-mica schist of the upper unit. Its analysis is very similar to that of Pettijohn's (1949, p. 256) average subgraywacke. Specimens JR–2–54 and JR–1–52 are from less massive schist richer in mica, and their analyses are somewhat similar to that of Pettijohn's (1949, p. 250) average graywacke. All three samples contain somewhat less iron, magnesium, and calcium than their sedimentary counterparts. However, the calc-silicate ellipsoids are rich in these constituents; this richness may compensate for their deficiency in schist. The general chemical similarity of the schist with presumed sedimentary counterparts, the thick-bedding, the occurrence of limited crossbedding, and the presence of apparent relict clastic grains indicate that the schist was graywacke and impure sandstone that probably originated in a geosynclinal environment (Pettijohn, 1949, p. 252–255).

The calc-silicate ellipsoids are undoubtedly metamorphosed calcite- and quartz-rich concretions, as concluded by Runner and Hamilton (1934, p. 59–61). In the Ducktown area, Tennessee, Emmons and Laney (1926, p. 20) found calc-silicate rocks of similar shapes and mineralogy, and were able to show a progressive change from metamorphosed ellipsoids to relatively unmetamorphosed concretions. Their excellent photographs of the calc-silicate rock could be duplicated in the Black Hills. The present shape of the ellipsoids is believed the result of the recrystallization of the various minerals in the concretions under unequal conditions of stress (Runner and Hamilton, 1934, p. 59–61).

CROW FORMATION

A thin unit overlying the Bugtown formation and underlying the Mayo formation is here named the Crow formation. It consists of interbedded calcareous and amphibolitic rocks, quartzite, microcline schist, and various other distinctive rocks. It was called the "amphibolite unit" by Lang and Redden (1953, p. 2–4). The present name is derived from Crow Creek just north of the Fourmile quadrangle, where the formation is well exposed. The formation crosses the northeastern part of the Fourmile quadrangle and continues northwestward several miles until it passes beneath the Paleozoic rocks. To the southeast it extends as far as Beecher Rock (fig. 75), where it thins and either pinches out or is cut off by the large fault a short distance to the south.

The upper contact is placed at the top of a thin resistant bed of quartzite that is conformable with the schist of the overlying Mayo formation. This contact is well exposed because the quartzite bed is very resistant and generally forms outcrops or abundant float. The map pattern (fig. 75) suggests that to the north in the Berne quadrangle, the formation interfingers with the overlying rocks, but even there the upper contact of each tongue of the Crow formation is marked by a quartzite bed.

The type section is 151 feet thick, and most of the better exposed sections are between 100 and 200 feet thick. The thickness is as little as 12 feet one quarter of a mile north of Beecher Rock (fig. 75). On the other hand, in the western part of the Berne quadrangle the width of outcrop is 1,600 feet, and though the rocks are so greatly folded that the actual thickness cannot be determined, it must be very great. The differences in thickness generally appear to be caused mostly by deformation.

The formation is generally poorly exposed. It can, however, be seen in many prospect pits, and its contacts can be located within a few tens of feet laterally by float and by the characteristically dark nonmicaceous soil.

The Crow formation contains hornblende schist, tremolite-actinolite schist, carbonate-bearing amphibole schist, impure marble and calc-silicate rock, cordierite-biotite schist, microcline-biotite schist, quartz-mica-feldspar schist, and quartzite. Gradational types exist between most of these rocks. Tremolite is found only in rocks of lower metamorphic grade north of the Fourmile quadrangle. Sillimanite appears in high-grade areas, but only in the few aluminum-rich rocks.

A relatively thick unit of amphibole-bearing schist occurs at the base of the formation. This schist becomes increasingly rich in calcite higher in the section. It is followed by cordierite-bearing schist and microcline-biotite schist, and then by diopside-rich calc-silicate rocks and biotite-rich schist. The uppermost unit is a quartzite bed, 3 to 8 feet thick. The rock units between the amphibole schist and the upper quartzite pinch out southeast of the Fourmile quadrangle so that near Beecher Rock the section is nearly all amphibole schist, carbonate-amphibole gneiss, and quartzite. The type section of the Crow formation is as follows:

Stratigraphic section of the Crow formation at Crow Creek in the NW¼ sec. 30, T. 3 S., R. 4 E., Berne quadrangle, South Dakota

	Thickness (feet)	Cumulative thickness (feet)
Precambrian:		
Mayo formation	---	---
Crow formation:		
Quartzite, grayish-white. Contains graphite-rich streaks	5	147–152
Covered. Inferred from other exposures to be calc-silicate gneiss	10	137–147
Partly covered. Brownish-green medium-grained biotite-hornblende-diopside-calcite-plagioclase gneiss. Biotite-rich layers alternate with diopside-rich layers.		
Partly covered. Dark-gray to black fine-grained quartz-biotite schist and microcline-biotite schist	12	115–127
Microcline-biotite-cordierite schist, grayish-brown, fine-grained, medium-bedded	10	105–115
Covered	6	99–105
Hornblende schist, dark-green, medium-grained, medium-bedded	4	85–99
Diopside-calcite-biotite-hornblende-scapolite gneiss (calc-silicate gneiss), light-green, medium-bedded	3	82–85
Microcline-biotite schist, black, fine-grained, thin-bedded	11	71–82
Hornblende schist, dark-green, medium-grained, thick-bedded	10	61–71
Calcite-hornblende gneiss, light-green; thin hornblende-rich layers alternate with calcite-rich layers	25	36–61
Hornblende schist, dark-green, fine-grained, massive	11	25–36
Hornblende schist, dark-green, medium-grained. Contains thin beds of marble(?)	25	0–25
Bugtown formation	---	---

AMPHIBOLE SCHIST

In the Fourmile quadrangle, the amphibole schist unit consists largely of actinolitic hornblende, but elsewhere its mineralogy varies. Several miles north of the quadrangle the unit is mostly brown tremolite schist, whereas near Beecher Rock (fig. 75) it is mostly dark-green hornblende schist. These changes are correlated with a northwestward decrease in metamorphic grade.

The unit is generally about 50 feet thick, but rarely it is as much as 100 feet thick. In the Berne quadrangle the thickness is considerably greater near the axis of the large syncline (fig. 75).

The actinolitic hornblende schist in the Fourmile quadrangle is moderately schistose, green to dark green, and medium to coarse grained. Most of the amphibole is in poorly oriented needles, but rarely it forms flattened spraylike aggregates approximately parallel to the plane of the schistosity. The refractive index β of the hornblende ranges from 1.65 to 1.68; the average is 1.665. The composition of small plagioclase grains interstitial to the hornblende ranges from An_{22} to An_{80} and averages An_{50}. Quartz and calcite are minor constituents. Three modes of the hornblende schist are listed in table 4 (specimens 2–4).

TABLE 4.—*Estimated modes of rocks of the Crow formation*

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
Quartz			56	Tr.	5	9	61				10	40		4		Tr.	93	53	58	64	
Biotite		2	Tr.			36	9		77		58	14	19	33	35	20		20	14	11	
Muscovite						Tr.						16			3	9	2	10		8	
Plagioclase		17	9	51	18	5	4	17		6	17			4	16	Tr.		5	Tr.		
(Percent cyorthite in plagioclase)		(33)	(78)	(28)	(35)	(75)		(40)			(26)				(80)						
Microline							4		12	19		22	48	33	6	64				11	
Hornblende		78	19	44	19	49	16	13	11	5	14										
Tremolite	93																				
Cummingtonite																					37
Diopside								26		56											
Epidote			3					8													
Cordierite													24	22	36				9		
Sillimanite																		5			
Kyanite																		3			
Scapolite						Tr.				Tr.											
Chlorite		Tr.			3											Tr.		4			
Magnetite	3	Tr.	Tr.	3							Tr.			1	Tr.	6					
Fuchsite(?)																			15	3	
Apatite								Tr.			Tr.				Tr.		Tr.			Tr.	2
Tourmaline		Tr.	Tr.				Tr.										Tr.				35
Sphene	4	Tr.	3	1			Tr.	5		3			7								3
Rutile														2					1	1	1
Zircon											Tr.					Tr.					
Calcite			8		55		5	29		13		5									
Graphite			Tr.			Tr.					Tr.				3	Tr.	3		Tr.	Tr.	22
Sulfides												3									2

1. Tremolite schist.
2. Hornblende schist.
3. Quartz-hornblende gneiss.
4. Plagioclase-hornblende schist.
5. Calcite-hornblende gneiss.
6. Hornblende-biotite-quartz schist (associated with calc-silicate gneiss).
7. Quartz-hornblende schist (interbedded with calcite-hornblende gneiss).
8. Calcite-diopside-hornblende gneiss (calc-silicate gneiss).
9. Biotite-hornblende schist interbedded with rocks represented by specimen 8.
10. Diopside-calcite-microcline gneiss (calc-silicate gneiss).
11. Biotite-gneiss layer associated with calc-silicate gneiss.
12. Quartz-microcline schist.
13. Microcline-cordierite-biotite schist.
14. Biotite-microcline-cordierite schist.
15. Cordierite-biotite-microcline schist.
16. Microcline-biotite schist.
17. Quartzite.
18. Quartz-mica-plagioclase-sillimanite schist.
19. Quartz-biotite-fuchsite(?) schist.
20. Quartz-mica-microcline schist.
21. Cummingtonite-tourmaline-graphite schist (associated with gold-bearing quartz vein).

Northwest of Crow Creek the amphibole schist unit is distinctly lighter green and typically contains cavities lined with iron oxide. Magnetite and ilmenite are somewhat more abundant there than in the Fourmile quadrangle. The average β index of the amphibole of about 1.64 indicates that it is actinolite. In the most northern exposures of the Crow formation (fig. 76), the amphibole schist is darker green and fine to medium grained. In a few places it contains ellipsoidal aggregates of calcite, quartz, chlorite, and ilmenite. These aggregates are about 1 inch long, flattened parallel to the foliation, and elongate parallel to the plunges of fold axes. The calcite commonly has been removed by weathering and the other minerals of the original aggregates line the cavities.

CALCITE-HORNBLENDE GNEISS

Pale green calcite-hornblende gneiss is locally interbedded with the amphibole schists, but most of it lies stratigraphically above the schists. There are probably all gradations between the two rock types. The gneiss may be as much as 40 feet thick, but rarely is exposed; a few outcrops are present in the Fourmile quadrangle southeast of U.S. Highway 16.

The gneiss consists of alternating dark-colored hornblende-rich and light-colored calcite-rich layers that range from 0.05 to 5 inches in thickness, and a few thin lenses of white sugary-grained quartzite. A typical mode is that shown for specimen 5 in table 4. Locally the calcite-rich layers transect and vein the dark layers. Medium- to coarse-grained hornblende or calcite may form 90 percent of different layers. Other minerals are pale-brown magnesium-rich biotite, clinozoisite, plagioclase (An_{35-80}), and sphene. In a few layers biotite is as abundant as hornblende.

CALC-SILICATE GNEISS

Locally there are thin lenses of calc-silicate rock in the calcite-hornblende gneiss, and transitional phases between the two rocks can be recognized. The thickest unit of calc-silicate gneiss, however, ordinarily is at a higher stratigraphic level, above the cordierite and microcline schists.

The overall color of the rock is brownish green to almost black, but it consists of alternating light-colored layers rich in calcite and diopside and of dark layers rich in hornblende and biotite (specimens 8 and 9 respectively, table 4). The light-colored layers consist predominantly of calcite and diopside, with lesser plagioclase, microcline, hornblende, and sphene. The dark layers consist almost entirely of biotite and hornblende. Small grains of scapolite may occur in both the light and dark layers. A weak schistosity is formed by flakes of biotite that are matted together and surround larger green hornblende grains. Some poikilitic diopside and hornblende are oriented with their long axes parallel to bedding. The diopside occurs as short crystals of medium- to coarse-grain size, a few of which are as much as 5 inches long. The biotite is pale brown to black; the brown variety has relatively low refractive indices and is rich in magnesium. Fresh untwinned very fine grained plagioclase (An_{40-90}) is most abundant on the outer edges of the light-colored beds. Microcline is fine grained and disseminated throughout the rock. Some cordierite(?) is present in the biotite-rich layers. Sphene is locally abundant and easily visible in hand specimens. It forms approximately 10 percent of some specimens from the SE 1/4 sec. 33, T. 3 S., R. 4 E. Megascopic crystals of rutile are associated with the more amphibole-rich rock, and in thin sections sphene is observed surrounding rutile grains. Several specimens contain acicular tourmaline crystals as much as 2 cm long.

The mineralogy of the gneiss is somewhat similar to that of the calc-silicate ellipsoids in the Bugtown formation but it has more potassium- and titanium-bearing minerals than the ellipsoids and does not contain garnet.

Northwest of Crow Creek in the Berne quadrangle (fig. 75) where the associated amphibole schist unit is characterized by light-colored actinolite, the calc-silicate gneiss unit is characterized by beds of nearly pure tremolite which appears to have taken the place of the diopside and some of the associated calcium and magnesium minerals. This conclusion is borne out by the general absence of cordierite-bearing rocks in the area that contains the rich tremolite rocks. The tremolite has a β index of about 1.627. The needles of tremolite are 2 to 5 mm long and form an intergrown network, so that the rock is only slightly schistose. Specimen 1 in table 4 is typical of the tremolite beds. The richness of the sample in ferromagnesian minerals is evident from the analysis of a similar sample (JR–8–54, table 2) and from the corresponding norm (table 3).

CORDIERITE-BIOTITE AND MICROCLINE-BIOTITE SCHISTS

Interbedded cordierite-biotite and microcline-biotite schists, about 30 to 40 feet thick, lie above the amphibole-bearing schist and hornblende-calcite gneiss. Though these schists rarely crop out, there are a few good exposures in the SE¼ sec. 33, T. 3 S., R. 4 E. (pl. 21).

The cordierite-rich schist has a mottled appearance caused by chalky white to gray cordierite grains, almost 1 mm in diameter, in a finer grained matrix of biotite. Individual cordierite grains protrude from weathered surfaces and are somewhat flattened parallel to an imperfect schistosity marked chiefly by biotite flakes. Some of the rock extremely rich in cordierite is fine grained and has the distinct purplish blue tint characteristic of cordierite.

The mineral content of the cordierite-rich gneiss is very variable as shown in three typical modes in table 4 (specimens 13–15); any of the minerals cordierite, microcline, plagioclase, or biotite may predominate in individual specimens. Sphene locally forms as much as 10 percent of the gneiss.

The grains of cordierite are nearly equant, are very poikilitic, and have the characteristic pseudohexagonal twinning. Boundaries between the twin segments are very commonly irregular, jagged, and possibly sheared. Most of the cordierite is optically negative with a large 2V and a β index of about 1.550.

The other minerals are varied in habit and appearance. Biotite forms medium-grained shiny black to brown plates. Most of the feldspar can be identified only in thin sections. Microcline occurs both as small poikilitic grains bordering and replacing cordierite and as pale blue easily recognizable aggregates several millimeters in diameter. Plagioclase ranges from 0.2 to 1 mm in diameter and commonly is untwinned. The appearance of the grains in hand specimens and thin sections is very similar to the cordierite. The probable range in composition of the plagioclase is from An_{15} to An_{80}. Grains of sphene locally exceed 1 mm in length and are easily visible on the weathered rock.

An analyzed sample of this rock (JR–4–54, table 2) consists of biotite-feldspar gneiss that proved to be unusually rich in biotite (45 percent) and plagioclase (30 percent) and lean in microcline (20 percent) and cordierite (trace). In the norm (table 3) Al_2O_3 is used up in micas and feldspars, and MgO appears in biotite, with the result that no cordierite is calculated.

The microcline-biotite schist is dark gray to black, is much finer grained than the cordierite-rich rocks, and has a better developed schistosity than the other rocks of the Crow formation. Its maximum thickness is probably less than 20 feet. The schist contains thin beds, an inch or less in thickness, which are easily recognized by small differences in color. Some of the schist is interbedded with cordierite-biotite schist and calc-silicate gneiss. Biotite is visible in hand specimens and microcline, muscovite, quartz, and small opaque blebs, presumably magnetite, are found in thin sections. The potassium-bearing minerals constitute 75 to 95 percent of the schist. The cordierite-rich rocks are also rich in potassium, and there are probably all gradations between the two rocks.

The mode of specimen 16 (table 4) and the chemical analysis and norm of a similar specimen (JR–3–54, tables 2 and 3) from the same outcrop are typical of this rock. The high Al_2O_3 content of the analysis is such that microcline of the mode appears largely as muscovite in the norm. This excess of Al_2O_3 suggests that some of the fine-grained unidentifiable material in the rock consists partly of cordierite.

QUARTZITE

A bed of quartzite adjacent to the upper contact of the Crow formation characteristically has thin lenses and streaks of graphite and hematite or limonite. The thickness is about 5 feet on the limbs and as much as 30 feet in the noses of folds. Contacts of the quartzite with the enclosing schists are sharp and quite unlike the gradational contacts between the other types of rocks in the Crow formation.

A second thin bed of similar quartzite is locally exposed near the center of the Crow formation in a few outcrops east of U.S. Highway 16.

The lenses and streaks of graphite and hematite are generally parallel to bedding and probably are relict sedimentary structures. The high purity of the quartzite is shown by the mode of specimen 17 (table 4). Traces of tourmaline, of muscovite, and of apatite are the only accessory minerals besides graphite and hematite. Many outcrops contain cavities that may have been derived from the weathering of sulfides or carbonate minerals.

OTHER ROCKS

Other rocks are relatively rare in the Crow formation. The most abundant are a few thin beds of quartz-mica-feldspar schist that are scattered throughout the formation, but are mainly in the upper part. In the Fourmile quadrangle, the quartz-mica-feldspar schist is generally sillimanite-bearing. Some kyanite occurs in the schist in lower grade areas north of the quadrangle (specimen 18, table 4). Cordierite and a green mica (fuchsite?) also appear in some of these beds. A spectrographic analysis of the mica by N. M. Conklin of the U.S. Geological Survey indicated a Cr content of almost 1 percent; a more accurate determination was not obtained because the sample was too small.

Cummingtonite and biotite-tourmaline schist adjacent to quartz veins north of the Fourmile quadrangle apparently are altered varieties of other rocks of the formation.

ORIGIN

The rocks of the Crow formation are in many respects different from the predominantly clastic rocks of the Bugtown and the Mayo formations. The composition and structure suggest that the original rocks of the Crow formation were mainly dolomite, shale, chert, and perhaps volcanic rocks.

The well-layered amphibole rocks rich in MgO in the upper part of the formation can be reasonably explained only as impure carbonate sediments that lost CO_2 during metamorphism. The only igneous rocks comparable with specimen JR–8–54 (tables 2 and 3) are ultramafics, and there is no evidence that the amphibole rocks in the Crow formation are intrusive. The metamorphism of carbonate rocks to form very different metamorphic products is illustrated in table 2 by an analysis of a garnetiferous rock that Rosenbusch (Clarke, 1924, p. 625) showed was derived from a carbonate-rich sediment.

Some of the Crow rocks that contain hornblende instead of tremolite, and more plagioclase than the tremolite-rich schist, resemble basalt in composition. Furthermore, in a few places they contain small ellipsoidal aggregates of calcite, quartz, chlorite, and ilmenite that resemble amygdules found by Harker (1939, p. 106) in metamorphosed basalts. Thus the available evidence is in accord with the supposition that these rocks have a volcanic origin.

The calcite-hornblende gneiss and calc-silicate gneiss, like much of the amphibole schist were probably derived from impure carbonate rocks. They are relatively well bedded and rich in calcium and magnesium minerals as well as carbonate. Lenses of pure sugary-grained quartz in the calcite-hornblende gneisses suggest derivation from chert lenses. The relicts of thin beds and the high Al_2O_3 content of the microcline-biotite schist indicate that it is a metamorphosed shale. The K_2O content is considerably higher than the 3 to 4 percent in the average shale. The literature contains at least three analyses of bedded rocks of similar K_2O content (other than the salines). One of these is an analysis of an Ordovician shale described by Gruner and Thiel (1937) that is very similar to the analysis of the microcline-biotite schist (see table 2 for comparison). Another shale described by Schwartz (1935, p. 528) had 9.84 percent K_2O. The high K_2O content of some bentonites (Ross and Hendricks, 1954, p. 66) also indicates the possibility that this schist may have been a volcanic ash.

The cordierite-biotite gneiss also has an Al_2O_3 content that suggests shale or clay, but its high content of

MgO and CaO suggests that the shale was rich in carbonate. The rock is also rich in K_2O, some of which may have been introduced from neighboring beds of microcline-biotite schist, as suggested by the textural evidence that microcline replaces cordierite.

The purity of the quartzite in the Crow formation and the virtual absence of relict grains or structures indicate quite a different origin from the clearly clastic quartzose rocks of the Bugtown and the Mayo formations. The probable sedimentary equivalent of the quartzite is chert rather than sandstone.

MAYO FORMATION

A thick unit of quartz- and mica-rich schist, conformably overlying the Crow formation, is here named the Mayo formation. The name is derived from the old settlement of Mayo, 1 mile east of the Fourmile quadrangle (fig. 75). A well-exposed section is found between the contact with Paleozoic rocks in the eastern part of the Fourmile quadrangle and Mayo, and from Mayo northeastward to Beecher Rock (fig. 75). The most complete section in the Fourmile quadrangle extends from 2 miles northeast of Fourmile along U.S. Highway 16 to 4 miles south-southwest of Fourmile along Fourmile Creek (pl. 21). At least 14,000 feet of the formation is exposed in the Fourmile quadrangle. However, the total thickness is not known because of the cover of Paleozoic rocks.

The Mayo formation is similar to the Bugtown, but generally is not so uniform in composition. Outcrops of the two formations are about equally abundant. Small outcrops of some of the schists in the Mayo cannot be distinguished from those of the Bugtown formation. However, beds of the Mayo are generally thinner and there are greater compositional differences between adjacent beds. Biotite is more abundant, as are calc-silicate ellipsoids, and in general the rocks of the Mayo are richer in iron and alumina than those of the Bugtown formation. Schist in the lower part of the Mayo contains abundant feldspar, whereas that in the upper part is richer in garnet and sillimanite and also contains staurolite.

The predominant rocks are medium-grained light- to dark-gray quartz-mica-feldspar schists. In addition, the formation has: (a) medium-grained quartz-mica schist that locally contains one or more of the minerals staurolite, garnet, feldspar, and sillimanite; (b) medium- to coarse-grained metagrit and metaconglomerate; (c) coarse-grained calc-silicate gneiss containing calcite, microcline, garnet, epidote, plagioclase, and diopside; and (d) cummingtonite-quartz schist.

The mineralogy of the Mayo formation in the staurolite and the sillimanite metamorphic zones is generally similar. The appearance of sillimanite and a decrease in the amount of muscovite in the higher-grade zone are the only significant changes. Sillimanite is in the more aluminous beds, which are chiefly of staurolite- and garnet-bearing quartz-biotite schist. Thin sections showing micaceous minerals partly converted to sillimanite indicate that the conventional view that staurolite breaks down to sillimanite and garnet does not apply to these rocks. Grain sizes in the schist increase somewhat toward the area of more intense metamorphism in the southeastern part of the quadrangle.

A generalized section of the Mayo formation is as follows:

Generalized stratigraphic section of the Mayo formation along the line of section A–A′ and Fourmile Creek (pl. 21)

	Thickness (feet)	Cumulative thickness (feet)
Precambrian:		
Mayo formation:		
Upper contact concealed by Paleozoic rocks.		
Quartz-biotite-muscovite schist, gray, medium-grained, thin- to thick-bedded; staurolite and sillimanite schist interbeds; garnet a common accessory; calc-silicate ellipsoids Partly concealed	2,400	14,000
Garnet schist, dark-gray, medium-grained; abundant calc-silicate ellipsoids	30	11,600
Quartz-mica schist interbedded with staurolite schist; unit light- to dark-gray, medium-grained, medium- to thick-bedded; sillimanite and garnet common accessories in staurolitic beds; calc-silicate ellipsoids. Largely concealed	3,900	11,570
Quartz-mica schist and quartz-feldspar-mica metagrit; unit gray, medium- to coarse-grained, medium-bedded; schist contains some sillimanite and garnet; four discontinuous metagrit beds, as much as 15 feet thick, equally spaced throughout interval; calc-silicate ellipsoids and lenses abundant in metagrit beds	1,400	7,670
Quartz-biotite-feldspar schist, gray, medium-grained, thin- to medium-bedded; some garnet, staurolite, and sillimanite; calc-silicate ellipsoids	1,000	6,270
Quartz-mich-feldspar schist, light-gray, medium-grained; accessory garnet	3,000	5,270
Quartz-feldspar metaconglomerate, gray-brown, medium- to coarse-grained, thick-bedded; pebbles as much as 1 inch in diameter; locally a thinner bed lies about 50 feet above main bed; metaconglomerate pinches out to the northwest; abundant calc-silicate ellipsoids	20	2,270

Generalized stratigraphic section of the Mayo formation along the line of section A–A′ and Fourmile Creek (pl. 21)—Con.

	Thickness (feet)	Cumulative thickness (feet)
Precambrian—Continued		
Mayo formation—Continued		
Quartz-mica schist, light-gray, medium-grained, thin- to medium-bedded	800	2, 250
Quartz-biotite-feldspar schist, gray, medium-grained; 2 calc-silicate beds about 6 feet thick at top and bottom of interval; these locally split into several thinner beds	150	1, 450
Quartz-mica-feldspar schist, gray, medium-grained, medium- to thick-bedded; clastic grains visible in specimens; some interbeds of quartz-mica schist; calc-silicate ellipsoids; poorly exposed	1, 200	1,300
Quartz-biotite-sillimanite gneiss, dark-gray, medium- to coarse-grained; sillimanite forms "knots" 2 to 3 mm across by 5 mm long; quartz-mica schist interbeds	100	100
Total exposed thickness	14, 000	
Crow formation.		

This section is generalized because in the field it is impossible to make sharp divisions between the units, except where there are thin distinctive units such as the beds of calc-silicate and metamorphosed grit and conglomerate. These thin units generally pinch out to the northwest in the Berne quadrangle, but they are generally traceable southeastward across the Fourmile quadrangle. Mica- and staurolite-rich beds tend to increase in abundance to the northwest.

QUARTZ-MICA-FELDSPAR SCHIST

The quartz-mica-feldspar schist of the Mayo formation differs very little in appearance from most of the schist in the Bugtown formation. The schist in the Mayo generally has more feldspar (both plagioclase and microcline) and is somewhat darker gray owing to a higher biotite content. Modes of samples 1 through 6, in table 5 are typical of this schist. The plagioclase in the Mayo and the Bugtown formations has approximately the same composition (An_{10-35}) and is more abundant than microcline. The schist in the Mayo has more of the apparently clastic grains of feldspar and quartz surrounded by a fine-grained matrix and in addition has quartz aggregates 1 to 2 mm long which may be recrystallized rock fragments. The grain size of the quartz-mica-feldspar schist averages 0.4 mm. Plates of mica have this size in most of the schist, but they increase to nearly 1 mm in the southeastern part of the quadrangle, where many large plates of muscovite transect the foliation, showing that the plates formed after the foliation.

TABLE 5.—*Estimated modes of the Mayo formation*

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
Quartz	69	68	31	67	40	66	68	52	50	40	26	40	21	63	34		60	28	70	62	25
Biotite	24	16	28	11	23	15	31	32	20	26	25	27	33	19	33	54	23	29	10	19	
Muscovite	Tr.	4	6	14	18	2	10	10	26	3	23	16		6	4	2	7	35	2	2	
Plagioclase	6	5	23	2	12	15		Tr.		15		2	5	3	10	43	10	3	5	2	4
Microcline		3		4	Tr.	2														15	15
Garnet			9		1			Tr.	3	1	20	3						2	Tr.		3
Staurolite			1							15		10	41	6							
Sillimanite			2		2		Tr.			Tr.		1		2	19						
Hornblende																					24
Epidote																					5
Calcite																					21
Chlorite	Tr.	2			2					1	2	Tr.			Tr.		Tr.		4		1
Tourmaline	Tr.			Tr.			Tr.			Tr.		1		Tr.			Tr.				
Apatite	Tr.	1				Tr.	Tr.				2		1				Tr.	1	Tr.		1
Sphene		Tr.			Tr.						Tr.								Tr.		
Zircon	Tr.	Tr.	Tr.				Tr.										Tr.	Tr.	Tr.		
Rutile		Tr.																			
Allanite																					Tr.
Graphite								5	2			Tr.									
Magnetite					2										Tr.						

1. Quartz-biotite-plagioclase schist.
2. Quartz-biotite-feldspar schist.
3. Quartz-mica-plagioclase-garnet schist.
4. Quartz-mica-feldspar schist.
5. Quartz-mica-feldspar schist.
6. Quartz-mica-feldspar schist.
7. Quartz-mica schist.
8. Quartz-biotite schist.
9. Quartz-mica schist.
10. Quartz-biotite-staurolite-plagioclase schist.
11. Quartz-mica-garnet schist.
12. Quartz-mica-staurolite schist.
13. Staurolite-biotite-quartz schist.
14. Quartz-mica-staurolite schist.
15. Quartz-biotite-sillimanite schist.
16. Biotite-feldspar schist.
17. Quartz-mica schist.
18. Mica-quartz schist.
19. Quartz-biotite-feldspar schist.
20. Quartz-mica feldspar gneiss (metagrit).
21. Quartz-hornblende-calcite-microcline (lime-silicate) gneiss.

The chemical analysis and the norm of a specimen of typical quartz-mica-feldspar schist of the Mayo formation (JR–7–54, tables 2 and 3) indicate a similarity to average graywacke (table 2). The texture of this schist also suggests that it is metamorphosed graywacke.

QUARTZ-MICA SCHIST

The quartz-mica schist of the Mayo formation is somewhat darker than the quartz-mica-feldspar schist and characteristically has beds containing large porphyroblasts of staurolite and garnet. Biotite is more abundant than in the feldspar-bearing schist and darkens the color. Typical modes of the quartz-mica schist are specimens 7 through 18 (table 4).

The schistose matrix of the staurolite- and garnet-rich quartz-mica schist is medium-grained like the other schist. The staurolite is in euhedral grains as much as 30 mm long. Much of the staurolite, especially within the sillimanite metamorphic zone, is in part psuedomorphically replaced by mica, predominantly muscovite. Some of the pseudomorphs have cores of staurolite surrounded by muscovite that in turn is surrounded by a rim of biotite. Many other pseudomorphs consist solely of muscovite, without relicts of staurolite or rims of biotite.

Rose to pink garnet crystals as much as 5 mm in diameter occur both in association with the staurolite and in beds lacking staurolite. The refractive index of the garnet is about 1.81, and the average of 2 X-ray determinations of the unit cell is 11.46 A. The garnet, therefore, is almandite with some pyrope (Fleischer, 1937, p. 754–57), and is similar to the garnet in the Bugtown formation.

An analysis of garnet-bearing quartz-mica schist (table 2, JR–6–54) shows it to be very similar to average shale. Although the schist is somewhat high in aluminum and potassium, there can be little doubt that it is the metamorphic equivalent of shale.

In the more highly metamorphosed areas in the southeast part of the quadrangle sillimanite appears as flattened and elongated waxy aggregates or knots, that form 5 to 10 percent of most of the quartz-mica schist. These knots consist of about equal quantities of quartz and sillimanite and average half a centimeter in diameter. Typically they are flattened parallel to the dominant foliation.

A few beds of gray quartz-mica schist from the area just north of the Tin Mountain mine contain light-colored aggregates, a few millimeters in diameter, consisting of quartz, mica, chlorite, and apatite. They are surrounded by a biotite-rich rim (fig. 76). It is unlikely that these aggregates were derived from garnet or staurolite because no intermediate stages of replacement can be seen and because fresh garnet and staurolite occur in the same rock. Probably they are relicts of an unknown earlier mineral.

METAGRIT AND METACONGLOMERATE

Metagrit and metaconglomerate gneiss occur in several beds throughout the Mayo formation. They are brownish gray and consist of pebbles of quartz and feldspar in a finer grained quartz-rich matrix. The metagrit and metaconglomerate are similar in composition. Quartz and oligoclase are the dominant minerals; microcline is a minor constituent but is more abundant than in the schist (specimens 20 and 21, table 5). Metagrit, containing flattened granules 1 to 5 mm thick, forms several beds in the upper part of the formation. Metaconglomerate occurs as a single bed or several adjacent beds, having a total thickness of 5 to 25 feet, that are well exposed just north of Fourmile (pl. 21). They pinch out several miles away along strike. The gneiss contains recognizable pebbles, some of which are arranged in a typical graded bedding structure indicating that the beds are upright. The diameter of the pebbles is as much as 2 inches. Most of them are flattened parallel to foliation and are slightly elongated parallel to the plunge of folds, so that the average axial ratios are about 2:1.5:1. Pebbles of finer grain size are elongated more than the coarser pebbles.

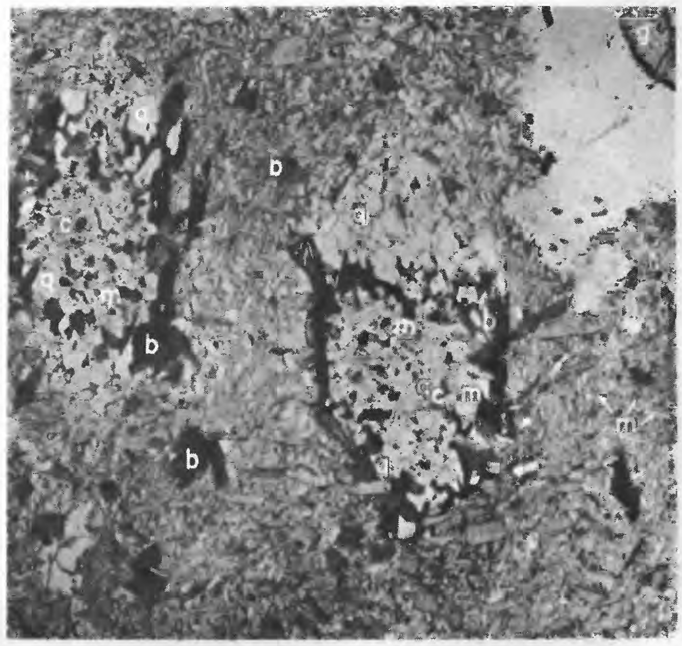

FIGURE 76.—Photomicrograph of schist from the Mayo formation containing aggregates of quartz (q), muscovite (m), and chlorite (c) which are surrounded by biotite (b) rims. One porphyroblast of garnet (g) is shown in upper right of photograph. × 30.

CALC-SILICATE GNEISS

Calc-silicate gneiss forms several discontinuous 2- to 15-feet-thick beds in the lower part of the Mayo formation, and ellipsoids like those in the Bugtown formation are found throughout most of the section. In places a single bed splits into two beds separated by a few feet of medium- to coarse-grained quartz-mica feldspar schist. The gneiss consists of layers rich in hornblende, or rarely diopside, intercalated with light-colored layers rich in calcite, microcline, and quartz (specimen 21, table 5). The calcite weathers easily, leaving pits outlined by dark minerals. In some samples hornblende, calcite, quartz, microcline, garnet, epidote, and plagioclase each form more than 10 percent of the rock and thus can be considered essential minerals. The accessory minerals are sphene, apatite, allanite(?), and chlorite. Most of the essential minerals are medium-grained, but hornblende crystals are as much as 2 cm in length, and plagioclase (An_{40-80}) and microcline are fine grained. The index of refraction of the garnet is about 1.785; evidently the garnet is rich in grossularite, as is the garnet in the calc-silicate ellipsoids in the Bugtown formation. Hornblende, diopside, and garnet are very poikilitic both in the ellipsoids and in the beds. The gneiss differs from that in the Crow formation in that the latter does not contain garnet. The richness of the gneiss in carbonate and calcium and magnesium minerals, its bedded nature, and the associated clastic sediments indicate that it was derived from impure carbonate rocks formed in relatively shallow water.

The calc-silicate ellipsoids of the Mayo formation are identical in structure and mineralogy to those in the Bugtown formation. They are very abundant about 1 mile north of the Tin Mountain mine where they are in massive quartz-mica schist that is interbedded with thin staurolite-bearing schist. They probably constitute one percent or more of the total volume of the formation, and locally form as much as 7 percent of the rock. One large ellipsoid, about 600 feet northeast of pegmatite 28 (pl. 21, C–1) is composed chiefly of hornblende and is indistinguishable in hand specimen from amphibolite. The plagioclase (An_{60}) in this ellipsoid, however, is much more calcic than that in the amphibolite bodies.

CUMMINGTONITE-QUARTZ SCHIST

Pale green cummingtonite-quartz schist occurs in prospect pits and as scattered float in the northwestern part of the quadrangle in the NE¼ sec. 34, T. 3 S., R. 3 E. This schist is very poorly exposed and for this reason it is not shown separately on plate 21. It apparently forms a single unit a few tens of feet thick that trends more or less northward parallel to bedding. It consists of thin layers of light-green cummingtonite-grunerite interlayered with sugary quartzite. A few feet of graphitic schist is at the borders of the amphibole rock.

AMPHIBOLITE

Small sills, dikes, and irregularly-shaped bodies of amphibolite are common throughout the Precambrian rocks of the Black Hills. Darton and Paige (1925, p. 4) described them and believed that they were derived from intrusive diorite. Dodge[3] made an intensive study of many amphibolite bodies in the northern Black Hills and concluded that virtually all of them are intrusive metagabbro.

[3] Dodge, T. A., 1935, Amphibolites from the Lead area: Harvard Univ. Ph. D. thesis.

In the Fourmile quadrangle (pl. 21), more than 85 separate amphibolite bodies are in the west central part of the area along the major south-southeast trending synclinal axis. Most of the bodies are thin sills although there are a few dikes and irregular discordant bodies. Just north of the quadrangle are some dikelike bodies that cross the bedding but are parallel to the well-developed axial plane schistosity of the major synclinal axis. All the amphibolite bodies have a poor foliation parallel to that in the enclosing rocks.

The amphibolite weathers to low knolls and ridges and generally is poorly exposed. The rock is dark green to black and characterized by a "salt and pepper" texture in hand specimens. The composition is generally uniform; about 50 percent of the rock is hornblende, 45 percent is plagioclase, and 3 to 8 percent is sphene and magnetite. Rock from near the contacts of some of the bodies contains as much as 90 percent fine-grained hornblende. Common accessory minerals are biotite, quartz, chlorite, garnet, arsenopyrite, and apatite. Biotite, however, may constitute as much as 50 percent of the amphibolite in a thin altered zone next to some contacts with pegmatite. Both biotite and quartz are more abundant in the outer parts of some of the amphibotite bodies.

The grain size and texture of the amphibolite is generally uniform. Most of the rock is medium-grained and has a diablastic texture. Large hornblende crystals, 2 to 8 mm long, are surrounded by a finer grained matrix of plagioclase and hornblende. Some specimens have textures superficially resembling an unmetamorphosed igneous rock, but alinement of the hornblende crystals is recognizable nearly everywhere. The mineral alinement is ordinarily developed best in the outer parts of the amphibolite bodies.

The β index of the hornblende ranges from 1.660 to 1.675. The pleochroism is Z, dark green to dark blue green; Y, pale green to dark blue green; and X, pale straw yellow to brownish green. There are no detectible differences in optical properties between the larger hornblende grains and the smaller grains of the rock matrix.

The plagioclase composition ranges from An_{20} to An_{50}, and the average is about An_{35}. Most of the grains have reverse zoning so that the centers of the grains have 10 to 15 percent less anorthite than the rims. Twinned grains are very rare, and sericitic material is only slightly formed along cleavage. Inclusions of opaque minerals are common.

The occurrence of sphene as small grains characteristically clustered around magnetite suggests that its titanium was obtained from titaniferous magnetite or ilmenite. Traces of arsenopyrite occur in some of the amphibolite, especially in the body that crosses the highway south of the Tin Mountain mine.

Quartz, biotite, and garnet, are generally in the outer edges of the amphibolite bodies, but they are disseminated throughout small sills in the eastern part of the area. Only rarely do they form as much as 10 percent of the rock although garnet forms 20 percent of the amphibolite in a few outcrops. Quartz and biotite are especially abundant in the small sills near the Tin Mountain mine. Both the biotite- and the garnet-rich amphibolites contain less hornblende and more quartz than average amphibolite. The biotite and garnet therefore formed largely from the ferromagnesian elements commonly utilized by the hornblende.

Only a few dubious relicts of original minerals or textures have been recognized in the amphibolite. Intrusive bodies 1 mile southeast of Custer contain light-colored augen, as much as 1 cm in diameter, consisting of fine-grained plagioclase in a matrix of hornblende and plagioclase. These augen are only slightly flattened and sheared parallel to the foliation, and they may well be relicts of plagioclase phenocrysts. Similar augen in small bodies of amphibolite near Pringle, southeast of the Fourmile quadrangle (fig. 75), are composed of a single twinned plagioclase grain (An_{60}) as much as 5 mm in diameter. Finer grained plagioclase of the matrix (approximately An_{30}) is reverse zoned.

The high percentage of mafic minerals in most of the amphibolite suggests that the original composition approximated diabase. A chemical analysis of typical amphibolite (table 2, JR–5–54) shows it to be very similar to the average diabase shown in the same table.

Some of the amphibolite bodies contain inclusions of the country rock. The best exposures of the inclusions are in the amphibolite along U.S. Highway 16 south of the Tin Mountain pegmatite. A body of schist large enough to be shown on the map is in the amphibolite near pegmatite 85 (pl. 21, E–2). The contacts of amphibolite with these inclusions are relatively sharp, as are the contacts with the country rock. The schist inclusions consist mainly of plagioclase, quartz, and approximately 10 percent biotite. In a few places, the quartz-mica schist adjacent to the amphibolite contains scattered needles of actinolite. However, the actinolite is not found more than a few inches from the contact with amphibolite.

The abundance of the intrusive bodies along the axis of the major syncline and the parallelism of some bodies of amphibolite to axial plane foliation indicate that the intrusions occurred after the area had undergone at least some deformation. Foliation in amphibolite, conformable with that in the country rock, shows that there was additional deformation after intrusion, and in the absence of evidence for two metamorphisms it may be presumed that this was all part of the same tectonic episode. The similarity of these amphibolite bodies to those in the northern Black Hills described by Dodge [3] suggests that they were intruded at the same time.

QUARTZ-RICH VEINS

Veins containing quartz and a variety of other minerals are widely distributed throughout the area. Their thickness ranges from a few inches to about 10 feet and their length from several feet to a few tens of feet. Only the larger ones are shown on plate 21. Most of the larger veins are crosscutting and have sharp contacts, whereas smaller podlike to irregular veins are generally subparallel to bedding and have less sharp contacts. The veins may be divided on the basis of their mineralogy into (a) quartz veins, (b) quartz-feldspar veins, (c) quartz-sillimanite veins, and (d) quartz-staurolite veins. All gradations exist among the last three types.

QUARTZ VEINS

The largest veins in the area, including all those shown on plate 21, consist almost entirely of quartz. Most of these veins trend either north-northeast or northwest, and have a nearly vertical dip. The adjacent wallrock generally has a layer, a few inches thick, containing abundant tourmaline and graphite (fig. 77). The quartz in these veins is white and massive; individual grains cannot be distinguished. Five samples of these veins near Fourmile contained from 0.1 to 0.3 oz. of gold per ton (analyst Charles Bently, South Dakota School of Mines and Technology, Rapid City, S. Dak.).

A few quartz veins of diverse attitudes and sizes have no recognizably introduced minerals along their contacts, and are almost barren of gold. A faint cleavage in the individual white quartz grains shows that the average grain size is about 0.5 inch. Though these veins otherwise resemble the gold-bearing veins, they presumably are in some respect different in origin.

QUARTZ-FELDSPAR VEINS

Quartz-feldspar veins are common in all the quartzose schist in the area but are especially abundant in the Bugtown and the lower part of the Mayo formation. Most of them are between 1 and 3 feet thick and between 10 and 50 feet long. They are approximately parallel to the schistosity of the enclosing rock but pinch and swell along strike, somewhat like the concordant vein shown in figure 77. The greater part of each vein consists of quartz that is mostly white to clear, but rarely

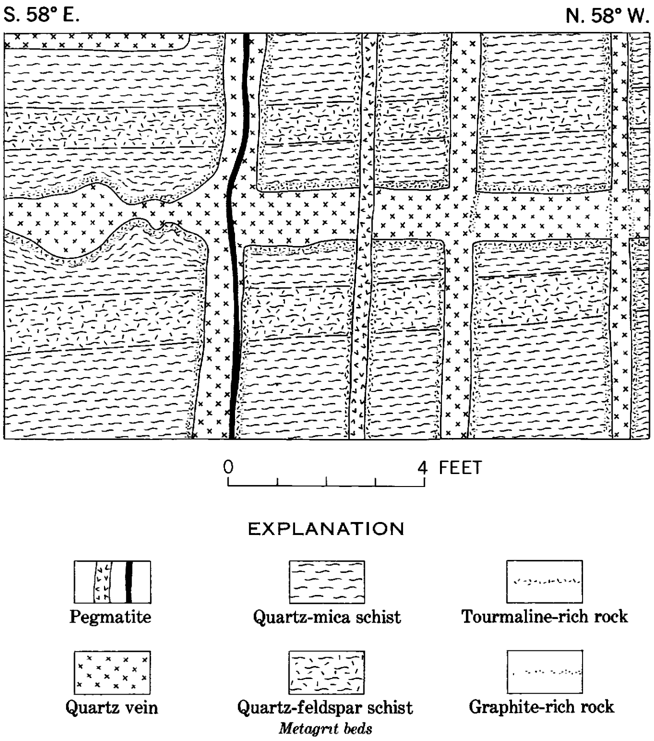

FIGURE 77.—Sketch showing relation of quartz vein to pegmatite. Introduced tourmaline and graphite are in the wallrock along the contacts of quartz veins; similarly introduced tourmaline is in the wallrock, at the pegmatite contacts. Vertical face in trench 300 feet S. 70° W. from pegmatite 84 (pl. 21).

is somewhat smoky. Plagioclase and microcline are abundant near the contacts of the veins, but they generally form less than 10 percent of an entire vein. Traces of biotite, muscovite, chlorite, tourmaline, and apatite commonly occur with feldspar, and in a few places are more abundant than the feldspar. The country rock adjacent to some of the veins has abundant biotite, garnet, tourmaline, graphite, and chlorite in a layer 1 inch to 1 foot thick. All these minerals except chlorite are also present in unaltered schist but in much smaller quantities. The layer of dark minerals in the country rock grades outward from the veins into normal quartz-mica schist.

The outer parts of some of the quartz-feldspar veins contain sufficient feldspar and muscovite to resemble thin pegmatites. The main differences between the veins and the pegmatites are: (a) large crystals are not found in the veins; (b) the composition of the plagioclase is about An_{20}, whereas in pegmatites in the area it rarely is greater than An_{15}; (c) the plagioclase in the veins has matted aggregates of subhedral crystals averaging about 0.2 inch in diameter; the plagioclase of most pegmatites is anhedral; (d) the quartz of the veins tends to be clearer than the dense white quartz in the pegmatites; and (e) the microcline in the veins is commonly not perthitic, as it is in most pegmatites of the area. No gradation has been recognized between the quartz-feldspar veins and the pegmatites.

QUARTZ-SILLIMANITE VEINS

Quartz-sillimanite veins are found only in aluminous schists in the sillimanite zone (pl. 21). They are 1 to 2 feet thick, are a few tens of feet long, and have irregular contacts with the country rock. Sillimanite is restricted to the outer parts of the veins, where it is in silky gray to white aggregates as much as 6 inches long that form no more than 10 percent of the vein material. The associated minerals are generally biotite, chlorite, tourmaline, feldspar, and apatite. One small vein about 25 feet south of the east end of pegmatite 35 (pl. 21, D–2), near the edge of the sillimanite zone, contains kyanite as well as sillimanite; the schist in the immediate vicinity contains no megascopic sillimanite or kyanite.

The quartz-sillimanite veins generally lack the well-developed layer of dark minerals found in schist adjacent to quartz-feldspar veins. In other respects however, the mineralogy and appearance suggest they are gradational with the quartz-feldspar veins.

QUARTZ-STAUROLITE VEINS

Relatively rare quartz-staurolite veins are found chiefly in association with staurolite-bearing schist. They occur approximately 0.4 mile east of the Warren Draw mine (pl. 21) and 1 mile north of the Tin Mountain mine in the southern part of the Berne quadrangle. They are only a few inches thick and several feet long. The centers of the veins consist only of white quartz and the borders have crystals of staurolite, 0.2 to 1.0 cm long, and small amounts of biotite and chlorite.

AGE AND ORIGIN OF THE VEINS

The age of the quartz veins with respect to the other Precambrian rocks can be determined by study of their structural relations with adjacent rocks and of their mineralogic relations with metamorphic mineral assemblages of the schist. None of the veins have been found in Paleozoic rocks, nor is there other evidence that any of them formed after the Precambrian.

The veins with feldspar, sillimanite, and staurolite contain minerals that are also characteristic of the associated metamorphic rocks. It seems likely that the staurolite and sillimanite in these veins formed under the conditions of temperature and pressure that caused the development of the same minerals in the schist. Furthermore, the irregular shapes of these veins and the halo of dark minerals surrounding them suggest that they formed by recrystallization of the schist accompanied by migration of ferromagnesian constituents to the edge and of silica to the center of the veins. Chapman (1950, p. 699–709) described somewhat similar veins in Vermont and presented evidence for their formation by metamorphic processes. Detailed sampling and chemical analyses are needed to prove that these veins in the Fourmile quadrangle were formed in a similar manner.

The main quartz and gold-bearing quartz veins have attitudes that are parallel to some of the joints, and, if their emplacement was controlled by these joints, they must have formed after metamorphism. These veins are cut by pegmatites, as shown in figure 77, and so preceded the pegmatites. The gold content of these quartz veins supports a hydrothermal origin.

Some of the pegmatite bodies contain thin fracture-filling units of white quartz that cannot be traced into the schistose wallrock. Three of the largest of the fracture fillings are in pegmatites 87 and 90 (pl. 21, F–2). Presumably these, like other fracture fillings in the pegmatites, were formed as a part of the crystallization history of the pegmatite.

A few veins and stringers consisting entirely of quartz are in country rock immediately beyond irregular contacts at blunt ends of some pegmatite bodies. It is possible that they formed by a process of metasomatism, such as that described by Ramberg (1956, p. 210), in which silica migrates toward a low pressure area in a crack between boudins of competent rock. The blunt ends of the pegmatite may have effectively acted as the ends of boudins. Regardless of the origin of these quartz veins, their association with pegmatites indicates they are no older than the period of pegmatite emplacement.

PEGMATITE

Approximately 2,300 separate bodies of granitic pegmatite more than 1 foot thick and 20 feet long have been mapped in the Fourmile quadrangle. They are most abundant in the eastern and southeastern parts of the quadrangle and are absent near its northwest corner. They occur in all Precambrian rock units of the area.

Numerous age determinations on uraninite and other minerals indicate that the pegmatite is Precambrian. The most recent and probably the most accurate determinations on uraninite from the Bob Ingersoll mine, Keystone, S. Dak., give an age of about 1,600 million years (Davis and others, 1955, p. 146–147; Kulp and others, 1956, p. 1557). Two zircon age determinations from the granite of Harney Peak gave 1,350 and 1,640 million years, respectively (Ahrens, 1949, p. 255). It appears therefore, that the age of both the granite of Harney Peak and the associated pegmatite is close to 1,600 million years.

The major minerals in the pegmatite are plagioclase (albite-oligoclase), quartz, perthite, and muscovite. The main accessory minerals are tourmaline, apatite, and garnet. Lithium-rich pegmatite may contain amblygonite, spodumene, and lepidolite as major constituents. Rarer accessories are beryl, lithiophilite-triphylite, loellingite, columbite-tantalite, microlite, tapiolite, sphalerite, allanite, graftonite, and uranium minerals.

The individual crystals from the pegmatite are mostly between 0.1 inch and 1 foot in size, but some are smaller than 0.1 inch and a few are larger than 10 feet across. The terminology of Cameron and others (1949, p. 16) for pegmatitic textures is used throughout this report; the additional term "very fine grained" is used here for pegmatite with an average grain size of less than a quarter of an inch. The other textures are: (a) fine, ¼ to 1 inch; (b) medium, 1 to 4 inches; (c) coarse, 4 to 12 inches; and (d) very coarse, greater than 12 inches.

The pegmatite is here classified into three categories: (a) layered, (b) homogeneous, and (c) zoned. The distinctive feature of the layered pegmatite is the presence of alternating layers, each of which differs from its neighbors in composition or texture; the layers are lenticular to tabular in form and are generally parallel to the pegmatite-wallrock contact. The homogeneous pegmatite has almost the same composition and texture throughout each body, though there ordinarily is a fine-grained border phase of insignificant size. The zoned pegmatite contains zones of differing compositions and texture, as described by Cameron and others (1949, p. 16–24). Though zoned pegmatite bodies are much less abundant than the other varieties, they have received by far the greater share of attention by geologists because of the presence in them of the rarer minerals and of minable deposits of various other minerals.

DISTRIBUTION

The Fourmile quadrangle is in the southwest part of the area of abundant pegmatite surrounding Harney Peak. Figure 78 indicates by isopleths the number of pegmatite bodies per square mile at any single place throughout the southern Black Hills. This map was made partly from geologic maps and partly by counting pegmatites visible in aerial photographs.

Figure 78 shows that the Fourmile quadrangle is just west of a major concentration of pegmatites that extends southwest from the main granitic area around Harney Peak. The scarcity of pegmatites on the other sides of the Harney Peak body suggests that the main intrusion plunges beneath the east side of the Fourmile quadrangle.

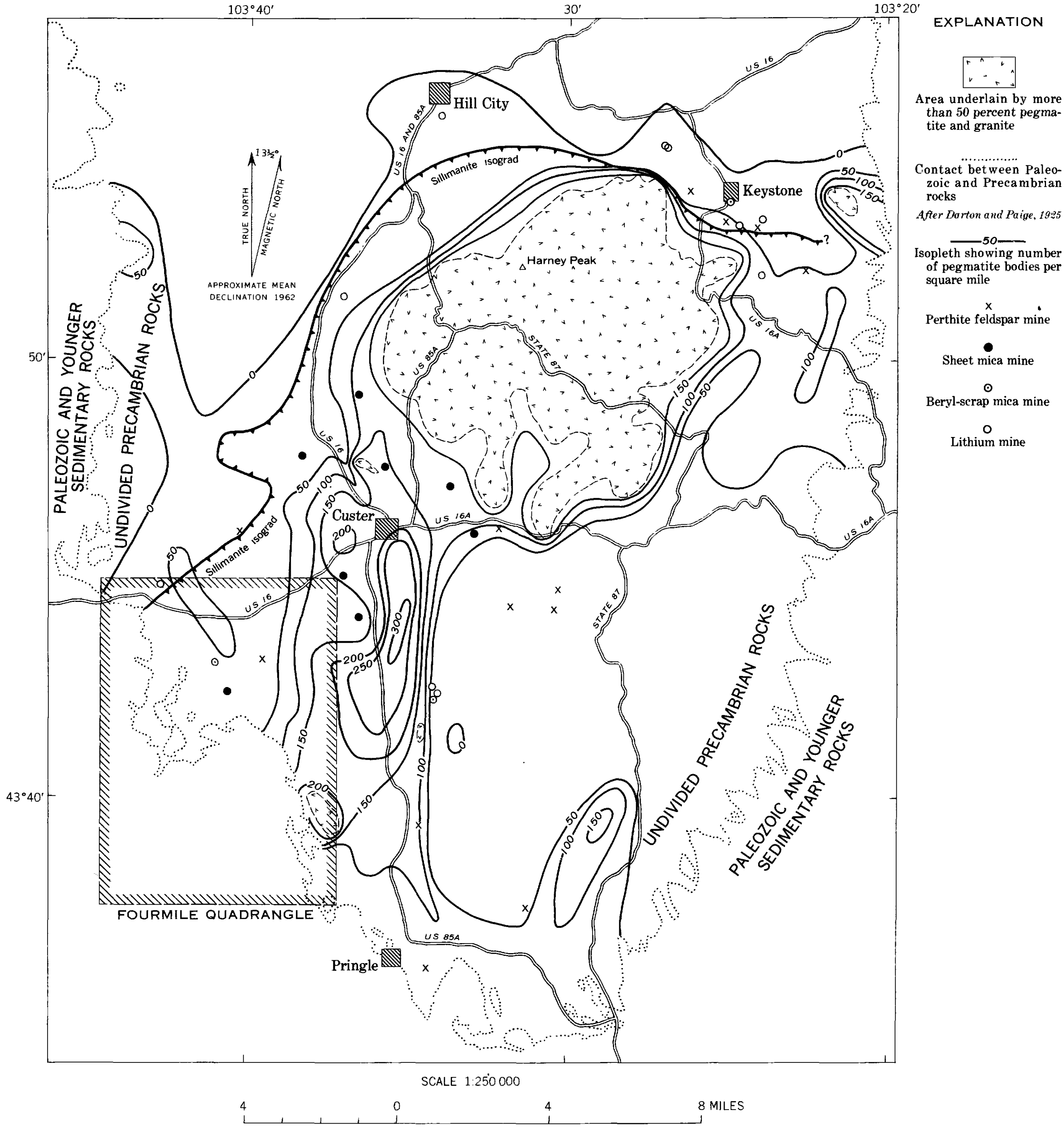

FIGURE 78.—Map showing distribution of pegmatite in the southern Black Hills, South Dakota.

Within the Fourmile quadrangle the abundance of pegmatite is shown on figure 79, which was constructed by calculating the amount of pegmatite exposed in each quarter of a square mile and drawing isopleths indicating the percentage of the rock that consists of pegmatite. Though the isopleths based on the quantity of rock that is pegmatite in the area shown on plate 2 differ somewhat from isopleths based on the number of pegmatite bodies per square mile shown in figure 78, the gross pattern in the 2 illustrations is the same. The quantity of pegmatite increases greatly from the northwestern to the eastern and southeastern part of the quadrangle. No pegmatite is exposed in the northwestern part of the quadrangle and very few are in the

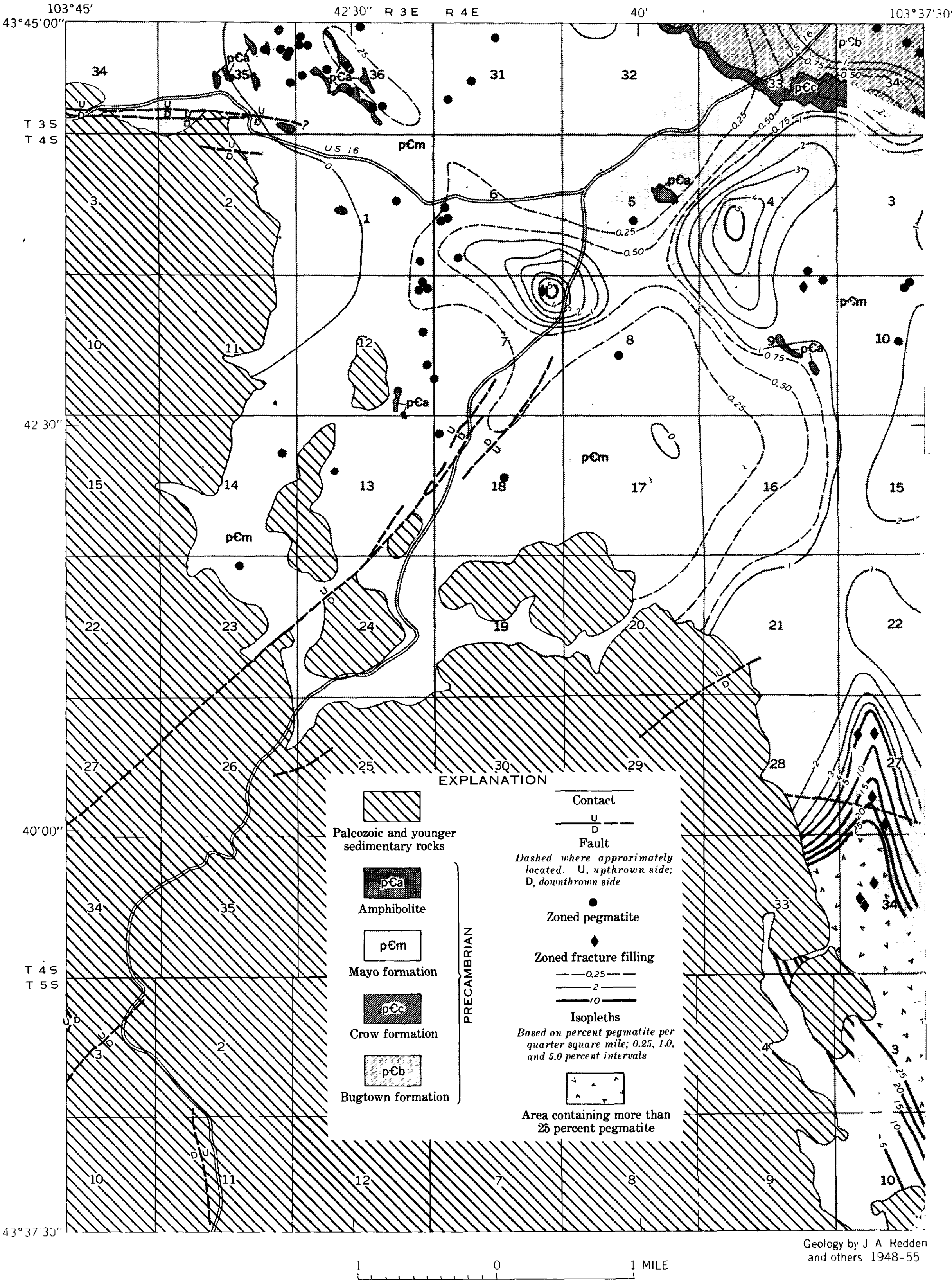

FIGURE 79.—Isopleth map showing approximate distribution of pegmatite in the Fourmile quadrangle, South Dakota.

western half. The distribution of areas of high and low pegmatite concentration on figure 79 shows no correlation with the composition of the country rock. In a broad sense this lack of correlation is true of areas to the east and north of the quadrangle, but locally certain metamorphic rocks are less favorable host rocks than others. The sole example of an unfavorable host rock in the Fourmile quadrangle is the Crow formation, which contains only one pegmatite in this and adjacent quadrangles.

SIZE, SHAPE, AND ATTITUDE

The pegmatite bodies vary greatly in size, shape, and attitude. The smaller ones are only a few inches thick and a few feet long; the larger ones are as much as 4,400 feet long and 300 feet thick (pl. 21, pegmatite 159, G–6); however, about 95 percent are less than 10 feet thick and very few are more than 20 feet thick. In general, the irregular bodies are the largest, and the sill-like bodies the smallest. Many of the larger bodies shown on plate 21 are exposed on dip slopes and are actually less than half as thick as the width of outcrop. Pegmatite 118 (pl. 21, F–3) is only about 25 feet thick, yet the outcrop width exceeds 100 feet.

The contacts on the opposite sides of many of the thicker bodies have dips that indicate they converge at depth and the pegmatite becomes thinner at depth. At the New York pegmatite (pl. 21, D–4), mining on the 300-foot level showed that the body was approximately one-third as thick as at the surface (Norton, J. J., *in* Page and others, 1953, p. 164–167). Diamond drill holes at several of the pegmatite mines prove a decrease in thickness at depth, although rarely the pegmatite bodies thicken at shallow depths and then thin at greater depths. It is likely that the smaller bodies bottom at shallow depths. The depths to which long sill- and dike-like pegmatites such as pegmatites 90 and 166 (pl. 21, F–2 and G–7) extend is presumably great because their contacts commonly are nearly parallel. Relatively long (2,000–3,000 feet) spodumene-bearing pegmatites near Kings Mountain, N.C., are known to extend to at least 900 feet in depth (Griffitts, 1954, p. 2).

The pegmatite bodies vary in shape from thin sills to irregular discordant bodies (fig. 80). Approximately 70 percent are thin lenticular sills; 25 percent are tabular or irregularly shaped dikes; and 5 percent are thickly tabular, oval, or teardrop shaped and in part discordant. Most of the pegmatite bodies, being concordant, strike north-northwest, parallel to the bedding and predominant schistosity, and dip southwest. Some are separated by layers or septa of country rock from a few inches to a few feet thick. These were mapped as separate bodies where possible, but where alternating thin layers of country rock and pegmatite are abundant they have been shown as single bodies of pegmatite. Several bodies near pegmatite 134 (pl. 21, G–3) in the east part of the quadrangle are of this type. They are apparently similar to the multiple pegmatites in the Crystal Mountain district, Colorado, described by Thurston (1955, p. 31). However, there is no field evidence that they were emplaced at different times, although a few examples, where one pegmatite body is in contact with another of slightly different composition, suggest multiple intrusion. Septa of country rock in some pegmatites in the Fourmile area end along strike and the two limbs merge into single bodies of pegmatite. Other wallrock septa, where there is enough vertical exposure, are seen to pinch out down dip. Although most of the pegmatites showing these structures are concordant, the Castle Rock pegmatite is a discordant body that has the same features, as shown by figure 81. Vertical exposures indicate that some and perhaps all of the schist septa shown in figure 81 pinch out downward. At a higher erosion surface this pegmatite might have appeared as two or more bodies of the same composition. It is likely, therefore, that some of the parallel bodies which are separated by thin layers of country rock and which appear to be unconnected, are actually parts of a single larger mass in the third dimension and that the thin layers of schist are roof pendants. Rhyolite dikes in the Homestake mine in the northern Black Hills exhibit similar relations. They are single bodies at depth, but nearer the surface they separate into parallel, concordant lenses (Noble and others, 1949, p. 344).

Discordant pegmatites commonly are more complex in outline than concordant ones; furthermore, as might be expected, the larger the pegmatite mass, the more likely it is to be discordant. Figure 80 and plate 21 illustrate the fact that there are 2 main discordant trends; one is about N. 60° E. and the other is about N. 10° E. These trends are even more apparent in figure 82 where the poles of planes determined by the average strike and dip of 235 discordant bodies are plotted on a Schmidt net. The relatively few northeast-trending dikes dip steeply southeast and the abundant north-northeast ones are nearly vertical. The more north-trending dikes are concentrated in the north-central part of the quadrangle, whereas those with more easterly trends are most abundant in the east-central part. The north-northeast trend is apparent in examples *B*, *D*, *G*, and *H* of figure 80; the more easterly trend is apparent in examples *C*, *F*, and *H*. Pegmatite 142 (*H* of fig. 80) is located between the 2 areas of different trends and contains both north- and the more

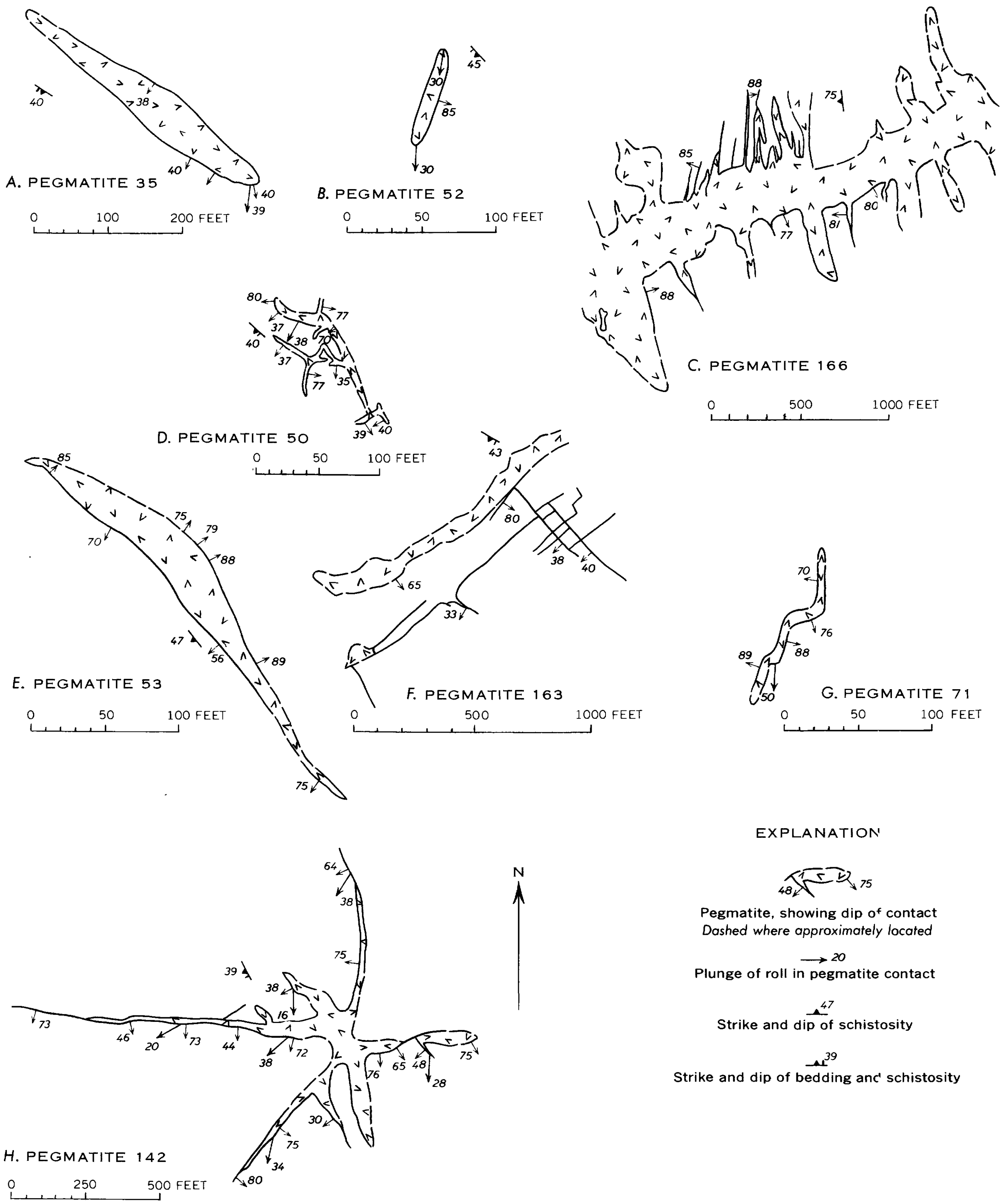

FIGURE 80.—Sketch showing shapes and attitudes of pegmatites, Fourmile quadrangle, Custer County, S. Dak.

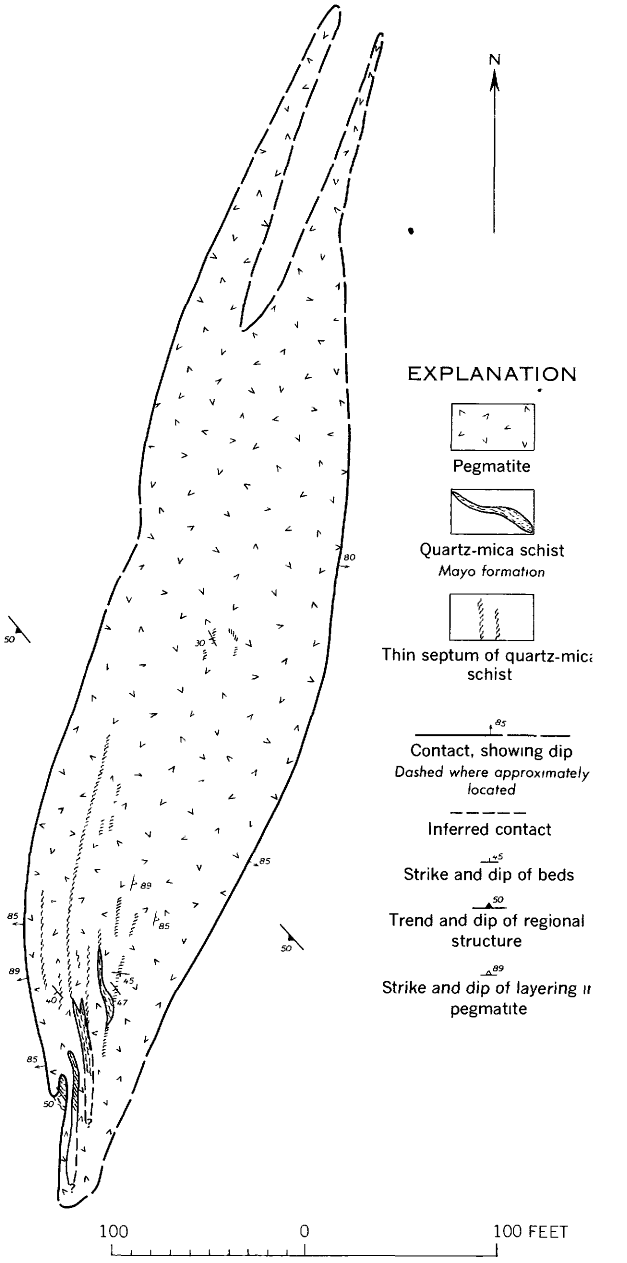

FIGURE 81.—Sketch map showing orientation and distribution of schist septa in the Castle Rock pegmatite.

east-trending segments as well as northwest-trending segments that are concordant with the country rocks. The large irregular pegmatite 166 (pl. 21; fig. 80 *C*) is a typical discordant body with a small limblike apophyses parallel to the main schistosity. A few pegmatite dikes west of Pleasant Valley trend northwest and dip steeply northeast. Others in this same area trend north to slightly northeast and have "right-hand" jogs such as that in pegmatite 71 (pl. 21; fig. 80 *G*). The short limbs of these jogs are approximately parallel to

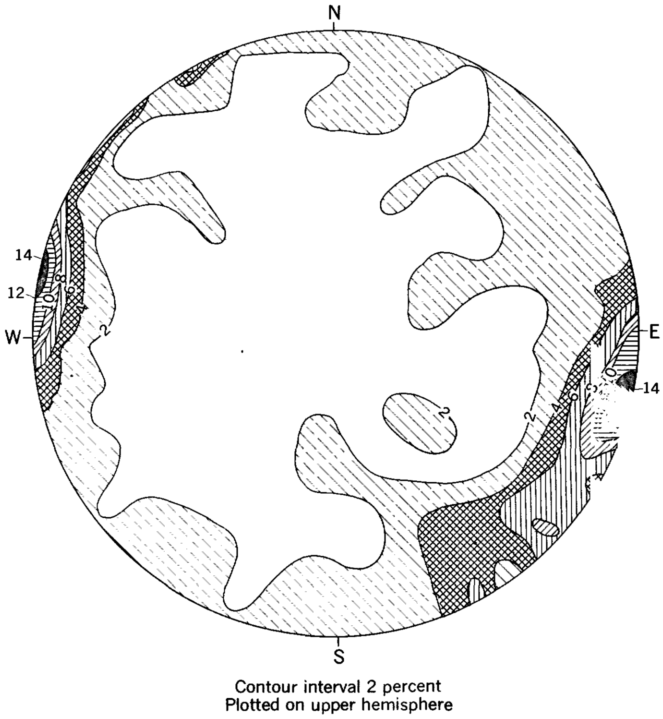

FIGURE 82.—Contour diagram of poles of 235 discordant pegmatites in the Fourmile quadrangle.

bedding and the other limbs are along a well-formed joint set. The short limb may be only a few feet long and some cannot be shown at the scale of plate 21. There are no examples of the reverse or "left-hand" jog.

The Corky, Helen Beryl, and New York pegmatites (pl. 21) are oval in plan, and are shaped like an inverted teardrop in cross section. This is a fairly common shape in the pegmatites that have been mined in the southern Black Hills.

The size, shape, and attitude of the pegmatite bodies can be correlated to some degree with the type of host rock. Approximately 95 percent of the pegmatites in the Bugtown formation are small, tabular, and concordant. The same can be said for the pegmatites in the lower part of the Mayo formation, but more than half the pegmatites in the middle part of the exposed Mayo, are discordant, and very large irregular bodies are common. The country rock in the Bugtown and in the lower part of the Mayo formation is uniform schist of moderate competency, but higher in the Mayo formation there are competent beds of feldspathic schist and grit interbedded with less competent schist. These competent beds tended to break across the foliation, and the pegmatites follow these fractures. The pegmatites within bodies of amphibolite occur as dikes that strike generally parallel to the foliation of the amphibolite but dip to the northeast at an angle to the foliation.

INTERNAL STRUCTURES

Cameron and his associates (1949, p. 10–13) and Jahns (1955, p. 1039–46) summarized the development of concepts regarding zones and other internal structures in pegmatite bodies. These structures were first classified into zones, fracture fillings, and replacement bodies by Cameron and others (1949, p. 14) after a detailed study of many pegmatite mines. All these structures appear in pegmatites of the Fourmile quadrangle, and many of the unzoned pegmatites have also a layered form that will here be described as a separate structure.

ZONES

"Zones" as originally defined (Cameron and others, 1949, p. 14) are mappable units of distinctly different mineralogy or texture, or both; where ideally formed, they reflect the shape of the pegmatite body. They may be divided into four types:

1. Border zones or outermost zones.
2. Wall zones.
3. Intermediate zones.
4. Cores or innermost zones.

In a single pegmatite body the zones may be designated inward from the outer contact as zone 1, zone 2, and so forth.

Ideally, the zones contain specific mineral assemblages arranged in a definite sequence. The sequence of such assemblages in the zoned pegmatites in the Black Hills, compiled by Page and others (1953, p. 66), is presented in abbreviated form in table 6. This sequence is about the same as that in other major pegmatite districts of the United States (Cameron and others, 1949, p. 61). In general in the pegmatites in the Fourmile area, outer zones are rich in plagioclase, quartz, and muscovite (assemblages 1 and 2); intermediate zones are rich in perthite, quartz, and plagioclase (assemblages 1 and 2); and cores are rich in quartz and perthite (assemblages 4 and 12). A few pegmatites contain intermediate zones or cores having lithium minerals (assemblages 5 and 7–9). Assemblages 10 and 11 occur only at the New York mine in small units that were not mapped separately by Page and others (1953, p. 163–169).

TABLE 6.—*Sequence of mineral assemblages in zoned pegmatites, Black Hills, S. Dak.*

[Condensed from Page and others (1953, p. 16, table 3)]

Assemblage	Essential minerals composing more than 5 percent of each zone	Zonal distribution			
		Border	Wall	Intermediate	Core
1	Muscovite, quartz, plagioclase	×	×	×	----
2	Quartz, plagioclase	×	×	×	×
3	Perthite, quartz, muscovite, plagioclase	×	×	×	×
4	Perthite, quartz	----	----	×	×
5	Perthite, quartz. plagioclase (usually cleavelandite), amblygonite, spodumene.	----	----	×	×
6	Plagioclase (cleavelandite), quartz	----	----	×	×
7	Plagioclase, quartz, spodumene	----	----	×	×
8	Quartz, spodumene	----	----	×	×
9	Plagioclase, quartz, lepidolite	----	----	----	×
10	Quartz, microcline	----	----	×	×
11	Microcline, plagioclase, lithia-mica	----	----	----	×
12	Quartz	----	----	----	×

BORDER ZONES

Border zones can be recognized at all contacts of pegmatites with metamorphic rocks in the Fourmile quadrangle. They make up a very small part of the total quantity of pegmatite because they are nearly everywhere less than 6 inches thick, and some are as thin as one-half inch. A 1-inch-thick border zone is shown in figure 88.

Border zones consist mainly of quartz, muscovite, and plagioclase, and the common accessory minerals are apatite, tourmaline, perthite, biotite, beryl, and garnet (table 7). All these minerals are very fine grained, and most are less than an eighth of an inch across. Though many of the accessory minerals are subhedral to euhedral, the major minerals are ordinarily anhedral to subhedral.

The mineralogy of border zones varies somewhat—apparently in response to the mineralogy of the wall-rock. Border zones adjacent to mica schist are rich in quartz and mica and have a relatively low plagioclase content. On the other hand border zones in pegmatites that are adjacent to amphibolite are exceptionally rich in plagioclase and apatite, and the plagioclase in such units is rich in calcium. Pegmatite 2? (pl. 21, C–1) contains oligoclase-andesine (An_{30}) immediately adjacent to contacts with amphibolite, but a few inches inward from the contact the plagioclase is albite (An_8), and where this same pegmatite crosses quartz-biotite schist, the border zone contains albite throughout.

WALL ZONES

Wall zones are found only in the zoned pegmatites shown on plate 21. These zones commonly are much thicker (1 to 10 feet) than the border zones. Though most of them are fine grained, they are coarser than the adjacent border zones, and the grain size generally increases from the outer to the inner part of the wall zone. A few wall zones are medium grained, and some of the minerals may be coarse grained. Table 7 shows that plagioclase (An_{6-12}) is commonly the most abundant mineral; most of it is blocky in form, but rarely it is the platy variety, cleavelandite. Quartz and perthite are generally the second and third most abundant minerals. Nearly all the wall zones have more than 5 percent muscovite, and in many pegmatites the high muscovite content of the wall zone is the main difference between it and the first intermediate zone. Accessory minerals

TABLE 7.—*Mineralogy of the principal zoned pegmatites of the Fourmile quadrangle*

Wallrock type: A, amphibolite; P, pegmatite; QBS, quartz-biotite schist; QMS, quartz-mica schist; QMSS, quartz-mica-sillimanite schist; QMSTS, quartz-mica staurolite schist.
Relation to wallrock: C, concordant; D, discordant.

Shape: Ir, irregular; L, lenticular; Ov, oval; T, tabular.
Internal structure: BZ, border zone; C, core; FF, fracture filling; IZ, intermediate zone; WZ, wall zone.
Mineralogy: Cb-Ta, columbite-tantalite; M, massive.

Pegmatite (pl. 21)	Wallrock type	Pegmatite: Relation to wallrock	Shape	Internal structure	Mineralogy: Plagioclase Volume (percent)	Plagioclase Size (in.)	Perthite Volume (percent)	Perthite Size (in.)	Quartz Volume (percent)	Quartz Size (in.)	Muscovite Percent	Muscovite Size (in.)	Garnet Percent	Garnet Size (in.)	Tourmaline Percent	Tourmaline Size (in.)	Lithium minerals: Mineral	Lithium Percent	Lithium Size (in.)	Other minerals: Mineral	Other Percent	Other Size (in.)
1 (Tin Mountain).[1]	AQMS	D	Ir	BZ	17	0.1	Tr.	0.2	44	0.1	39	0.1								Apatite Beryl Cassiterite	Tr. Tr. Tr.	0.1 .1 .1
				WZ	[2] 47	1	2	6	35	1	12	2			Tr.	2	Spodumene Lithia mica Amblygonite	4 Tr. Tr.	6 .1 3	do Beryl Apatite	Tr. .8 Tr.	.1 4 .2
				1st IZ	[2] 10	1	50	30	33	3	7	4			Tr.	4	Lithiophilite Spodumene	Tr. Tr.	6 10	Cassiterite	Tr.	.3
				2nd IZ	[2] 2	1	89	36	6	8	Tr.	4					do Amblygonite	1 Tr.	30 12	Beryl Cb-Ta	.4 Tr.	8 1
				C	6	1			55	M	1	6					Lithiophilite Spodumene Lepidolite	1 23 14	8 36 .1	Apatite do Beryl Cassiterite Cb-Ta Pollucite Microlite	Tr. Tr. .5 Tr. Tr. Tr. Tr.	1 1 10 .5 1 2 .5
				FF	2	1			55	4	5	3					Spodumene	37	6			
3	A	D	T	WZ	70	.5			10	1	20	1.5			Tr.	.5						
				C	10	.7			80	10	10	.7										
5	QMS	D	T	WZ	45	.5	20	4	30	.5	5	2					Lithiophilite	Tr.	2	Beryl	.1	2
				C			65	8	30	2										do	.1	2
8	A	D	T	BZ					60	.2	40	.1			Tr.	.1				Apatite	Tr.	.05
				WZ	45	1			39	.7	14	2			2	1				do	Tr.	.4
				C	10	1			85	1	4	2										
9 (Lithia Lode).	A	D	T	WZ	55	1	4	2	25	1	14	2			Tr.	.8	Amblygonite Lithiophilite	Tr. Tr.	3 1	Beryl	.5	2
				C	40	2	10	6	38	5	10	3					Amblygonite Lepidolite	1 Tr.	8 .1	Beryl	.5	2
10	QMS, A.	D	T	WZ	50	.5			38	2	12	3								Apatite Beryl Tapiolite	Tr. Tr. Tr.	.1 2 .2
				C	[2] 3	2			95	14	Tr.	3										
11	QMS, A.	D	T	WZ	[2] 70	2			10	.7	20	3								Apatite	Tr.	.1
				C	[2] 85	4			5	1										do	Tr.	.1
12 (Warren Draw).	QMS, A.	D	T	WZ	43	.5	6	3	41	.5	10	1	Tr.		2	.8				Garnet Beryl	Tr. .1	.3 1.5
				WZ	60	.2			30	.2	10	1	Tr.	.1	Tr.	.1	Amblygonite Spodumene	Tr. Tr.	1 1	Garnet Beryl	Tr. Tr.	.1 .2
				C	20	.5	50	10	27	1	2	4					Lithiophilite	Tr.	4	do Apatite	Tr. Tr.	3 1
				FF	4	1	4	6	70		5	1					Amblygonite Spodumene	8 10		Beryl	1	2
19	A, QMS.	D	L	WZ	50	.5			35	.8	15	1.0										
				C	3	.8	45		50	6.0	1	1.5								Beryl	Tr.	1.0
20	A	D	T	WZ	63	.5			26	.5	10	.2								Beryl	.1	1
				C	35	2	18	5	44	2	2	2								do	Tr.	2
22 (Wright Mica Lode).	QMS	C	Ov	BZ	5	0.1			60	0.1	35	0.1								Apatite	Tr.	0.1
				WZ	35	.8	17	3	30	1	15	3			3	.7				Beryl	.5	3
				IZ	3	1	75	20	20	3	2	6			Tr.	.8				do	.5	6
				C	10	2	10	30	80	M	Tr.	6								do	.1	3
24	QMS, A.	D	T	WZ	[2] 70	2			10	.7	20	3								Apatite	Tr.	.3
				C	[2] 95	4			5	1										Beryl	Tr.	1
27	A, QMS	D	T	WZ	42	.2	15	1	37	.3	5	.8			1	.3						
				C	15	.6	50	10	35	.8										Beryl	Tr.	1.5
28	QMS, A	D	T	WZ	42	.2	12	1	35	.8	6	1			2	1	Lithiophilite	Tr.	6	Pyrite Beryl	Tr. .3	.1 2
				C	15	.5	15	10	65	1	4	1			Tr.	1				Apatite Beryl	Tr. .3	.1 2
33 (Royal Flush Claim).	QMS, A.	D	T	BZ	45	.1			40	.1	15	.1			Tr.	.1				Apatite	Tr.	.1
				WZ	40	1	5	2	45	1	5	2			4	1	Lithiophilite Amblygonite	Tr. Tr.	8 2	Beryl Cb-Ta	.1 Tr.	3 .2
				C	3	2	70	18	25	M	Tr.	2			Tr.	2				Beryl	.2	5

See footnotes at end of table.

TABLE 7.—*Mineralogy of the principal zoned pegmatites of the Fourmile quadrangle*—Continued

Pegmatite (pl. 21)	Wallrock type	Pegmatite: Relation to wallrock	Shape	Internal structure	Mineralogy: Plagioclase Volume (percent)	Plagioclase Size (in.)	Perthite Volume (percent)	Perthite Size (in.)	Quartz Volume Percent	Quartz Size (in.)	Muscovite Percent	Muscovite Size (in.)	Garnet Percent	Garnet Size (in.)	Tourmaline Percent	Tourmaline Size (in.)	Lithium minerals: Mineral	Lithium Percent	Lithium Size (in.)	Other minerals: Mineral	Other Percent	Other Size (in.)
(Gayle Claim).	QMS, A	D	T	WZ	42	.7	8	4	42	1	5	1	Tr.	0.5	3	1				Beryl	.1	1
																				Cb-Ta	Tr.	.2
				C	8	2	21	16	60	M	4	3					Amblygonite	Tr.	3	Beryl	.2	6
36	QMS	C	T	WZ	40	.6	6	2	40	.9	8	1.5	Tr.	.2	6	.8				Beryl	Tr.	.5
				C	20	1.0	38	5	38	1.0	2	4.0	Tr.	1.0	2	1.0						
41 (Pleasant Valley).	P	D	T	WZ	40	.8	10	6	35	1	11	2			3	1	Amblygonite	1	4	Beryl	.5	2
																				Cb-Ta	Tr.	.2
				C	Tr.	1.5	53	20	44	8	1	5								Beryl	Tr.	4
																				Sphalerite	Tr.	.1
46 (Rainbow No. 4).	QMS	D	Ov	BZ	25	.2			50	.3	25	.3			Tr.	.4				Apatite	Tr.	.1
				WZ	43	.7	10	6	36	1	5	2	Tr.	1	4	1				Beryl	.3	2
				C	[2] 10	1	45	30	40	12	4	6					Lithiophilite	Tr.	4	do	.5	6
																	Purpurite	Tr.	.5			
48	QBS	D	T	WZ	40	.6	18	7	25	.7	15	2			2	.8				Beryl	.2	2
				C	15	1	50	12	34	1	Tr.	4										
49 (Rainbow No. 3).	QMS	D	Ir	WZ	46	.5			42	.7	8	1			3	.7				Apatite	Tr.	.1
																				Beryl	Tr.	1
				C	16	.8	40	16	40	4	3	3			1	1	Lithiophilite	1	5	do	.7	3
53 (Volcano)	QMS	D	T	WZ	45	.7	4	4	44	1	4	1.5			1	.7				Beryl	.5	3
				IZ	17	1	45	12	45	2	Tr.	1.5			Tr.							
				C	38	2	15	18	40	M	5	3					Amblygonite		4			
																	Lithiophilite		8			
58 (Corky)	QMS	D	T	WZ	45	.8	4	4	40	1	10	1			1	.6				Apatite	Tr.	.1
																				Beryl	Tr.	1
				C	5	.7	45	24	43	12	4	3			Tr.	1	Amblygonite	Tr.	12	do	.8	6
64 (Big Tom)	QMSTS, QMSS	D	T	WZ	40	1			40	1	20	1								do	.3	1.5
				1st IZ	55	3			20	2	23	1.5								do	.3	2
				2nd IZ	55	3			45	3										do	Tr.	2
				C	Tr.	3			95	14	Tr.	.5										
66 (Helen Beryl).[1]	QBS, QMS.	C	Ov	WZ	43		4		33		13				Tr.		Spodumene	Tr.		Biotite	5	
																				Beryl	.5	1.0
																				Apatite	Tr.	
				1st IZ	5		52		22		10				Tr.					Beryl	.5	
																				Apatite	Tr.	
				2d IZ	5		59		26		10				Tr.		Spodumene	Tr.		Beryl	.03	
																				Biotite		
				3rd IZ	2		65		10		3						Spodumene	20		Beryl	Tr.	
																	Lithiophilite	Tr.				
				C	[2] 10		11		54		8						Spodumene	15		Beryl	1.2	
																				Cb-Ta	.1	
																				Cassiterite	Tr.	
68 (Michaud Beryl).	QBS	D	T	WZ	34	.8	10	3	48	1	6	1			2	.6				Apatite	Tr.	4
				C	[2] 15	2	40	10	42	M					Tr.	2	Amblygonite	Tr.	2	Beryl	1.5	5
																				CB-Ta	Tr.	.5
70 (Red Spar)	QMS	C	T	WZ	42	.9	3	4	44	1	7	3	Tr.	.5	2	3				Apatite	Tr.	.3
																				Beryl	.5	2
				C	5	1	36	16	55	12	3	4	Tr.	1			Lithiophilite	Tr.	6	do	.5	5
																				Cb-Ta	Tr.	.5
74 (New York).	QMS	C	L	BZ	30	.5			50	.2	20	.2			Tr.	.2				Apatite	Tr.	.1
				WZ	35	1	Tr.	10	47	2	15	6			2	2				Beryl	.3	3
				1st IZ	4	1	50	30	43	3							Amblygonite	Tr.	4	do	Tr.	4
																	Spodumene	Tr.	8	Cb-Ta	Tr.	.5
				2nd IZ	40	2			57	1.5	Tr.	1					do	Tr.	8	Beryl	Tr.	3
																	Lepidolite(?)	Tr.	.1			
				3rd IZ	[2] 70	3			29	2	Tr.	3					do	Tr.	.1	Beryl	Tr.	2
																				CB-Ta	Tr.	.8
				C	[2] Tr.	1			98	M	Tr.	1					do	Tr.	.1			
79 (Tip Top)	QBS	C	Ov	WZ	41	1	15	10	40	2	3	3			1	2				Biotite	Tr.	4
																				Apatite	Tr.	.6
																				Beryl	Tr.	2
				C	2	1	40	30	37	3	Tr.	6			Tr.	5	Amblygonite	Tr.	10	do	Tr.	4
																	Lithiophilite	Tr.	10	Lollingite	Tr.	2
																				Phosphates(?)	Tr.	2
81 (Lincoln No. 4).	QBS	D	Ov	WZ	15	.5			75	1	10	.5			1	1				Apatite	Tr.	.2
				C	4	.5	50	12	41	M	3	.5	Tr.	.5	Tr.	1.5						
93	QMS	C	T	WZ	16	.5			65	.7	16	1			3	.6						
				C	8	.5	43	7	46	M	3	1.5										

See footnotes at end of table.

TABLE 7.—*Mineralogy of the principal zoned pegmatites of the Fourmile quadrangle*—Continued

Pegmatite (pl. 21)	Wallrock type	Pegmatite																				
		Relation to wallrock	Shape	Internal structure	Mineralogy																	
					Plagioclase		Perthite		Quartz		Muscovite		Garnet		Tourmaline		Lithium minerals			Other minerals		
					Volume (percent)	Size (in.)	Volume (percent)	Size (in.)	Volume (percent)	Size (in.)	Percent	Size (in.)	Percent	Size (in.)	Percent	Size (in.)	Mineral	Percent	Size (in.)	Mineral	Percent	Size (in.)
99 (White Spar).	QMSS	C	Ov	WZ	43	2	4	12	41	2.0	10	3			2	2				Beryl Apatite	Tr. Tr.	4 .3
				IZ	12	3	40	24	45	3.0	1	3			1	2				Apatite Beryl	Tr. Tr.	.6 6.0
				C			5	24	95	M												
106	QMSS	C	O	WZ	40	.6	4	2	44	.9	10	1			2	.5				Apatite	Tr.	.1
				C	3	.8	55	10	40	1.0	1	2			1	.9						
113 (Rock Ridge).	QMS	D	T	WZ	44	2	10	12	40	.5	2	1			1	2	Lithiophilite	Tr.	8	Apatite	Tr.	.5
				C	7	4	50	30	38	M	4	3			Tr.	2	Spodumene	Tr.	6	Beryl	Tr.	6
114	QBS	D	Ov	WZ	45	1.5	10	3	43	1	1	.5			Tr.	.5				Apatite	Tr.	.2
				C	9	3	50	14	40	M	1	1.5			Tr.	1						
116	P	D	T	WZ	68	1.5			10	.7	20	1.5			Tr.	.5	Lithiophilite	Tr.	2	Apatite	Tr.	.2
				C	4	2.5	32	14	60	M	3	2.5			Tr.	.8						
126	QMS	CD	L	WZ	45	.8	8	4	41	.9	16	1	Tr.	.2	<1	.6						
				C			20	10	75	M	5	2								Beryl	Tr.	1.0
137 (MacArthur).	QMS, QBS	C	T	WZ	41	1.0	3	6	40	1.0	15	2	Tr.	0.2	1	1.0				Beryl	Tr.	3
				C	Tr.	1.0	52	12	46	2.0	1	2			1	1.0						
139	QMS	C	T	WZ	45	.8	5	3	42	.8	5	1	Tr.	.2	3	.6						
				C	10	.6	45	10	43	.8	1	1.5			Tr.	.8						
160	P	D	T	WZ	15	.9	15	12	50	1.5	15	3.0	Tr.	1.0	Tr.	1.0				Biotite	5	2.0
				C	5	1	50	24	45	M	Tr.	4.0										
161 (Punch)	P	D	T	WZ	10	.8			10	1.0	90	1.5			1	1.0				Beryl Biotite	Tr. Tr.	3.0 4.0
				C					100	M												
164	P	D	T	WZ	20	1.0	10	12	55	2.0	5	3	Tr.	1.0	Tr.	1.0				Biotite	10	5
				C	7	1.0			90	M	3	5										

[1] Mineralogy from M. H. Staatz, L. R. Page, H. G. Stephens, and J. J. Norton (written communication, 1956).
[2] Cleavelandite.

may include tourmaline, apatite, beryl, lithiophilite-triphylite, amblygonite, columbite-tantalite, biotite, cassiterite, or garnet. Beryl, lithiophilite-triphylite, and columbite-tantalite characteristically occur along the inner part of the wall zone and the outer part of the adjacent intermediate zone or core. Muscovite books and prismatic crystals of tourmaline and beryl are oriented subperpendicular to the outer contact of the pegmatite. Tourmaline and beryl crystals are commonly tapered and increase in size away from the contact.

INTERMEDIATE ZONES

Intermediate zones occur in only about 15 percent of the zoned pegmatites of the area. The thickness of the intermediate zones is highly variable. Some of the intermediate zones are so thin that they are best mapped as part of the wall zone or the core. In small bodies, such as pegmatites 7, 8, and 9 (pl. 21, C–1), they are 1 to 2 feet thick; in the larger bodies, such as the Tin Mountain pegmatite, they are as much as 30 feet thick.

Intermediate zones commonly have medium- to coarse-grained pegmatite textures, but a few are very coarse grained. Perthite crystals as much as 10 feet long are found in several of the intermediate zones of large zoned pegmatites. The essential minerals of most intermediate zones are perthite, quartz, and plagioclase (assemblages 3 or 4 of the general sequence in table 6). Accessory minerals include muscovite, beryl, tourmaline, columbite-tantalite, cassiterite, lithiophilite, and apatite (table 7). In several pegmatites, such as the Big Tom and the New York, the first (outer) intermediate zone consists of quartz and plagioclase, and the second (inner) intermediate zone consists of cleavelandite and quartz. Traces of amblygonite and spodumene in the cleavelandite-quartz zone indicate a gradation between assemblage 6 and assemblages 5 and 7.

The perthite-rich intermediate zones in several of the larger zoned pegmatites are hood-shaped and pinch out or thin at depth on each side of the pegmatite body. These units ordinarily extend farther down the hanging wall side than down the footwall side. In general, the perthite content is greatest near the crest of the zone where the rock may consist almost entirely of perthite. As the zone thins downward, quartz and plagioclase increase.

CORES

The innermost zones or cores are generally coarser grained than the outer zones, and range in thickness from 1 foot in thin bodies to as much as 40 feet in the thicker ones. Grain sizes are coarse to very coarse. The crystals of individual minerals may be of giant size where the core is sufficiently thick to accommodate large crystals. Spodumene crystals as much as 15 feet long and 3 feet thick were exposed in the Tin Mountain pegmatite in 1955. Plagioclase, however, is mainly fine to medium grained, but a few pegmatites like the Big Tom contain rosettes of cleavelandite 12 inches or more in diameter.

Quartz and perthite are the most abundant minerals in cores of the Fourmile pegmatites (table 7). About 15 percent of the zoned pegmatites have massive quartz cores containing less than 5 percent of any other mineral (assemblage 12, table 6). About one-half of the zoned pegmatites contain quartz and perthite (assemblage 4, table 6) in the innermost exposed zone, but some of these may have a concealed quartz core. Cores in other less distinctly zoned bodies consist of quartz, perthite, and plagioclase (assemblage 3, table 6). In most of these, quartz is the predominant mineral, but about 30 percent of the cores have more perthite than quartz. Pegmatite 11 (pl. 21, C–1; table 7) contains an apparent core consisting largely of cleavelandite, but inasmuch as the exposure is only 3 feet thick, there may be other units, such as a quartz core, in thicker unexposed parts of the body. Accessory minerals include muscovite, tourmaline, beryl, columbite-tantalite, lithiophilite-triphylite, other lithium and phosphate minerals, microlite, tapiolite, and garnet. Beryl and lithium-bearing minerals are most abundant in the outer edges of the quartz or quartz-perthite cores and in the inner part of the adjacent outer zone, although large crystals of beryl may also be disseminated through perthite-rich cores. In some of the thinner zoned pegmatites columbite-tantalite is associated with beryl or phosphate minerals at the outer edges of the cores.

FRACTURE FILLINGS

Fracture fillings are defined as generally tabular bodies that fill fractures in previously consolidated pegmatite (Cameron and others, 1949, p. 14). They are present in nearly all the pegmatites in the Fourmile quadrangle. In some, they are only a few inches thick and several feet long, whereas in others they are several feet thick and extend across the entire host pegmatite. None was observed penetrating country rock outside the host pegmatite. Pegmatites 41, 116, and 161 (pl. 21, F–3 and G–6) and a few others are fracture fillings that are large enough to be mapped separately from the enclosing pegmatites. The Pleasant Valley mine (pl. 21, pegmatite 41, D–3) is entirely in a fracture filling unit about 10 feet thick.

The contacts of the fracture fillings are generally planar and parallel (figs. 83 and 84). The attitudes of the fracture fillings vary, but many, if not most, are nearly perpendicular to the plunge of the host pegmatite. Where such fracture fillings intersect the outer contacts of the host pegmatite, the pegmatite is indented by country rock as in a boudinage structure (figs. 84 and 101) and the fracture fillings separate the host pegmatite into boudins. A fracture filling in the New York pegmatite is normal to the dip but inclined to the plunge. Other fracture fillings are parallel to shear or joint surfaces in the pegmatite or in the country rock. A few others fill fractures that have no obvious relation to the structure of associated rocks.

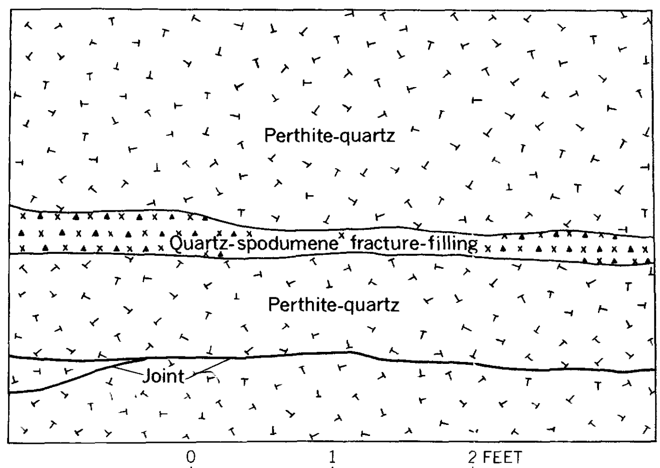

FIGURE 83.—Quartz-spodumene fracture filling crossing perthite-quartz pegmatite. Vertical exposure in opencut, Warren Draw pegmatite.

The contacts of the fracture fillings and host pegmatite may be either very sharp or somewhat gradational. In general, the contacts are more gradational where the mineralogy of the fracture filling resembles that of the host. Some fracture fillings have offset earlier units as shown in figure 91, but in other places, as in fracture fillings that lie between boudins of pegmatite, there is no clear evidence of forceful injection of the material in the fracture filling.

The mineral composition of the fracture fillings varies somewhat in different types of pegmatite. Those that cut zoned pegmatites contain minerals characteristic of one or more of the inner zones, and these fracture fillings cross only the outer zones of the host peg-

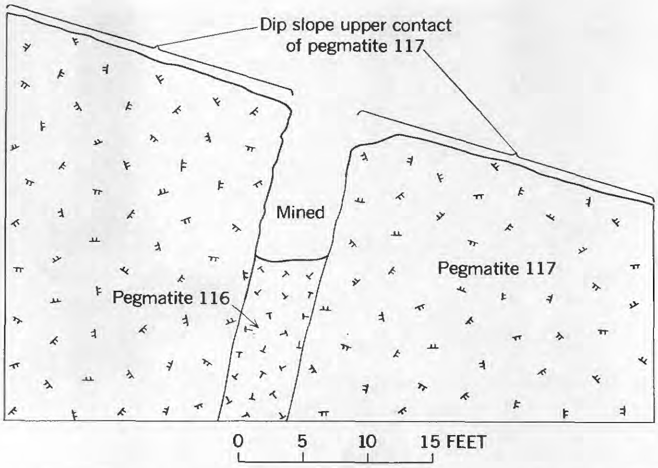

FIGURE 84.—Section showing pegmatite 116 transecting pegmatite 117. Shape of dipslope upper contact of pegmatite 117 suggests that pegmatite 116 was emplaced between boudins of pegmatite 117.

matite. The fracture fillings themselves may be zoned, as at the New York pegmatite, where they have a quartz core and a cleavelandite-quartz outer zone having the same composition as the two innermost zones of the pegmatite. Small quartz fracture fillings in the Big Tom pegmatite cut the outer zones and merge with the quartz core. In the Tin Mountain pegmatite, fracture fillings containing quartz, spodumene, plagioclase, amblygonite, and muscovite cut outer zones and are connected with an inner zone of similar mineralogy. The nearby Warren Draw pegmatite has similar fracture fillings, and it may be predicted that this pegmatite has lithium-bearing inner zones, though none have yet been found.

In structurally simple quartz-feldspar pegmatites the minerals of the fracture fillings are, in order of decreasing abundance, quartz, perthite, plagioclase, and muscovite. Commonly these fracture fillings also are zoned (or poorly zoned as in fig. 84), and their structure and mineralogy are similar to those of zoned pegmatites. Many of them contain a muscovite-rich outer zone, a perthite-rich intermediate zone, and a quartz-rich core.

Most of these fracture fillings have the same essential minerals as the host pegmatite and in addition have beryl, lithiophilite-triphylite, or other minerals that either are not present, or are much scarcer in the host pegmatite. In many places where the host pegmatite contains graphic perthite crystals, the fracture fillings contain nongraphic perthite. All the minerals in the fracture fillings are coarser grained and more nearly euhedral than the same minerals in the host.

There are probably all gradations among quartz fracture fillings, perthite-rich fracture fillings, and those that are mixtures of minerals in more or less the same proportions as in the host pegmatite. In general, the richer the fracture fillings are in quartz, the sharper are the contacts with the enclosing pegmatite. Some quartz-rich fillings transect the entire host pegmatite and cut earlier perthite-rich fracture fillings. Quartz fracture fillings like those shown on the geologic map cutting pegmatites 87 and 90 (pl. 21, F–2 and F–3) are restricted to the pegmatite and do not penetrate the country rock, and so were probably associated genetically with the pegmatite though they resemble quartz veins appearing elsewhere in the schists.

Many of the layered pegmatites contain fracture fillings, rich in potassic feldspar and quartz, that can be traced into lenticular segregations of coarser grained pegmatites that commonly are richer in perthite than the host. These segregations generally are parallel to the outer contacts of the host pegmatite and may have gradational contacts with layered bodies. Many of these segregations present no evidence that they have transected previously consolidated pegmatite in the manner of a true fracture filling. There are all gradations between such segregations and fracture fillings that effectively displace earlier crystallized units in the pegmatite.

REPLACEMENT BODIES

Replacement bodies have been defined as units formed by replacement of pre-existing pegmatite (Cameron and others, 1949, p. 14). Sizable replacement bodies have not been recognized in the pegmatites in the Fourmile quadrangle. The New York pegmatite (pl. 21) contains bodies a few feet across that are rich in cleavelandite and lithia mica and may be of replacement origin; mineralogically, they resemble replacement units in the Peerless and Hugo pegmatites, Keystone, S. Dak., that contain relicts of previously formed zones.

In layered pegmatite some of the coarser grained units rich in potassic feldspar contain apparent relicts of finer grained pegmatite layers that suggest a small amount of replacement of one rock by another.

LAYERS

Layers are parallel with the nearest contact of the pegmatite and the wallrock, but near the ends of individual pegmatite bodies, they commonly fade out into a rock of uniform composition, instead of adopting the concentric form that is characteristic of zones. Layers of one mineral assemblage alternating with layers of a second mineral assemblage may be repeated tens or even hundreds of times within a single pegmatite. The systematically progressive changes in mineralogy characteristic of zones, as shown in table 6, are not evident in layers.

Staatz and Trites (1955, p. 20–24) used the term "banding" for similar structure in pegmatites of the Quartz Creek district, Colorado, and Jahns and Wright (1951, p. 21–22, 27–30) referred apparently to similar structures in the pegmatites of the Pala district, California as "layered varieties of pegmatite." Jahns and Wright also described composite dikes that seem to be composed of layers (1951, figs. 26, 34, and pl. 12). The repeated units are clearly three dimensional and the term "layers" seems preferable to "bands".

Although there are many variations in the structure and texture of the layers, the combination of thickness, grain size, and composition enables one to divide the layers into two general types. Where the layers have a thickness of more than a few inches, each layer differs from its immediate neighbors both in grain size and in mineral composition. Typically in layers of this kind, there is an alternating sequence of relatively coarse-grained layers, consisting mainly of perthite, quartz, and plagioclase, and of finer grained layers where quartz and plagioclase are the dominant minerals. The most evident difference between adjacent layers is the large grain size of the perthite and associated minerals in the coarser layers.

Many of the finer grained layers are themselves divided into a thinner variety of layers that differ from each other mainly in composition because the dark minerals, tourmaline or garnet, are most abundant in alternate layers. These thinner layers are fine-grained or very fine grained and remarkably uniform in thickness. Pegmatite in the Pala area, California, with similar thin layers, was called "line rock" by Schaller (1925, p. 263), and this name is used here for such rock. Jahns and Wright (1951, p. 22), Staatz and Trites (1955, p. 23–24), and McLaughlin (1940, p. 62), described line rock or its equivalent in many pegmatite districts. There are apparently all gradations between the coarse- and fine-grained layers and the line rock.

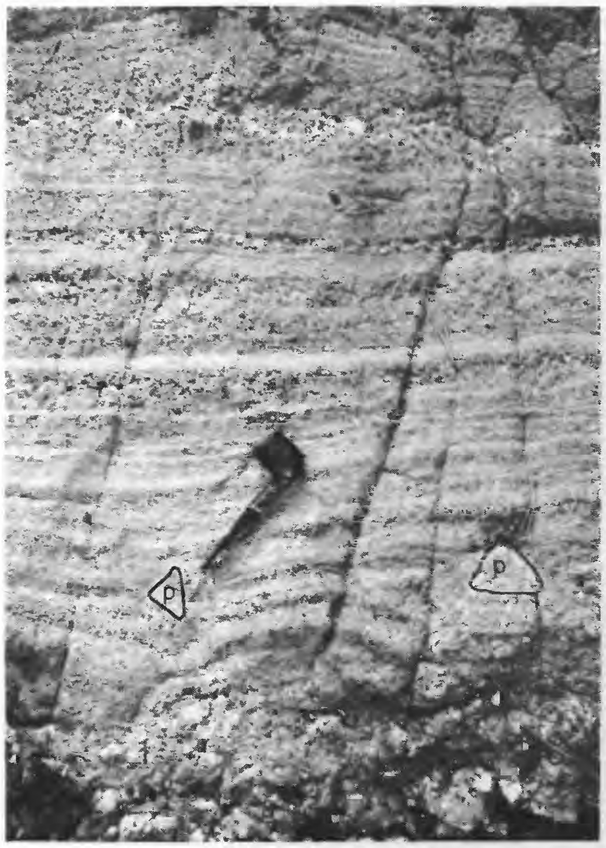

FIGURE 85.—Line rock in pegmatite 16 (pl. 22 loc. A). Dark layers are rich in tourmaline and quartz; light layers are mostly plagioclase. Graphic perthite crystal (p) near stem of pipe transects layers on bottom; layers bow around on top. Outer contact of pegmatite is below the photograph; center of pegmatite is above.

SIZE, SHAPE, AND DISTRIBUTION

The layers of alternating coarse- and fine-grained pegmatite range from a fraction of an inch to as much as 20 feet thick. Thin layers may be only a few feet long but thicker layers may persist along strike for hundreds of feet in rare instances. Line rock layers are commonly only about 0.3 inch thick (fig. 85) but single layers are as much as 4 inches thick.

The layers are generally almost planar and almost parallel to the contacts of the pegmatite. Though many of the host pegmatites are tabular and the layers are correspondingly simple in form, some of the pegmatites are irregular bodies and the layers concordantly follow bends or flexures in the contacts. In a few places the coarser grained layers cut adjacent fine-grained layers, thus forming fracture filling bodies. Near the ends of individual pegmatite bodies, the contacts between layers become diffuse as the compositional and textural differences of layers tend to disappear. Jahns and Wright (1951, p. 21) and Staatz and Trites (1955, p. 20) found a similar merging of layers.

Layering is somewhat more abundant near the outer contacts of individual pegmatite bodies than in the interiors, although it commonly persists throughout. Many of the line rock layers occur as isolated remnants adjacent to outer contacts (pl. 22).

The layers appear in pegmatites from all parts of the Fourmile quadrangle, but they are most abundant in the many large discordant bodies in the southeastern part of the area.

PETROGRAPHY

The coarser grained layers consist of perthite-quartz, perthite-quartz-plagioclase, perthite-quartz-plagioclase-muscovite, or quartz-plagioclase-muscovite pegmatite. Accessory minerals are tourmaline, apatite,

TABLE 8.—*Flame photometric and semiquantitative spectrographic analyses of layered and homogeneous pegmatite and of schist from the Calamity Peak area, 2 miles east of Custer, S. Dak.*

Samples collected by D. H. Kupfer.
Data on K_2O and Na_2O determined by flame photometer; right column by Sarah Berthold, left column by J. M. Dowd.
Spectrographic analyses by Janet Flecther. The concentrations of the elements as determined by semiquantitative spectrographic analysis are bracketed into groups each of approximately one-third of an order of magnitude, X+ indicating the higher portion (10 to 5 percent); X the middle portion (5 to 2 percent) and X− the lower portion (2 to 1 percent). Comparisons of this type of semiquantitative results with those obtained by quantitative methods, either chemical or spectrographic, show that the assigned group includes the quantitative value in about 60 percent of the analyses.
Data on BeO obtained from plate exposed for general scanning, and not for precise determination of BeO.

Description	Sample	K_2O		Na_2O		Sum of K_2O, Na_2O		K_2O ÷ Na_2O		CaO	MgO	Fe_2O_3	Cu	Pb	Zn	Ti	V	Cr	Mn	Ni	Sr	Zr	Ba	Ga	Y	La	Li_2O	BeO	B
Pegmatite																													
Fine grained layer	1	3.78	3.91	4.42	4.46	8.20	8.37	0.86	0.87	0.3–.6	0.1–.3	0.3–.6	0.000x	0.00x	-----	0.00x	-----	-----	0.x	-----	0.00x	0.00x	0.000x	0.00x	-----	-----	<0.10	0.008	0.00x
Do	6	3.78	3.91	4.62	4.71	8.40	8 62	.82	.83	.1–.3	< .1	.3–.6	.00x	.00x	-----	.00x	-----	0.0x	.0x	0.00x	.00x	.00x	.00x	.00x	-----	-----	< .10	.001	.0x
Do	8	1.04	1.06	6.42	6.67	7.44	7.73	.16	.16	.3–.6	< .1	.3–.6	.000x	.000x	-----	.00x	-----	.0x	.0x	.00x	.00x	.00x	.00x	.00x	-----	-----	< .10	.001	.0x
Do	9	n.d.	2.90	n.d.	4.65	n.d.	7.55	n.d.	.62	.3–.6	.06–.1	.6–1.0	.000x	.00x	-----	.0x	-----	.000x	.0x	.00x	.00x	.00x	.00x	.00x	-----	-----	n.d.	.001	.0x
Coarse grained layer	2	6.47	6.62	3.28	3.26	9.75	9.88	1.97	2.03	.1–.3	< .1	.3–.6	.00x	.00x	-----	.00x	-----	.00x	.0x	.00x	.00x	.00x	.0x	.00x	-----	-----	< .10	.0006	.0x
Do	5	7.42	7.62	2.92	3.03	10.34	10.65	2.54	2.51	< .1	< .1	.1–.3	.000x	.00x	-----	.00x	-----	-----	.00x	-----	.00x	-----	.0x	.00x	-----	-----	< .10	.001	.0x
Do	7	8.61	8.88	3.15	3.29	11.76	12.41	2.73	2.77	< .1	< .1	.1–.3	.00x	.00x	-----	.00x	-----	-----	.00x	-----	.00x	-----	.00x	.00x	-----	-----	< .10	.0008	.0x
Do	10	n.d.	5.78	n.d.	3.37	n.d.	9.15	n.d.	1.72	.1–.3	.06–.1	.6–1.0	.000x	.00x	-----	.0x	-----	.000x	.0x	-----	.00x	.00x	.00x	.00x	-----	-----	n.d.	.001	.0x
Fracture filling	4	n.d.	4.25	n.d.	2.05	n.d.	6.30	n.d.	2.07	.3–.6	< .1	.3–.6	.00x	.00x	-----	.00x	-----	.00x	.0x	.00x	.00x	-----	.0x	.00x	-----	-----	n.d.	.002	.0x
Do	11	n.d.	3.54	n.d.	3.78	n.d.	7.32	n.d.	.94	.1–.3	.06–.1	1–3	.000x	.00x	-----	.0x	-----	-----	.0x	-----	.00x	.00x	.00x	.00x	-----	-----	n.d.	.001	.x
Do	15	n.d.	4.43	n.d.	2.89	n.d.	7.32	n.d.	1.53	.1–.3	.06–.1	1–3	.00x	.00x	-----	.x	-----	-----	.00x	n.d.	.00x	-----	.0x	.00x	-----	-----	n.d.	.001	.0x
Homogeneous pegmatite	16	n.d.	4.81	n.d.	2.73	n.d.	7.54	n.d.	1.8	.1–.3	n.d.	1–3	.000x	.00x	-----	.x	-----	.000x	.00x	-----	.00x	.00x	.0x	.00x	-----	-----	n.d.	.0008	.06–.1
Do	17	n.d.	4.87	n.d.	3.81	n.d.	7.68	n.d.	1.3	.06–.1	n.d.	1–3	.00x	.00x	-----	.0x	-----	.000x	.0x	-----	.00x	-----	.00x	.00x	-----	-----	n.d/	.001	.06–.1
Schist																													
Quartz-mica shist inclusion in pegmatite	12	n.d.	3.44	n.d.	2.14	n.d.	5.58	n.d.	1.6	0.1–.3	1–3	3–6	0.00x	0.00x	0.0x	0.x	0.0x	0.0x	0.0x	0.00x	0.0x	0.0x	0.0x	0.00x	0.0x	0.0x	n.d.	0.001	x.0
Quartz-rich schist	13	n.d.	1.45	n.d.	2.48	n.d.	3.93	n.d.	.6	.6–1	.6–1	1–3	.00x	-----	-----	.x	.00x	.00x	.0x	.00x	.0x	.0x	.0x	.00x	.0x	.00x	n.d.	.0004	.0x
Quartz-sillimanite schist	14	n.d.	3.51	n.d.	1.43	n.d.	4.94	n.d.	2.4	.06–.1	.6–1	3–6	.00x	-----	-----	.x	.0x	.0x	.0x	.00x	.0x	.0x	.0x	.00x	-----	.00x	n.d.	.0004	.0x
Quartz-mica schist	18	n.d.	4.20	n.d.	1.52	n.d.	5.72	n.d.	2.8	.1–.3	.6–1	3–6	.00x	.0x	-----	.x	.0x	.0x	.0x	.00x	.0x	.0x	.0x	.00x	.0x	.00x	n.d.	.001	.00x

and garnet. The finer grained layers contain plagioclase-quartz-perthite-muscovite, plagioclase-quartz-muscovite, or plagioclase-quartz pegmatite. Garnet is somewhat more abundant than in the coarser grained layers but otherwise the accessories are the same. Perthite-quartz-plagioclase pegmatite is probably the most common rock of the coarser grained layers, and plagioclase-quartz-perthite-muscovite pegmatite predominates in the finer grained layers. In general, the coarser grained layers correspond to assemblages 3 and 4 (table 6), and the finer-grained layers to assemblages 1, 2, and 3. The average grain size ranges from 0.4 to 4 inches in the coarser grained layers and from 0.03 to 0.5 inch in the finer layers. Perthite in both varieties of layers occurs as phenocrysts that may be 10 or more times as large as the average grain size.

Jahns and Wright (1951, p. 21) and Staatz and Trites (1955, p. 20–24) also described coarser grained layers that are rich in perthite and quartz and finer grained layers that are rich in plagioclase, quartz, and muscovite. McLaughlin (1940, p. 62) found apparently similar layering in the Bridger Mountain pegmatites of Wyoming in which the coarser grained layers consist of microcline, quartz, and muscovite and the finer grained layers consist of cleavelandite, quartz, muscovite, and garnet.

Great differences in the alkali content of the layers are indicated by their mineralogy and confirmed by the analyses shown in table 8. The samples in table 8 were collected by D. H. Kupfer from a large layered pegmatite a few hundred feet from Calamity Peak (fig. 75) in which scores and perhaps hundreds of layers are parallel to the outer contact of the pegmatite. The layers are about 1 foot thick and may be traced for hundreds of feet. The various layers probably do not contain more than 5 percent muscovite, and if the amount of K_2O necessary to make that much muscovite is subtracted from the total in table 8, the remaining K_2O and all of the Na_2O may be used to calculate the approximate amount of feldspar in each layer. If this calculation is done according to the CIPW method (assuming adequate Al_2O_3 and SiO_2), the content of albite in the coarser grained layers from Calamity Peak ranges from 19 to 22 percent and the microcline content ranges from 23 to 56 percent. The albite in the finer grained layers ranges from 36 to 59 percent and the microcline from 2 to 24 percent. The remainder of the rock is mostly quartz, and by subtracting the feldspar and mica values from 100 the quartz content is found to be 25 to 35 percent of both the coarser and finer grained layers.

The spectographic analyses also show that the finer grained layers tend to be richer in CaO, MgO, Fe_2O_3, BeO, Cr, Mn, Ni, and Zr than are the coarser grained layers. The only minor element that is clearly more abundant in the coarser grained layers is Ba, which presumably substitutes for K in the potassic feldspar (Rankama and Sahama, 1950, p. 471–472).

The line rock layers are rich in plagioclase, quartz, tourmaline, garnet, and muscovite. Sharp variations in the amount of tourmaline, garnet, muscovite, or some other mineral give the rock its striking layered appearance (fig. 85).

The commonest variety of line rock in the pegmatite of the Fourmile area consists of dark tourmaline-rich layers alternating with light feldspar-rich layers. Less abundant varieties of dark layers are rich in garnet or green muscovite. Table 9 contains modes for a few of the many light and dark layers of line rock. In general, the dark layers are rich in minerals containing Fe, B, OH, or K, and the light layers are poor in these minerals.

The layers of line rock that are unusually rich in tourmaline or muscovite are also very rich in quartz, and conversely, where a layer contains relatively small amounts of tourmaline or muscovite, it also is deficient in quartz. As the quartz content decreases, the plagioclase content increases.

TABLE 9.—*Estimated modes of layers in line rock*

[Samples from pegmatites 16 and 57, pl. 21]

Mineral	1	2	3	4	5	6
Dark layers						
Quartz	49	42	51	43	20	40
Plagioclase	28	39	35	43	49	12
Microcline	7	7	------	------	28	3
Muscovite	8	5	14	1	------	------
Tourmaline	8	5	------	12	3	45
Apatite	------	1	------	1	------	------
Garnet	------	1	------	------	------	------
Total	100	100	100	100	100	100
Light layers						
Quartz	47	42	17	7	26	------
Plagioclase	30	54	74	90	52	------
Microcline	20	------	5	2	18	------
Muscovite	2	4	1	1	2	------
Tourmaline	1	------	2	1	1	------
Apatite	------	------	1	------	1	------
Total	100	100	100	100	100	------

The average grain size of the line rock is about 0.05 inch, but crystals of perthite several inches across may occur between and crosscutting the layers (fig. 85). In thin section, the feldspar and quartz are seen to be mostly anhedral, muscovite and garnet subhedral, and crystals of tourmaline euhedral. Tourmaline is commonly oriented subparallel to the plane of each layer but with no noticeable lineation. The contacts between the layers are generally somewhat gradational in thin sections because differences in color of the minerals are not so evident as in hand specimens. The composition of the plagioclase does not change perceptibly in adjacent layers.

STRUCTURE

Most of the layers are approximately planar and are parallel to pegmatite contact. In many places the line rock layers follow the pegmatite contact or contacts of wallrock inclusions in great detail (fig. 86).

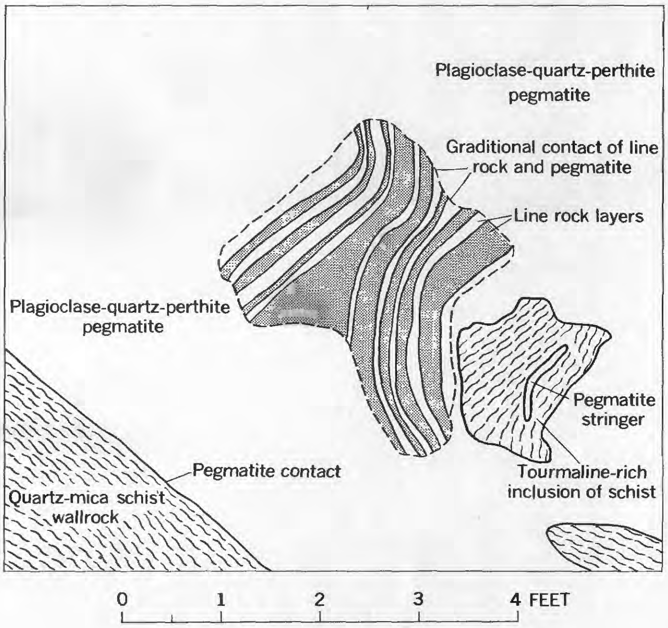

FIGURE 86.—Sketch showing inclusion of schist in pegmatite partly surrounded by line rock structure. Foliation of inclusion is at right angles to part of layering. SE¼NW¼ sec. 25, T. 3 S., R. 3 E., Berne quadrangle, South Dakota.

In a few outcrops the layers of line rock are moderately plicated or warped (fig. 87) and in part are not parallel to the contact of the pegmatite body. The plications generally die out rapidly inward and may be bordered on both sides by relatively straight layers (figs. 87 and 88). Some thickening and thinning occur in the noses of the flexures of the different layers. Rarely, some of the layers nearer the interior of a single body transect the noses of flexures in layers near the outer contacts.

Layers of line rock are deflected around large crystals of perthite and graphic perthite (figs. 85 and 88) that are locally disseminated in the line rock. Individual layers of line rock commonly curve inward toward the center of the pegmatite body in passing around these crystals but never outward toward the outer contact of the pegmatite body. Where the line rock has been deflected in this way toward the center of the pegmatite, it forms a crude hemisphere around the crystals of perthite. Many of the graphic perthite

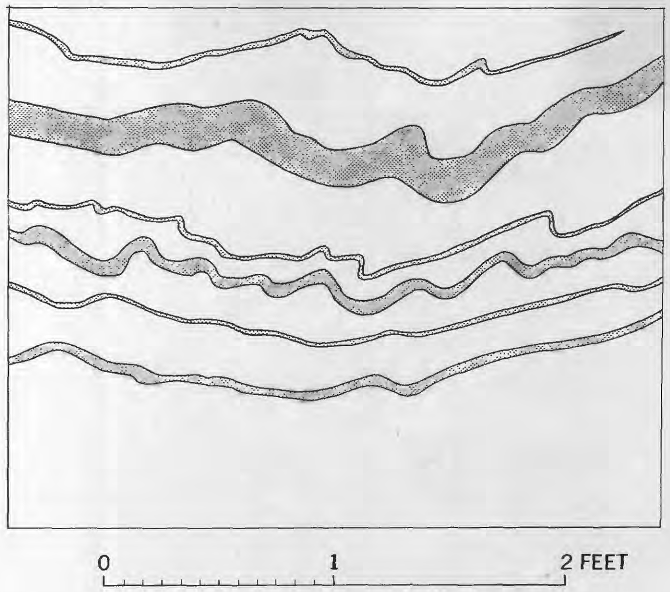

FIGURE 87.—Sketch showing crinkles and folds in line rock from Warren Draw pegmatite. Dark layers are rich in muscovite and quartz; light layers are rich in plagioclase.

crystals in the interiors of these hemispheres are fan shaped with the apex of the fan pointing toward the outer wall of the pegmatite. Only a few inches away from the perthite, however, layers are generally planar and uncurved.

Some layers of line rock that pass through perthite crystals resemble the phantom layers described by Jahns (1953, p. 584–585). These layers can be recognized mainly from the presence of tiny needlelike tourmaline inclusions in the perthite. The tourmaline inclusions are oriented approximately perpendicular to the layers (fig. 89), instead of being either parallel to the layers or nearly unoriented as they are outside the perthite crystals. The phantom layers commonly are planar rather than rounded, and probably all are parallel to rational crystallographic planes of the microcline. Seven oriented thin sections cut parallel to the plane of the layers in several perthite crystals, were examined with a universal stage; the layers were found to be parallel to the (201), the (110), or the (001) of the microcline. Angular bends appear in some layers where they change from one orientation to another.

The coarser layers commonly cut across line rock as shown in figure 88, and in places, as shown in figure 90, offshoots from the coarse layers form fracture fillings that cut and displace the layers of line rocks.

SEQUENCE OF CRYSTALLIZATION OF THE LAYERS

The relations among the layers indicate that in general they crystallized in a sequence from the pegmatite contact inward and that all the layers in the outer part of a pegmatite, regardless of their lithologic character,

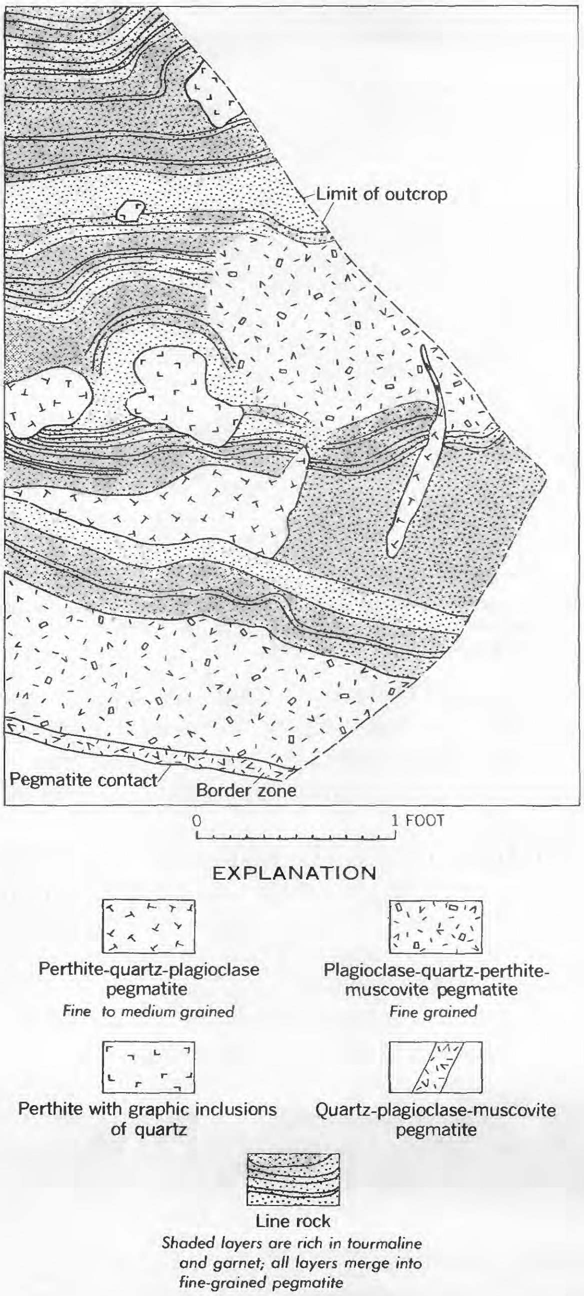

Figure 88.—Sketch showing layers of line rock cut by perthite-quartz-plagioclase pegmatite in pegmatite 16. Layers are deflected around the inner borders of graphic perthite crystals, but are cut by the outer borders of these same crystals.

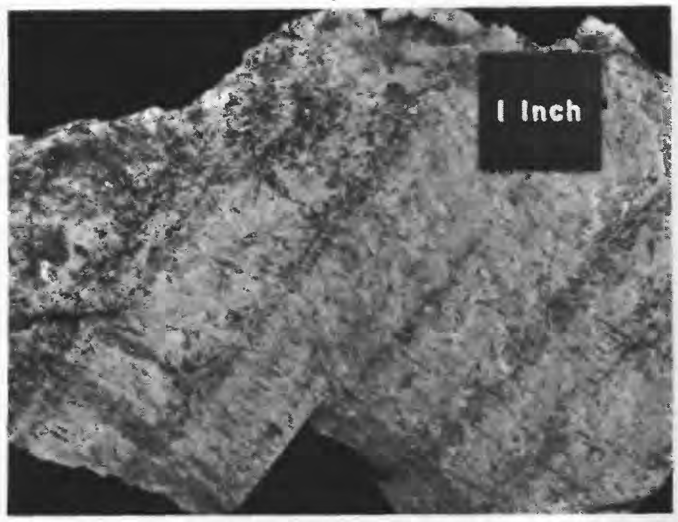

Figure 89.—Phantom layers of line rock in graphic perthite crystal. Layers are parallel to ($\bar{2}01$) of the microcline. Very tiny tourmaline needles (not visible individually in the photograph) and quartz rods are predominantly perpendicular to layers.

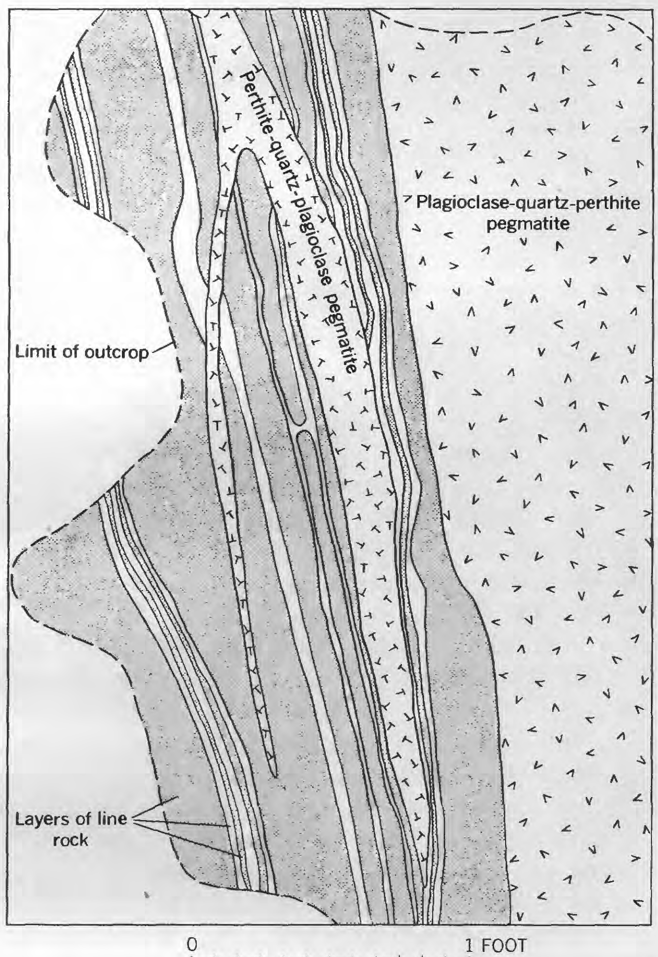

Figure 90.—Sketch showing displacement of line rock by coarser grained perthite-rich fracture filling in pegmatite 13. Dark layers are rich in tourmaline and quartz; light layers are rich in plagioclase.

crystallized prior to all layers in the center of the pegmatite.

The relatively coarse grained perthite-rich layers commonly have offshoots that form fracture fillings in adjacent line rock, but in all the observed examples these fracture fillings go outward toward the pegmatite contact, never inward toward the center of a pegmatite body. Furthermore, this line rock may be transected on one side of a large single crystal of perthite and deflected around the other side as shown in figure 88. The transecting side, where the line rock is unquestionably older, is toward the pegmatite contact, and the side with deflected layers, which presumably were controlled by the coarser material and thus are younger, is toward the center of the pegmatite. Where the crests of flexures in the line rock are transected by younger layers, the younger layers are invariably nearer the center of the pegmatite.

Tapered crystals of perthite or other minerals that expand inward toward the center of the pegmatite have commonly been used as evidence for crystallization of pegmatite from the wall inward (Jahns, 1953, p. 584–585). It has been supposed that these crystals started from a seed at what is now the small end, and as the crystal grew outward into the magma, new material was added on the inner end and at the sides. A tapered crystal shown in figure 91 has phantom layers of line rock that are displaced from the corresponding layers in the matrix. An especially striking feature of the displacement is that it is smallest at the small outer end of the crystal and becomes progressively greater toward the inner end. The only logical explanation is that the perthite crystal grew inward at a faster rate than the line rock. This interpretation is also in accord with the observed fact that phantom layers are parallel to several rational crystallographic planes that may have been crystal faces temporarily during the development of the microcline. Line rock apparently formed along these crystal faces and immediately thereafter became engulfed by later additions to the microcline crystal.

Some of the line rock, however, may have been replaced by coarser pegmatite or resorbed by the liquid prior to crystallization of coarser pegmatite. Not only is the line rock transected by the coarser pegmatite as along the discordant crystals shown in figure 88, but there are also remnants of line rock surrounded entirely by coarser pegmatite (fig. 92). Furthermore, a phantom tourmaline layer in a perthite crystal that is now surrounded entirely by coarse pegmatite suggests that the surrounding matrix was at one time line rock.

Line rock is generally more abundant in the outer parts of pegmatite bodies, and the coarser grained phases are generally more abundant in the inner parts. This relation, taken in conjunction with the evidence for crystallization from the contact inward, shows that in a broad sense the line rock was generally early and the coarser varieties of pegmatite formed later.

In interpreting the sequence of crystallization in any specific pegmatite, it is important to keep in mind the

FIGURE 91.—Tapered crystal of perthite (p) that contains phantom layers of line rock and is surrounded by a matrix of line rock. At the small end of the perthite crystal, which is nearest the pegmatite contact, the line rock layers are only slightly displaced. The displacement becomes progressively greater toward the inner end of the perthite crystal. Natural size.

fact that a structure much like layering may arise by multiple intrusion. Where several pegmatites, each having fine-grained outer units and coarse-grained inner units, coalesce to form a single body, this body will necessarily have alternating fine and coarse units that resemble the layers here described, but the innerward parts of these multiple intrusives did not necessarily form before the outer parts.

TYPES OF PEGMATITE BODIES

The pegmatite bodies in the Fourmile quadrangle have been divided into layered, homogeneous, and zoned varieties on the basis of the predominant internal structure. This division is not clearcut; there apparently are all gradations among the various types of

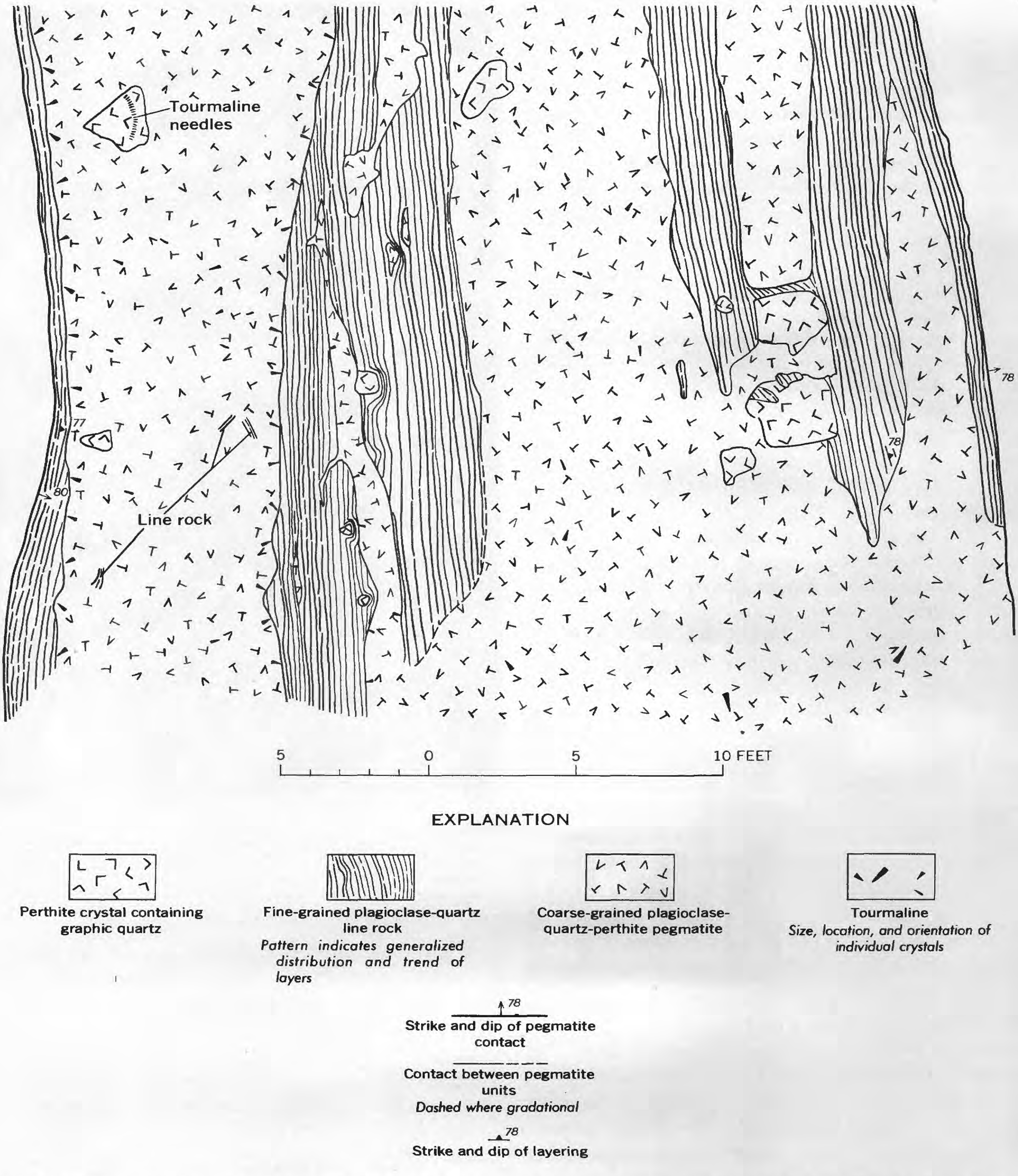

FIGURE 92.—Diagram showing relation between line rock and coarser grained pegmatite in pegmatite 16 (pl. 22, loc A). Phantom line rock layers in graphic perthite crystal, upper left, are inferred to be relict line rock layers that were replaced by the coarser grained pegmatite. The pegmatite cuts line rock in the left center of the diagram.

pegmatites. Some of the large layered bodies, such as pegmatite 166 (pl. 21, G–7), have limbs or branches without any noticeable internal structures. On the other hand, the outer parts of a few zoned pegmatites, such as the Warren Draw pegmatite, are layered (fig. 93).

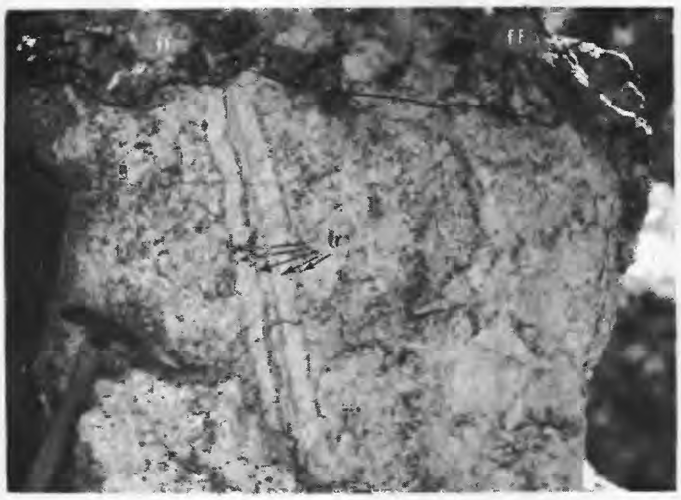

FIGURE 93.—Layered outer part (wall zone) of a zoned pegmatite cut by a fracture filling (ff). Contact with schist at left side of outcrop is nearly vertical. Parts of the fracture filling above the photograph and a muscovite-rich zone immediately to the right of the exposure have been removed by mining. Prominent line rock (lr) consists of plagioclase layers alternating with quartz-muscovite layers; less prominent layers on either side are not readily visible in the photograph. Fracture filling consisting of quartz, perthite, and muscovite is similar in composition to the inner zones of the pegmatite; though it extends outward across the wall zone, it does not penetrate the country rock. Wilhelm pegmatite, sec. 32, T. 2 S., R. 4 E., Berne quadrangle, South Dakota.

LAYERED PEGMATITES

A layered pegmatite is one in which the predominant internal structure is the layering described on page 229–230 of this report. An additional structure is a thin border zone, generally less than 2 inches thick, at the outer contact (fig. 88). Fracture fillings also are generally present. The layers may make up most of a single body or only a small part of it (pl. 22); consequently, the layers also occur in other types of pegmatites, but are not a prominent feature.

The layered pegmatite bodies of the Fourmile quadrangle are very similar to the granitic rocks of the large area near Harney Peak mapped by Darton and Paige (1925), who stated that this coarse-grained rudely layered rock could properly be called pegmatite. Furthermore, reconnaissance northeast of the Fourmile quadrangle confirms that layered pegmatites are more abundant nearer the granite area.

Layered pegmatites are most abundant in the southeast part of the Fourmile quadrangle, where they form as much as 25 percent of all the rocks. Even in other parts of the area they are sufficiently numerous and commonly so large that layered pegmatite is more abundant throughout the quadrangle than any other type. More than 99 percent of the layered bodies are either in the Mayo formation or in amphibolite; more than 90 percent of them are discordant; and many are irregularly shaped.

The essential minerals of the layered pegmatites are in order of decreasing abundance, plagioclase, (An_{2-14}), quartz, perthite (mostly with graphic quartz), and muscovite. The accessories are tourmaline, apatite, garnet, and rarely beryl. The estimated mineral content of some of the many layered bodies is given in table 10. The average mode of layered pegmatite is 40 percent plagioclase, 31 percent quartz, 23 percent perthite, 4 percent muscovite, 1 percent tourmaline and 1 percent other accessories. When potassic feldspar is fine grained it may have been mistaken for plagioclase, but otherwise the modes of table 10 are not likely to be significantly in error.

In some gently dipping layered bodies, coarse-grained perthite-rich layers are most common along the hanging wall. Similar positions of perthite-rich layers in Colorado were noted by Staatz and Trites (1955, p. 20) and in California by Jahns and Wright (1951, p. 21–22).

Between 25 and 75 percent of the rock in layered pegmatites has the layered structure. The layers are well developed in the thicker or tabular parts of bodies and gradually merge near the ends of such bodies into homogeneous pegmatite of intermediate grain size (pl. 22). Aside from the layers, the most distinctive feature of the layered pegmatites is a grain size generally smaller than that in the zoned and homogeneous pegmatites. The coarser grained layers have pegmatitic textures and about the same grain size as the homogeneous bodies. The finer grained layers, however, are predominantly granitoid.

HOMOGENEOUS PEGMATITES

Homogeneous pegmatites are those that have no prominent internal structure. In most respects they resemble the homogeneous pegmatites in northeast Brazil described by Johnston (1945, p. 1025). All have fine-grained border zones a few inches thick, and some have fracture filling units and podlike aggregates of coarse-grained minerals. In the northwest part of the Fourmile quadrangle many of the pegmatites are transitional in character between homogeneous and zoned pegmatites.

The homogeneous pegmatites are generally small, and although 85 percent of the pegmatite bodies in the quadrangle are of this variety, the total quantity of rock contained in them is less than in the layered ones. Though the sizes, shapes, and attitudes have a great range, the homogeneous pegmatites are predominantly thin, tabular sills ranging in length from tens of feet to more than 3,000 feet. Johnston (1945, p. 1025), in defining homogeneous pegmatites, mentioned that they have thicknesses of only 2 to 3 meters. In the Fourmile quadrangle, as in the area described by Johnston, nearly all the homogeneous pegmatites are less than 10

TABLE 10.—*Mineralogy of layered and homogeneous pegmatites*

Wallrock type: A, amphibolite; QBS, quartz-biotite schist; QMS, quartz-mica schist; QMSS, quartz-mica-sillimanite schist; QMSTS, quartz-mica-staurolite schist.

Relation to wallrock: C, concordant; D, discordant.
Shape: Ir, irregular; L, lenticular; T, tabular; TB, tabular-branching.
Internal structure: H, homogeneous; L, layered.

Pegmatite (pl. 21)	Wallrock, type	Pegmatite																				
		Relation to wallrock	Shape	Internal structure	Mineralogy																	
					Plagioclase		Perthite		Quartz		Muscovite		Garnet		Tourmaline		Lithiophilite-triphylite		Biotite		Beryl	
					Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)
2	QMS, A	C	T	L	47	0.5	15	4	35	0.7	3	0.7	Tr.	0.1	Tr.	0.5	-----	-----	-----	-----	-----	-----
4	QMS	D	T	H	40	.3	23	5	33	.5	4	.5	-----	-----	Tr.	.4	-----	-----	-----	-----	Tr.	0.7
23	QMS	D	Ir	L	45	.2	25	2	24	.2	5	.3	Tr.	.1	1	.1	-----	-----	-----	-----	Tr.	1.0
25	QMS, QMSTS	D	T	L	50	.1	10	1	35	.2	5	.1	Tr.	.1	<1	.1	-----	-----	-----	-----	Tr.	.2
30	QMS	D	TB	H	32	.3	38	6	28	.4	2	.3	Tr.	.1	Tr.	.6	-----	-----	-----	-----	Tr.	2.0
32	QMS	D	T	H	35	.2	30	3	30	.5	4	.6	Tr.	.2	<1	.6	-----	-----	-----	-----	-----	-----
35	QMS	C	T	L	40	.2	30	2	25	.2	4	.3	Tr.	.1	1	.2	-----	-----	-----	-----	Tr.	.6
37	QMS	D	T	H	30	.5	33	6	34	.5	5	1.0	Tr.	.2	1	1.0	Tr. +	1	Tr.	0.8	Tr.	1.5
38	QMS	D	TB	L	33	.4	30	6	34	.5	5	.8	Tr.	.2	1	.8	-----	-----	-----	-----	-----	-----
39	QMS	D	T	L	38	.2	30	6	25	.5	4	.3	Tr.	.2	3	.5	-----	-----	-----	-----	-----	-----
40	QMS	D	T	H	37	.3	25	4	30	.3	5	.2	Tr.	.1	2	.2	-----	-----	-----	-----	-----	-----
43	QMS	D	T	H	38	.5	25	10	30	.5	5	.5	-----	-----	2	.3	Tr.	1	-----	-----	Tr.	1.0
44	QMS	D	T	L	50	.1	10	1	30	.2	8	.2	Tr.	.1	2	.1	-----	-----	-----	-----	-----	-----
45	QMS	D	T	H	50	.2	16	1	28	.2	4	.2	Tr.	.1	2	.1	-----	-----	-----	-----	-----	-----
51	QMS	D	T	H	35	.3	30	5	27	.4	5	.5	-----	-----	3	.5	Tr.	2	-----	-----	Tr.	.5
54	QMS	D	L	L	32	.3	38	8	25	.3	3	.3	Tr.	.1	2	.2	-----	-----	-----	-----	Tr.	.8
57	QMS, A	C	T	L	45	.05	20	1	30	.1	4	.1	Tr.	.05	Tr.	.05	-----	-----	-----	-----	-----	-----
63	QMS	D	T	H	45	.2	25	2	25	.3	4	.3	Tr.	.1	1	.1	-----	-----	-----	-----	-----	-----
67	QMSS	C	L	H	35	.5	40	8	22	.5	2	.5	Tr.	.1	1	1.0	-----	-----	-----	-----	-----	-----
69	QBS, A	C	T	H	40	.3	35	3	22	.5	3	1.0	-----	-----	Tr.	.2				--	-----	-----
71	QMS, QMSS	D	TB	H	37	.3	29	2	32	.3	2	.3	-----	-----	1	.5	-----	-----	-----	-----	Tr.	.8
72	QMSTS	C(?)	T	H	33	.3	25	4	35	.5	3	.6	Tr.	.1	2	.5	-----	-----	-----	-----	-----	-----
75	QMS	D	T	H	40	.5	23	3	33	.7	4	.5	-----	-----	Tr.	.1	-----	-----	-----	-----	-----	-----
77	QMSTS	D	T	H	52	.7	20	2	25	.5	3	.5	-----	-----	Tr.	.2	-----	-----	-----	-----	-----	-----
80	QMS, QMSS	C	L	H	20	.5	50	18	25	1.0	3	2.0	Tr.	1.0	1	1.0	-----	-----	1	3	-----	-----
82	QMS	D	T	H	36	.3	28	6	33	.3	2	.3	Tr.	.1	1	.2	-----	-----	-----	-----	-----	-----
83	QMS	D	T	L	45	.1	16	2	33	.1	6	.2	Tr.	.1	Tr.	.1	-----	-----	-----	-----	-----	-----
90	QMSS	CD	Ir	L	40	.2	20	8	35	.3	4	.5	Tr.	.1	1	.2	-----	-----	-----	-----	-----	-----
91	QMS	D	L	H	28	.4	40	8	30	.6	1	.8	-----	-----	1	.2	-----	-----	-----	-----	-----	-----
92	QMS	C	TB	H	38	.4	29	8	30	.5	2	.7	Tr.	.1	Tr.	.3	-----	-----	-----	-----	Tr.	.8
95	QMSS	C	TB	L	42	.2	20	4	30	.4	5	.5	Tr.	.1	3	.2	-----	-----	-----	-----	-----	-----
96	QMS	C	T	H	35	.4	26	2	35	.2	3	.5	-----	-----	1	.1	-----	-----	-----	-----	-----	-----
97	QMS	C	T	H	33	.3	31	2	33	.4	2	.6	Tr.	0.1	1	.3	-----	-----	-----	-----	-----	-----
98	QMS	C	T	H	20	.4	50	4	27	.5	3	.5	-----	-----	1	.3	-----	-----	-----	-----	-----	-----
101	QMS	CD	L	H	27	.5	38	7	31	.6	3	.8	-----	-----	1	.2	-----	-----	-----	-----	-----	-----
102	QMS	C	T	H	37	.3	25	2	35	.3	2	.4	Tr.	.1	1	.3	-----	-----	-----	-----	-----	-----
103	QMSS	C	T	L	40	.3	20	2	35	.2	2	.3	Tr.	.1	2	.1	-----	-----	-----	-----	-----	-----
104	QMSS	C	L	L	34	.2	29	4	31	.2	5	.4	Tr.	.2	1	-----	-----	-----	-----	-----	Tr.	.5
107	QMS	C	T	H	33	.3	30	5	33	.2	3	.4	Tr.	.1	<1	.2	-----	-----	-----	-----	-----	-----
108	QMS	C	TB	L	37	.2	25	4	33	.2	3	.2	Tr. +	.1	1	.1	-----	-----	-----	-----	-----	-----
109	QMS	C	T	L	36	.2	28	3	31	.3	3	.3	Tr.	.1	<1	.2	-----	-----	-----	-----	-----	-----
110	QMS	C	T	H	31	.4	35	4	31	.4	1	.5	-----	-----	2	.3	-----	-----	-----	-----	-----	-----
111	QMS	C	T	H	39	.5	20	3	39	.5	1	.5	Tr.	.1	1	.4	-----	-----	-----	-----	-----	-----
112	QMSS	C	L	H	35	.2	25	5	35	.4	4	.3	Tr.	.1	1	.2	-----	-----	-----	-----	-----	-----
115	QMS	D	TB	L	43	.1	22	3	32	.2	2	.2	Tr.	.1	<1	.2	-----	-----	-----	-----	-----	-----
117	QMS	C	L	L	35	.4	32	6	25	.4	6	.2	Tr.	.1	1	.2	-----	-----	Tr.	1	-----	-----
118	QMS	C	L	L	40	.3	24	8	32	.3	4	.5	Tr. +	.1	<1	.2	-----	-----	-----	-----	-----	-----
119	QMS	C	L	L	32	.2	26	2	34	.2	5	.3	Tr.	.1	1	.2	-----	-----	-----	-----	-----	-----
122	QMS	C	T	H	40	.2	21	3	33	.2	3	.3	-----	-----	1	.1	-----	-----	-----	-----	-----	-----
124	QMS	C	L	H	35	.4	26	8	32	.5	5	.8	Tr.	.2	<1	.2	-----	-----	Tr.	8	-----	-----
125	A, QMS	D	T	H	33	.1	25	3	35	.2	3	.2	1	.1	3	.1	-----	-----	-----	-----	Tr.	.5
127	QMS	C	T	H	33	.5	30	4	33	.7	2	.8	-----	-----	1	.5	-----	-----	-----	-----	-----	-----
128	QMS	D	L	L	40	.2	20	4	35	.3	2	.5	Tr.	-----	2	.5	-----	-----	-----	-----	-----	-----
129	QMS	C	T	L	40	.1	20	3	35	.1	4	.2	<1	.1	<1	.1	-----	-----	-----	-----	-----	-----
130	QMS	CD	Ir	L	40	.2	20	3	35	.2	4	.2	Tr.	.1	<1	.1	-----	-----	-----	-----	-----	-----
131	QMS	C	T	H	37	.2	25	4	33	.2	4	.5	-----	-----	1	.1	-----	-----	-----	-----	-----	-----
132	QMSS, QBS	C	T	H	21	.3	40	3	30	.3	8	.6	Tr.	.1	Tr.	.2	-----	-----	-----	-----	-----	-----
133	QMSS	C	T	H	37	.2	20	3	33	.2	10	.2	Tr.	.1	-----	-----	-----	-----	-----	-----	-----	-----
135	QMS	C	T	H	27	.5	33	5	30	.5	8	.8	Tr.	.1	1	.3	-----	-----	-----	-----	-----	-----
136	QMS	C	T	H	26	.5	39	4	30	.5	4	.8	Tr.	.2	1	.5	-----	-----	-----	-----	-----	-----
141	QMS	CD	B	H	40	.2	20	1	36	.2	2	.4	Tr.	.1	1	.2	-----	-----	-----	-----	-----	-----
151	QMS	D	T	H	35	.4	27	5	33	.4	4	.6	-----	-----	Tr.	.4	-----	-----	-----	-----	-----	-----
152	QMS	D	T	H	27	.6	38	4	33	.4	4	.6	-----	-----	Tr.	.5	-----	-----	-----	-----	-----	-----
153	QMS	C	T	H	45	.2	20	3	30	.3	4	.5	-----	-----	1	.5	-----	-----	-----	-----	-----	-----
154	QBS, QMS	C	T	H	24	.6	30	7	38	.8	6	.9	Tr.	0.3	2	.8	-----	-----	-----	-----	-----	-----
155	QMSS, QBS	C	T	H	30	1.0	30	8	30	1.0	7	1.0	Tr.	.2	3	.8	-----	-----	-----	-----	-----	-----
156	QMSS, QBS	C	T	H	20	.5	32	6	40	.8	7	.9	Tr.	.1	1	.5	-----	-----	-----	-----	-----	-----
157	QMSS, QBS	C	T	H	33	.4	35	5	27	.4	4	.6	-----	-----	Tr.	.3	-----	-----	-----	-----	-----	-----
158	QMSS, QBS	C	T	H	36	.8	25	2	33	.5	3	.8	-----	-----	2	.6	-----	-----	-----	-----	-----	-----
162	QMS	C	T	H	34	.2	28	4	32	.4	4	.5	-----	-----	1	.2	-----	-----	-----	-----	-----	-----

TABLE 10.—*Mineralogy of layered and homogeneous pegmatites*—Continued

Pegmatite (pl. 21)	Wallrock, type	Pegmatite																				
		Relation to wallrock	Shape	Internal structure	Mineralogy																	
					Plagioclase		Perthite		Quartz		Muscovite		Garnet		Tourmaline		Lithiophilite-triphylite		Biotite		Beryl	
					Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)	Percent	Size (inch)
165	QBS, QMSS	C	L	H	36	0.4	28	8	28	0.7	7	0.8	Tr.	0.1	<1	0.4			Tr.	0.8		
166	QBS, QMSS	CD	Ir	L	37	.2	25	7	33	.3	4	.4	Tr.	.1	<1	.2						
176	QMS		Ir	L	32	.2	31	8	31	.5	5	.4	Tr.	.1	<1	.4						

feet thick, but a few have thicknesses of as much as 25 feet. Concordant relations with the country rock are more common for homogeneous than for layered pegmatites.

The major minerals in the homogenous pegmatites are plagioclase, quartz, perthite (some with graphic quartz), and muscovite. Minor accessory minerals are tourmaline, garnet, apatite, biotite, and beryl. The average mode of the homogeneous pegmatites is 36 percent plagioclase, 33 percent quartz, 25 percent perthite, 4 percent muscovite, 1 percent tourmaline, and 1 percent other minerals. Fairly large differences in mineral content of the different bodies are illustrated in table 10, and minor differences are common in different parts of single bodies. In particular, table 10 shows many modes that are unusually high in perthite, but these are mainly for small pegmatites that are not typical of the homogeneous pegmatites as a group. Perthite or plagioclase may form as much as 50 percent of a homogeneous pegmatite. The greatest concentrations of perthite, especially when it has graphic quartz, are along the hanging wall of gently dipping pegmatites. Graphic perthite is somewhat less abundant in homogeneous pegmatites than in layered ones. The plagioclase is mostly albite, as in the layered pegmatite, but more calcic plagioclase is found in very small stringers of pegmatite and in border zones that may have been contaminated by wallrocks. Some of the thin homogeneous bodies, mostly less than 3 feet thick, have abundant plagioclase and quartz but less perthite than do the larger bodies. Pegmatites that are poor in perthite tend to be rich in muscovite; a few have as much as 10 percent muscovite.

The quartz and feldspar of all the modes of layered and homogeneous pegmatites shown in table 10 were recalculated to total 100 percent and were plotted in the ternary quartz-albite-orthoclase diagram in figure 94. The anorthite content of the plagioclase was ignored because it is so low, and the plagioclase content of the perthite was disregarded because it would be largely offset if the K_2O in the muscovite were recalculated as orthoclase. The diagram indicates extensive overlap in the composition of the two types of pegmatite. Homogeneous pegmatites, however, tend to have more potassic feldspar and less plagioclase than layered pegmatites. On the other hand, the quartz-feldspar ratio of the two varieties of pegmatite is about the same.

The perthite and the graphic perthite in homogeneous pegmatite generally are coarse grained and in places they form crystals as much as 4 feet long. These crystals are subhedral to euhedral and are surrounded by a matrix of anhedral fine-grained quartz and plagioclase. In finer grained, more nearly graintoid homogeneous pegmatite, the perthite grains are as small as 1 inch. Plagioclase and quartz grains rarely average more than 0.4 inch in diameter in a homogeneous body.

Most of the accessory minerals are less than an inch across, but rarely tourmaline or muscovite occurs as crystals several inches in diameter.

Tourmaline, commonly having an euhedral form, is most abundant in the outer parts of the pegmatites, especially on the hanging wall side. Many of the tourmaline crystals are approximately perpendicular to the contact, and single crystals of tourmaline, perthite, or muscovite commonly increase in size inward from the contact. Rarely, tourmaline crystals near the contacts of a pegmatite body are partly or completely replaced by muscovite.

Homogeneous pegmatites in certain parts of the area have distinctive compositions or textures that suggest they are more closely related to one another than to more distant pegmatites. Many in the vicinity of the New York pegmatite (pl. 21), contain perthite in which the plagioclase lamellae weather to a darker red than does the enclosing microcline. In the southeastern part of the area and also in secs. 4 and 9, T. 4 S., R. 4 E., these pegmatites are finer grained and contain less perthite than elsewhere. In other small areas there are groups of homogeneous pegmatites that all contain more than the average amount of quartz, perthite, mica, or tourmaline. The many small pegmatites in sec. 22,

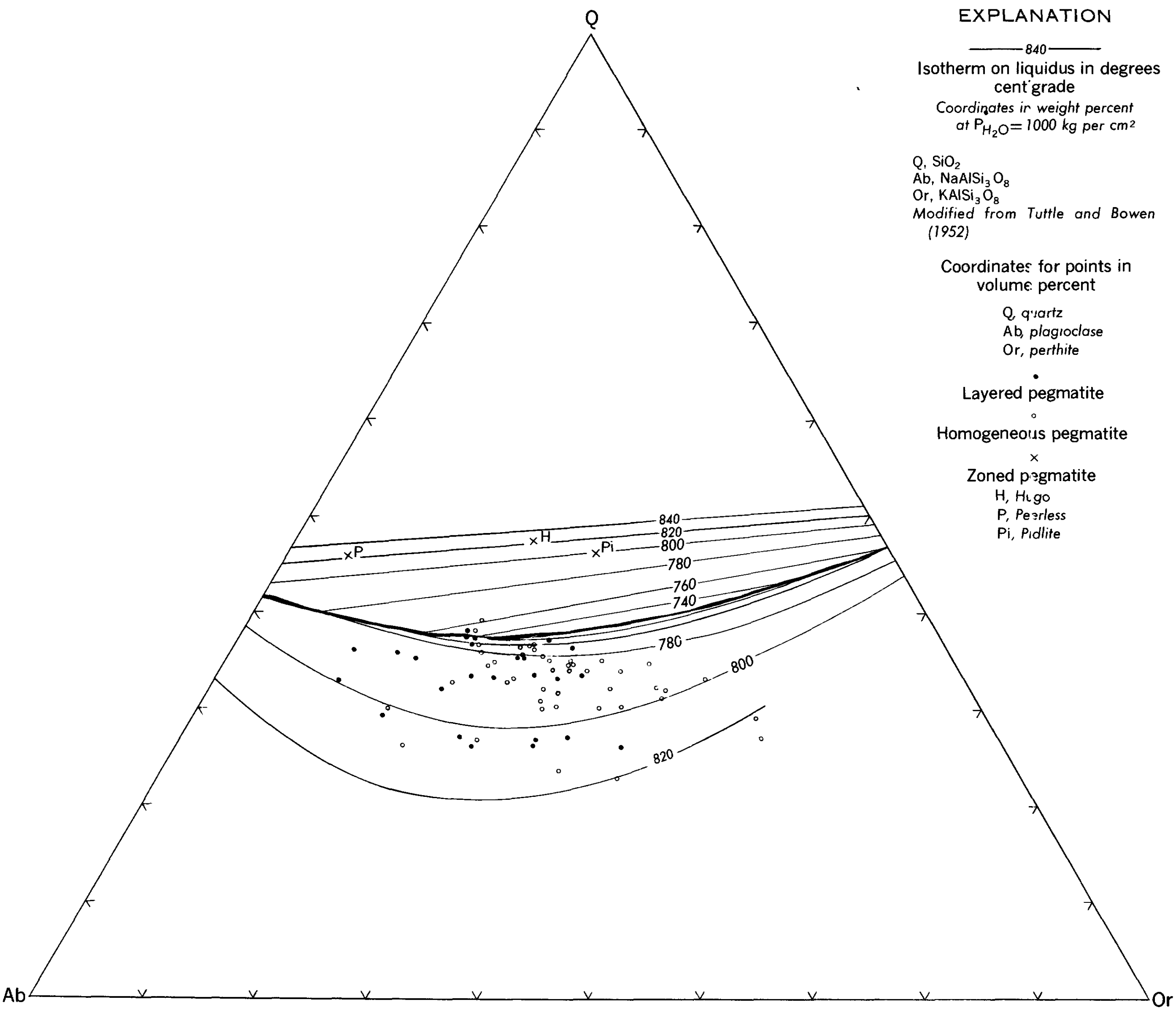

FIGURE 94.—Ternary diagram showing quartz-feldspar proportions in pegmatites as compared with the experimentally determined liquidus in the system $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2–H_2O.

T. 4 S., R. 4 E., are relatively rich in perthite, as the modes for pegmatites 155, 156, and 157 indicate (table 10). Plagioclase in many of the homogeneous bodies in secs. 2, 16, and 17, T. 4 S., R. 4 E., is bluish-gray and coarser grained than elsewhere. Fracture fillings in homogeneous bodies near Castle Rock (pl. 21) commonly contain lithiophilite-triphylite. Beryl is most common as an accessory in the western and northwestern parts of the quadrangle.

ZONED PEGMATITES

Zones are the only prominent internal units of zoned pegmatites, but there also may be a few fracture fillings, layers, and replacement bodies. These pegmatites correspond to the heterogeneous pegmatites of Johnston (1945, p. 1025) and to some of the complex pegmatites of Schaller (1925) and Landes (1933, p. 33–35). Page and others (1953) described many of the zoned pegmatites of the Black Hills.

Only 51, or about 2 percent, of the 2,200 pegmatites mapped in the Fourmile quadrangle are classed as zoned. In addition, 10 of the larger fracture filling bodies cutting layered pegmatites will be treated here as zoned pegmatites. Though the zoned bodies are widely distributed, they are recognizably more abundant in the northwestern part of the quadrangle, where in a belt extending 1½ miles eastward from the zero isopleth of figure 79, nearly 20 percent of the pegmatites are zoned. Zoned pegmatites that appear elsewhere on figure 79 are also on the fringes of pegmatite-rich areas. The dominant country rock in these areas is the Mayo formation, but amphibolite is also abundant in the belt of zoned pegmatites in the northwest part of the quadrangle and most of the pegmatites in amphibolite are zoned. In the Berne quadrangle the Bugtown formation also contains several zoned pegmatites near the zero isopleth. Apparently the zoning of pegmatites is more dependent on the areal distribution (on the fringes of pegmatite-rich areas) than on the type of country rock.

The zoned pegmatites in the Mayo formation are largely discordant, but those in the Bugtown formation in the Fourmile and Berne quadrangles are nearly all concordant with the dominant schistosity. Approximately 60 percent of all the zoned pegmatites in the Fourmile quadrangle are discordant, and half of these are partly or entirely in amphibolite.

The sizes and shapes of zoned bodies vary considerably, but none of the zoned bodies is as large as the largest layered or homogeneous pegmatites. The largest are the Warren Draw and the Tip Top pegmatites, and the smallest are too small to show on the map. Many zoned pegmatites are thickly tabular or oval or pipe shaped, and thus are in contrast to the pegmatites of tabular and irregular shape that are predominant among the layered and homogeneous bodies. Vlassov (1943) also recognized a somewhat similar relation of shape to zoning of pegmatite bodies.

Most of the zoned pegmatites in the quadrangle contain only border and wall zones and a core. Of the 51 zoned pegmatites, 7 contain intermediate zones distinctive enough to be mapped, and 1, the Helen Beryl, has 6 zones counting the border zone.

The modes of the principal zoned pegmatites (table 7) show that outer zones are generally enriched in plagioclase and muscovite, and that perthite is mainly in intermediate zones and cores. Quartz appears in all units, but it tends to be most abundant in inner zones. Lithium minerals are almost entirely in intermediate zones, cores, and fracture fillings that correspond in composition to inner zones. The Tin Mountain, Helen Beryl, and Warren Draw mines (pl. 21) have units containing more than 1.0 percent Li_2O, but even the lithium-rich units at these localities have a much smaller volume than the lithium-poor outer zones.

The greater size of outer zones, which are rich in plagioclase, also suggests that plagioclase is generally more abundant than perthite. Quartz is probably intermediate between plagioclase and perthite. The overall muscovite content of zoned pegmatites as a group is between 5 and 10 percent. Accurate estimates of the mineral content of a zoned pegmatite can be made only if the tonnage of each zone has been determined, and this can be done only after obtaining thorough knowledge of the structure of the pegmatite at depth. Data of this sort are presented below for the Peerless and Hugo pegmatites, Keystone, S. Dak., and the Pidlite pegmatite in New Mexico.

Mode, in percent

Pegmatite	Quartz	Plagioclase	Perthite and microcline	Muscovite and lithia mica	Lithium minerals	Remainder	Source
Peerless	40	43	5	10	0.3	1.7	Sheridan and others, (1957, table 11).
Hugo	41	27	19	7	1.8	4.2	Norton, J. J., written communication.
Pidlite	35.4	20.3	21.4	5.7	17.1	.1	Jahns (1953, table 4).

In a broad sense these three pegmatites are comparable with the Fourmile zoned pegmatites. Lithium minerals are more prominent in the Pidlite and Hugo pegmatites and potassic feldspar less prominent in the Peerless than in most of the zoned pegmatites of the Fourmile quadrangle. The quartz-feldspar proportions of these pegmatites as plotted on figure 94 may be taken as an approximation for the Fourmile area. Figure 94 indicates that these zoned pegmatites contain significantly more quartz than layered and homogeneous bodies and that they probably have about the same amount of plagioclase and less potassic feldspar. It should be noted, however, that zoned pegmatites are relatively rich in muscovite, which contains K_2O that would otherwise be in feldspar.

The zoned pegmatites contain more of the rarer minerals and therefore greater amounts of such minor elements as Be, Li, Cb, Ta, or Cs than other types of pegmatites. The total quantity of each of these minor elements in zoned pegmatites is so small that it may not be much greater than the amounts hidden in mineral solid solution in layered and homogeneous pegmatites.

MINERALOGY

The dominant minerals in all the pegmatites are plagioclase, quartz, perthite, and muscovite. The more common accessory minerals are tourmaline, garnet, microcline, and apatite. Other minerals are beryl, biotite, lithiophilite-triphylite, spodumene, amblygonite, lepidolite, pollucite, columbite-tantalite, microlite, tapiolite, cassiterite, and iron-magnesium phosphate minerals. Modes of all the large bodies of pegmatite and many of the smaller ones are presented in tables 7 and 10.

PLAGIOCLASE

Plagioclase is the most abundant mineral of most of the pegmatite, but in approximately 30 percent of the individual pegmatite bodies listed in table 10 it is subordinate to quartz or perthite. A similar statement would probably hold for pegmatites elsewhere in the Black Hills, though Higazy (1949, p. 557) emphasized the supposed predominance of potassic feldspar. Certainly there are no pegmatites in the Black Hills in which plagioclase is not a major constituent.

Though there are wide variations in the plagioclase content of the different pegmatites, it is probably most abundant in the layered bodies. It generally forms white to gray subhedral grains, typically about 0.2 inch in diameter.

Blocky crystals of plagioclase in the Rock Ridge pegmatite are as much as 12 inches across. White to bluish-gray plates of cleavelandite appear in many inner zones. Their average diameter is 0.8 inch but may be 4 inches or even more. Most of the pegmatites containing cleavelandite are in the northwestern and northcentral part of the area (fig. 95).

The minimum indices of refraction of cleavage fragments (X') and the apparent anorthite content of the plagioclase vary in a generally systematic manner throughout the area as is shown in figure 96. It was originally assumed that the change in anorthite content was absolute, but preliminary data of J. J. Norton (written communication, 1958) indicate that this may not be true. However, X' and possibly the anorthite content of the plagioclase do increase in areas containing more pegmatite. The apparent increase in the anorthite content seemingly continues to the northeast into areas of abundant granitic rocks; plagioclase from a few samples of the granite and pegmatite in the vicinity of Harney Peak (fig. 75) has an apparent anorthite content ranging from An_{11} to An_{30}.

Local variations in X' of the plagioclase are superimposed on the regional differences. Plagioclase from zoned pegmatites and fracture fillings generally has a smaller X' value on the cleavage fragments than that from nearby layered and homogeneous pegmatites. In some of the zoned pegmatites, plagioclase from inner zones has lower X' values than that from the wall zones, but differences in the indices of more than .003 are rare. In others there are no noticeable differences. There is no appreciable difference in the refractive indices of plagioclase from layered and homogeneous pegmatites in any small part of the area. Very fine grained pegmatite that is rich in microcline has plagioclase whose X' values are less than those of cleavage fragments of plagioclase in pegmatite containing little or no microcline. The X' values of plagioclase in the lamellae from large perthite crystals are generally about .002 lower than those of plagioclase in the surrounding groundmass. Very fine grained pegmatite locally contains relatively large rounded grains of plagioclase whose X' indices are slightly higher than those of the smaller grains in the groundmass.

QUARTZ

Most quartz of the pegmatites either forms white, cloudy anhedral masses or graphic intergrowths in perthite. Quartz in the zoned pegmatites is coarser grained than that in the other pegmatites. Large single crystals, several feet or more in diameter, may occur in massive quartz cores or in perthite-quartz zones. Many perthite-rich zones contain fine- to medium-grained quartz associated with plagioclase in a matrix around large perthite crystals. In the very fine grained pegmatite much of the quartz occurs either as rounded blebs or as an interstitial filling between grains of feldspar.

PERTHITE AND MICROCLINE

Perthite, much of which contains graphic quartz, is the third most abundant mineral in the pegmatite. Only a very few thin zoned bodies lack this mineral entirely. It occurs as white, cream, flesh, or pink to rose-colored subhedral to euhedral crystals. At the New York mine and in several nearby pegmatites, perthite occurs as flesh-colored crystals in which plagioclase lamellae weather to a rusty red color, darker than the enclosing microcline. Elsewhere in the quadrangle, the plagioclase is lighter colored than the microcline. The perthite typically forms large phenocrysts set in a matrix of quartz and plagioclase. It is more nearly euhedral than the quartz and plagioclase and is generally much coarser grained than the other pegmatite minerals. Where intergrown with massive quartz, perthite commonly has euhedral faces; these crystals are especially noticeable in some fracture fillings. The average size of the perthite crystals in all pegmatite is 5 inches; single crystals 10 feet long occur at the Tin Mountain pegmatite.

Plagioclase lamellae in large perthite crystals are approximately uniform in size and distribution in nearly all pegmatites and all types of units in the pegmatites; the ratio of plagioclase to microcline is about 1 : 5. The major exception to this constant ratio is in the quartz core of the New York pegmatite where there are a few relatively nonperthitic microcline crystals. Perthite is also not apparent where very fine or fine-grained pegmatite contains microcline.

In most of the homogeneous and layered pegmatites the perthite has graphic intergrowths with quartz. Nongraphic perthite is more common in the western part of the area than in the eastern; fracture fillings in pegmatites containing graphic perthite will also consist of nongraphic perthite.

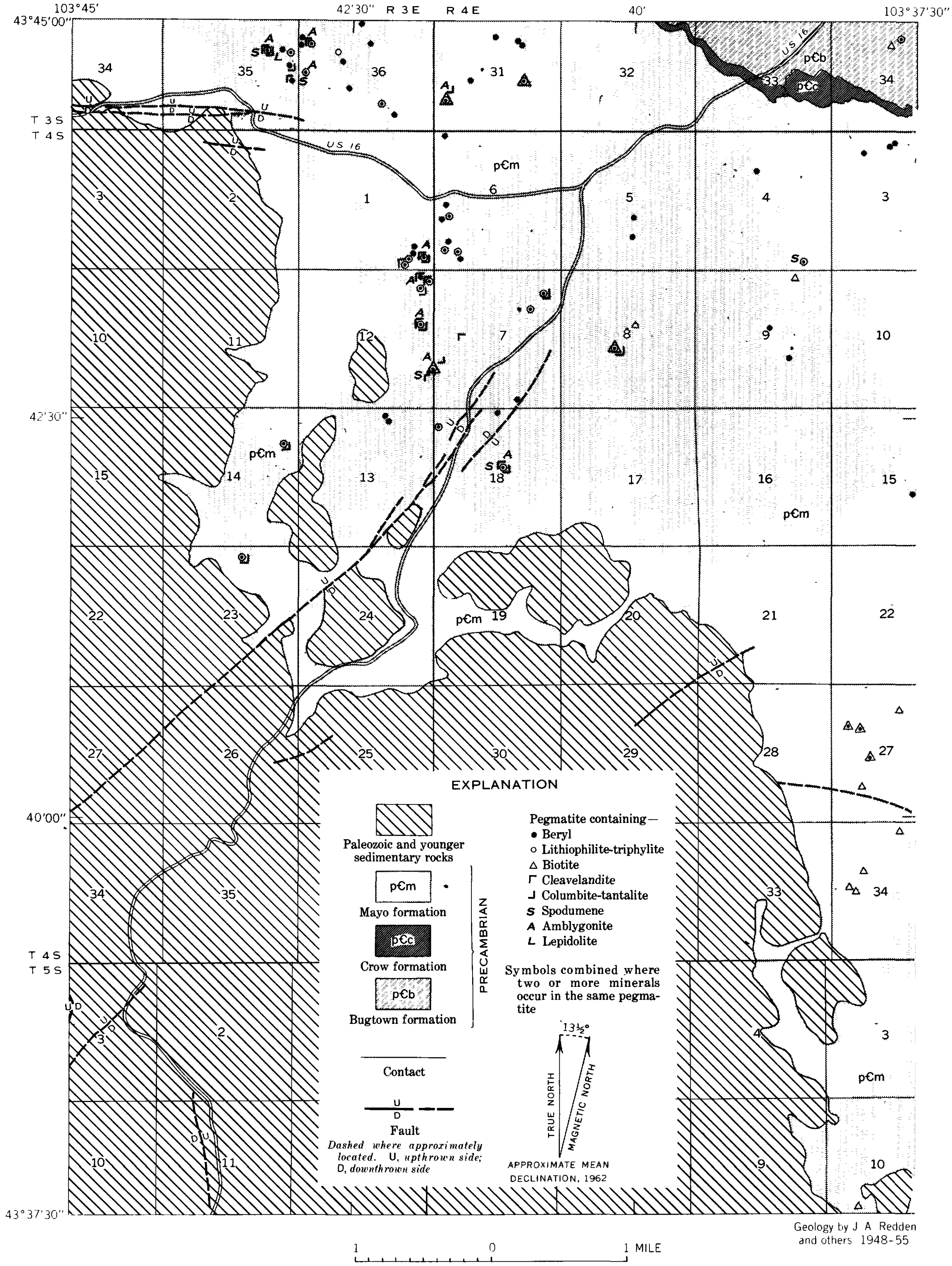

FIGURE 95.—Map showing distribution of pegmatites containing the rarer minerals in the Fourmile quadrangle.

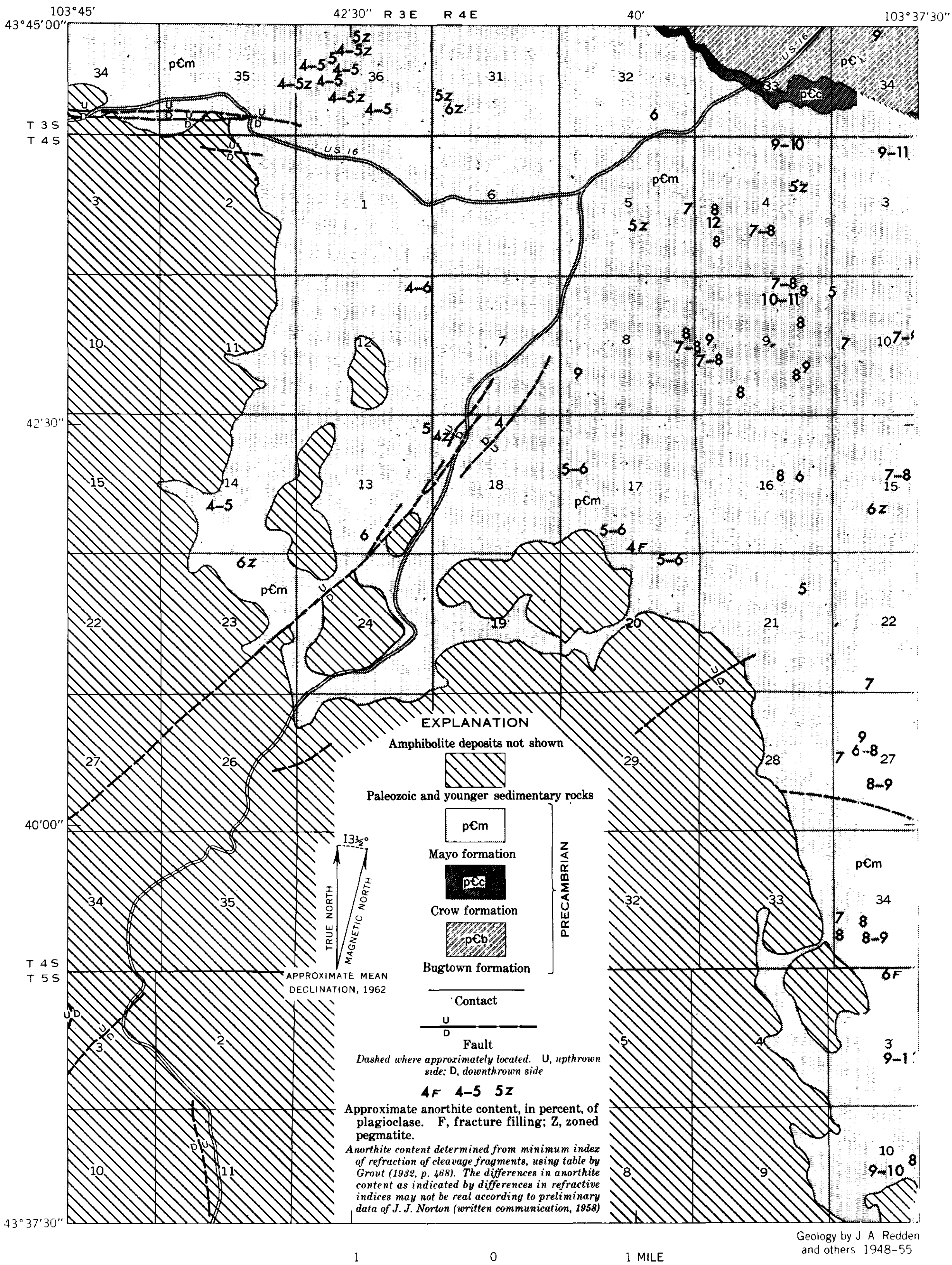

FIGURE 96.—Map showing apparent anorthite content of plagioclase from selected pegmatites in the Fourmile quadrangle, as determined by refractive indices.

The quartz rods in the graphic perthite ordinarily have a variety of orientations in different parts of a single perthite crystal, but those in any small part of the crystal are approximately parallel. In the universal stage study of graphic perthite in which phantom crystal faces were marked by inclusions of line rock (p. 233) the only thin section in which there was no preferred orientation of quartz rods was one which was cut along a (001) face of the microcline. In each of the other specimens the quartz rods were megascopically nearly perpendicular to the crystal face, and the optic axes of the rods and grains of quartz had pronounced maxima parallel or nearly parallel to the crystal face. In 4 specimens parallel to $(\bar{2}01)$ crystal faces of microcline, the optic axes of the quartz had maxima at an angle of about 63° to the *c*-axis of the microcline; in 2 specimens parallel to $(\bar{1}10)$ the optic axes of quartz had maxima at about 42° to the *c*-axis of the microcline. This relation seems best interpreted as meaning that the quartz grew contemporaneously with the microcline and was somewhat controlled by the microcline structure.

MUSCOVITE

Muscovite is present in all pegmatites, either as an accessory or major mineral, and generally as flat or wedge-shaped subhedral deep-ruby to nearly colorless crystals. At the Tin Mountain mine, lithium-bearing muscovite occurs as yellowish-green round aggregates. Although the average size of muscovite crystals is only about 0.6 inch, muscovite forms books as much as 2 feet across in zoned pegmatites like that in the New York mine. Many of the muscovite books contain inclusions of tourmaline, feldspar, and other minerals. The platy crystals of muscovite in the outer few inches or few feet of pegmatite bodies are oriented subperpendicular to the contacts between pegmatite and wallrock. Muscovite is concentrated in the outer zones of the zoned pegmatites and in the border zone of all pegmatites.

OTHER MINERALS

Tourmaline can be found in virtually all the pegmatites as shiny black subhedral to euhedral crystals. Green tourmaline is associated with lithium minerals in the Helen Beryl and Tin Mountain pegmatites. In most of the pegmatites tourmaline is most abundant near the contacts, especially along the hanging wall. It appears as prismatic crystals that commonly are nearly perpendicular to the contact. Most of the tourmaline crystals are 2 to 4 inches long and 0.5 to 1 inch in diameter, but very fine grained needles less than 1 mm in diameter are common in line rock. Many of the larger masses of tourmaline are aggregates consisting of several crystals having nearly parallel *c*-axes.

Reddish-cinnamon-colored subhedral to euhedral garnet is in most of the pegmatites. Crystals generally range from 1 mm to 5 cm in diameter, although aggregates of garnet crystals as much as 5 inches across are found in some zoned pegmatites. Many of the larger crystals have abundant fractures.

Apatite is an accessory mineral in nearly all the pegmatite. It generally occurs as small (average 0.1 inch) blue-green subhedral crystals, commonly in the outer part of individual bodies. Coarser grained apatite is found only in the larger zoned pegmatites. Bright-blue apatite is found in the Tin Mountain and Helen Beryl pegmatites.

Beryl is most abundant in outer zones of zoned pegmatites where quartz, plagioclase, and muscovite are the dominant minerals. Zones of this type in the Tin Mountain and Helen Beryl pegmatites contain the largest tonnage of beryl in the district. Locally, beryl is a common accessory mineral in all three types of pegmatites. Beryl is commonly found in the fracture fillings of homogeneous and layered pegmatites, but rarely in the host pegmatites. Beryl has a regional distribution and is most common in pegmatites in the northwestern part of the quadrangle (fig. 95).

Most of the beryl is pale green and somewhat milky, although rarely it occurs as small colorless crystals. In pegmatites rich in lithium minerals, most of the beryl is white to cream colored, and generally has higher refractive indices than the green or colorless beryl. The indices of specimens of beryl from 24 different pegmatites in the area range from 1.576 to 1.588, but no pattern was recognized in the areal distribution.

Long, thin, lathlike crystals of shiny black biotite were observed in 19 pegmatites, mostly in the southeastern part of the area (fig. 95). Much of the biotite is in large perthite-rich fracture fillings. At the Helen Beryl and Tip Top mines, biotite laths are as much as 4 feet long, 4 inches wide, and 1 inch thick. Muscovite ordinarily is intergrown with the biotite.

Lithiophilite-triphylite or its alteration products was found in 25 different pegmatites in the northwestern part of the area (fig. 95). The distribution of these minerals coincides closely with that of zoned pegmatite and of beryl (figs. 79, 95). The lithiophilite-triphylite generally occurs as irregular masses several inches in diameter, but at the Rainbow No. 4 pegmatite, it is in euhedral crystals that are diamond shaped in cross section. Fresh lithiophilite-triphylite is dark cinnamon brown to bluish green, but alters readily to brownish-purple or black products, probably chiefly heterosite or purpurite.

Spodumene has been found in only five pegmatites, all of which also have beryl and lithiophilite (fig. 95).

Only the Tin Mountain and Helen Beryl pegmatites contain a significant quantity of spodumene. It occurs in intermediate zones and cores as lathlike crystals as much as 15 feet in length. Most of the spodumene is chalky white, but some in the Helen Beryl pegmatite is green. One large spodumene crystal at the Tin Mountain mine had a clear, colorless interior surrounded by white spodumene. A greenish-gray claylike alteration product is at the outer part of many spodumene crystals, and in places it replaces entire crystals.

Amblygonite has a somewhat wider distribution than spodumene. It was found in 11 zoned pegmatites (fig. 95), largely in the northwestern part of the area. It is gray to white and generally forms equidimensional crystals, an inch or a few inches in diameter. It is most abundant in the Tin Mountain pegmatite, where crystals several feet across are exposed.

Lepidolite, microlite, and pollucite are known only in the Tin Mountain pegmatite. The lepidolite is pale lavender to colorless and occurs mainly as fine-grained aggregates. Some of the plates are curved, and aggregates of these have a hemispherical form. Small waxy brownish-green subhedral equidimensional crystals of radioactive microlite associated with cleavelandite, quartz, and lepidolite are at the edges of some of the large crystals of spodumene in the inner part of the pegmatite. Although pollucite was mined from an inner zone some years ago (Page and others, 1953, p. 20), none was found in place or on the mine dumps.

Columbite-tantalite was found in 12 zoned pegmatites in the northwestern part of the area (fig. 95). Most of it is in shiny black platy crystals a few inches long and a fraction of an inch thick. In the Tin Mountain and Helen Beryl pegmatites it is in equidimensional crystals associated with quartz and cleavelandite.

One crystal of tapiolite was reported from zoned pegmatite 10 (pl. 21) by Mrs. Gladys Wells of Custer (oral communication, 1952). The crystal, about 0.3 inch in diameter, was embedded in a book of muscovite from the mica-rich wall zone.

PALEOZOIC AND YOUNGER ROCKS

DEADWOOD FORMATION

The Deadwood formation unconformably overlies the metamorphic and igneous rocks along an irregular northwest-trending line, approximately connecting the southeast and northwest corners of the quadrangle. The formation crops out around the entire Black Hills and its age is late Cambrian. (Darton and Paige, 1925, p. 7.)

The formation is very poorly exposed; it forms moderately steep slopes covered by soil and talus that also extend over the adjacent Precambrian rocks. The lower contact, though rarely exposed, is marked by a flattening of the slope. A soft weathered zone in the Precambrian rocks extends to a depth of as much as 50 feet below the unconformity; it is characterized by bleaching, iron staining, and kaolinization of the plagioclase of the Precambrian rocks similar to that found in areas of semitropical or tropical climates. Locally, as along the Pleasant Valley road in the S½ sec. 24, T. 4 S., R. 3 E., fresh Precambrian rock crops out at the contact with the Deadwood, and there is no weathered zone or change in slope.

In several places in the northeast part of the quadrangle the unconformity may not have been far from the present surface of Precambrian rocks. A 5-foot block of conglomerate from the basal part of the Deadwood formation is exposed just south of U.S. Highway 16 in the center of the S½ sec. 33, T. 3 S., R. 4 E. The bedding in the block is nearly horizontal, and it seems likely that the rock is virtually in place. Furthermore, the surrounding topography is relatively flat, and the outcrops of pegmatite do not rise above the schist, as they do elsewhere. It is likely that this truncation is a relict of Precambrian topography. Further south, along upper Hay Creek, many boulders of Paleozoic rocks in gravel deposits are so large that it is unlikely they have been transported very far or reworked.

The contact of the Deadwood formation with the overlying Englewood limestone is concealed throughout the quadrangle, but it generally can be located within a few feet vertically.

The thickness of the Deadwood formation in the Fourmile quadrangle ranges from about 90 feet near the southeast corner to 170 feet in the northwest corner. The thickness decreases to as little as 4 feet farther southeast and increases to approximately 500 feet in the extreme northern part of the Black Hills (Darton and Paige, 1925, p. 6).

The Deadwood formation as seen in its few exposures is mostly reddish brown and consists predominantly of conglomerate, sandstone, glauconitic sandstone, and quartzite. A few feet of greenish shale and siltstone occur near the middle and top of the formation, but are exposed rarely. The general sequence from bottom to top is conglomerate and conglomeratic sandstone, sandstone, quartzite, glauconitic sandstone, siltstone and shale, and a coarse-grained upper sandstone.

The basal conglomerate contains sand and pebbles as much as an inch in diameter that were largely derived from nearby Precambrian quartz veins and pegmatite. Some of the pebbles are very angular and others are moderately rounded.

Grains of the upper sandstone are extremely well rounded. The best exposure of this unit in the quadrangle is in the small domal window in sec. 35 along Pleasant Valley. Good exposures have been opened up by recent trenching just west of the quadrangle in Layton Draw. Most of the area contains no exposures of the upper sandstone, and even where exposed it is rarely thicker than 5 feet. Apparently its thickness increases to the north in the Berne quadrangle where it is known to be as much as 20 feet thick. Fossils were not found in the sandstone nor in the few feet of silt beds underlying it. The lower part of the Ordovician section in the northern Black Hills (Furnish, Barragy, and Miller, 1936, p. 1338) is lithologically somewhat similar to these rocks, and it is possible that they belong to the Ordovician rather than to the upper part of the Deadwood formation. The excellent rounding of the sand grains suggests that this unit may be reworked material from the Deadwood formation.

Two of the best exposed sections of the Deadwood formation are described below:

Section of Deadwood formation in the SE1/4 sec. 23, T. 4 S., R. 3 E., Custer County, S. Dak.

Mississippian:
Englewood limestone: Lavender and purple pebbly limestone.
Unconformity.

Cambrian: Deadwood formation:	Thickness (feet)	Cumulative thickness (feet)
Sandstone, white, coarse-grained, well-rounded, in part crossbedded; conglomeratic at base	3	3
Siltstone, pale pinkish-lavender, tan, and cream; thin-bedded, laminated, calcareous	12	15
Cover	2	17
Sandstone, pale greenish-gray, medium grained, thin-bedded, very glauconitic; argillaceous laminae in sand beds; some carbonate cement	10	27
Cover	16	43
Sandstone, variegated, green to red, medium-grained, thinly laminated and bedded; glauconitic in part; cement partly carbonate	4	47
Cover; probably glauconitic sandstone and possibly shale beds	15	62
Sandstone, reddish- to greenish-brown, medium-grained, medium-bedded; in part a quartzite; slightly glauconitic	5	67
Cover	5	72
Sandstone, greenish- to brownish-gray, medium-grained, medium-bedded; interbedded glauconitic sandstone and sandstone beds; beds contain many small shiny calcareous brachiopods; small round iron-stained concretions are common	10	82
Cover	2	84

Section of Deadwood formation in the SE 1/4 sec. 23, T. 4 S., R. 3 E., Custer County, S. Dak.—Continued

Cambrian—Continued Deadwood formation—Continued	Thickness (feet)	Cumulative thickness (feet)
Sandstone, reddish-brown, coarse-grained, medium-bedded; poorly rounded quartz grains; conglomeratic near top	3	87
Cover	13	100

Unconformity.
Precambrian:
Schist and pegmatite.

Section of Deadwood formation in NE1/4 sec. 33, T. 4 S., R. 4 E., Custer County, S. Dak.

Mississippian:
Englewood limestone: lavender and purple pebbly limestone.
Unconformity.

Cambrian: Deadwood formation:	Thickness (feet)	Cumulative thickness (feet)
Cover	78(?)	78(?)
Quartzite, light- to brownish-tan, fine-grained, medium-bedded; well cemented with silica; contains numerous fucoid markings on bedding partings	4	82(?)
Sandstone, light-tan to brown, coarse-grained, medium-bedded; a few pale green glauconitic blebs associated with flakes of detrital mica; some half-inch thick spherical iron-stained concretions in white sandy matrix	27	109(?)
Conglomerate, friable, white to pink, fine-grained, medium-bedded; pebbles of quartz as large as 0.5 inch	3	112(?)
Cover	9	121(?)
Quartzite, light-tan to cream, fine-grained, medium-bedded; moderately well cemented	6	127(?)
Conglomerate, tan to brown, fine-grained, loosely cemented; subrounded to angular pebbles as large as 0.3 inch	4	131(?)

Unconformity.
Precambrian:
Schist and pegmatite.

ENGLEWOOD LIMESTONE

The thin Englewood limestone of Early Mississippian age (Darton and Paige, 1925, p. 7) overlies the Deadwood formation through the entire length of the quadrangle. Although somewhat better exposed than the Deadwood formation, it is generally covered by soil and blocks of the overlying Pahasapa limestone. The contact with the Deadwood formation is not exposed, but is marked by a steepening of slope in the more resistant limestone. Even though no rocks of Ordovician to Devonian age were recognized, there is no apparent discordance between the Deadwood and the Englewood.

The thickness of the Englewood limestone ranges from approximately 40 to 55 feet and averages 45 feet.

The thickness varies locally and does not change systematically throughout the area.

The exposed Englewood consists of a uniform thin-bedded pebbly to slabby very fossiliferous impure limestone interbedded with some dolomite. The weathered limestone is lavender to mauve, whereas fresh rock is somewhat lighter. The pebbly appearance is caused by the selective weathering of the limestone around fossils and iron-rich concretions. These "pebbles" are 0.5 to 3 inches in diameter and are slightly flattened parallel to bedding. Laminations in the bedding planes of the limestone also tend to curve around the fossils and concretions. Horn corals are the predominant fossils; brachiopods are also abundant, and crinoid stems are distributed throughout the limestone. The interiors of many of the fossils are replaced by white calcite.

PAHASAPA LIMESTONE

Most of the southwestern half of the Fourmile quadrangle is underlain by the Pahasapa limestone. The limestone dips gently west and forms cuestas that face east and northeast; vertical and near vertical cliffs are as much as 70 feet high. The Pahasapa also caps small outliers of Paleozoic rocks west of Pleasant Valley (pl. 21). The general topography formed on the Pahasapa is rugged, and valleys have steep walls. According to Darton and Paige (1925, p. 8), the age of the Pahasapa limestone is Early Mississippian.

The contact of the Pahasapa limestone with the Englewood limestone is exposed in a few places in the quadrangle. Probably the best exposure is in the SE¼ sec. 2, T. 4 S., R. 3 E., where the contact is conformable and is marked by an abrupt change from a thin- and medium-bedded mauve limestone characteristic of the underlying Englewood limestone to a thick-bedded pale-buff limestone of the Pahasapa.

The Pahasapa consists of a thick-bedded pale-buff to dove-colored dense massive relatively pure limestone. Beds generally range from 1 to 8 feet thick. Small-scale crossbedding was found in only one exposure. The lower 60 to 80 feet of the limestone is the very massive cliff-forming limestone of the cuestas. The middle and upper parts are uniform except for discontinuous beds and lenses of gray-brown waxy chert bed in the upper half of the formation. The lowest chert is about 180 feet above the base of the Pahasapa, and others are approximately at 220, 240, 265, and 280 feet above the base. Other cherty layers higher in the formation were noted, but could not be traced along strike. The upper part of the limestone is poorly exposed and less resistant than the remainder of the formation. Where exposed it commonly is broken and contains solution cavities. Veins and pods of white calcite locally cross the beds. Fossils are mainly brachiopods and crinoid stems, which are abundant in beds throughout the formation.

The upper part of the Pahasapa limestone is locally cavernous. Though large caves have been found in the Black Hills area, the largest known in this quadrangle are only a few tens of feet in maximum dimension. Along the lower part of Pleasant Valley the normally massive limestone has been somewhat altered and iron stained. Outcrops are rubbly in appearance and bedding is not visible. Layers of broken chert as much as 6 feet thick cap some of these altered exposures. Locally fragments of chert that appear to have come from a number of separate chert layers are concentrated in a single thick unit. These accumulations of fragmental chert and the broken and pitted appearance of the limestone are interpreted as resulting from widespread solution of the upper part of the Pahasapa limestone. Small flexures are characteristic of the upper contact of the Pahasapa, but they do not appear at the lower contact (fig. 97). The axes of these flexures apparently follow solution channels. They trend northwest, north, northeast, and east, and are parallel to the dominant joints.

Great local variations in the thickness of the Pahasapa are a consequence of solution activity and erosion along the unconformable upper contact. The limestone may be as little as 240 feet thick near the southwest corner of the quadrangle, and as much as 450 feet thick a few miles to the east or north (fig. 97). A distinctive fossiliferous chert-bearing bed in the SE¼ sec. 27, T. 4 S., R. 3 E., apparently drops 200 feet stratigraphically in less than a quarter of a mile, and the chert at the lower level is in fragments typical of the solution areas. The average thickness of the limestone where the effects of solution are not evident is between 400 and 450 feet.

Elsewhere in the southern Black Hills the limestone is extremely variable in thickness. Rothrock (1949, p. 28) reported that in wells drilled a few miles south of the Black Hills, the thickness ranges from 165 to 465 feet. One well on the east side of Fall River County, just south of Custer County, contained only a few feet of Pahasapa limestone. In Kansas and surrounding states, pre-Pennsylvanian erosion removed parts or all of the Mississippian as well as older rocks, and also formed a very porous zone in the limestone below the unconformity (Wallace, 1939, p. 19–41). The porous zone in the upper part of the Pahasapa limestone is probably associated with this same unconformity. Variation in the amount of solution of the limestone after deposition of overlying rocks may, however, have been responsible for local extreme differences in thick-

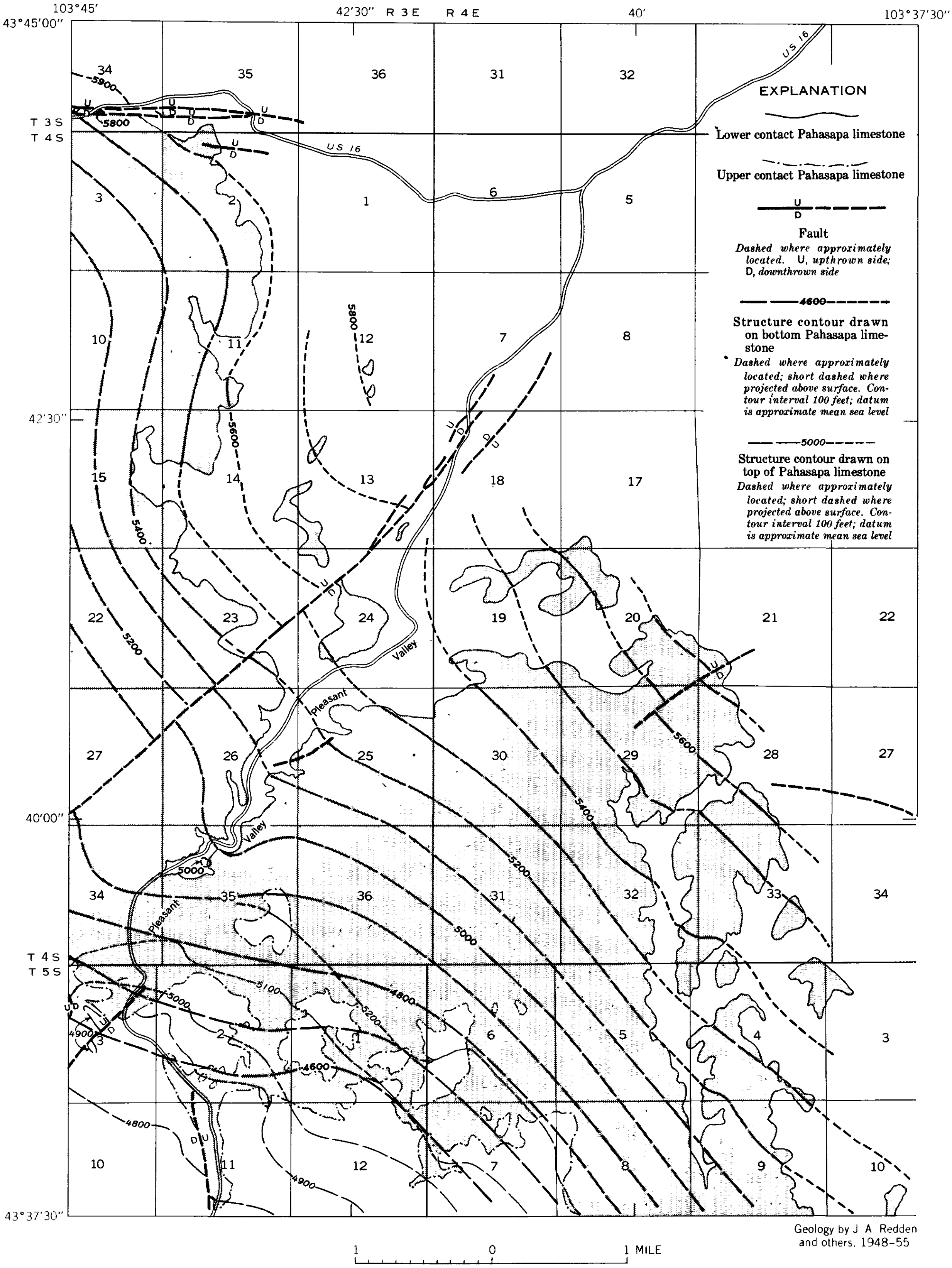

FIGURE 97.—Structure contours on top and bottom of Pahasapa limestone.

ness. This is discussed more fully in the description of the lower contact of the overlying Minnelusa sandstone, below.

MINNELUSA SANDSTONE

The Minnelusa sandstone of Pennsylvanian age is exposed in the extreme southwestern part of the quadrangle. It consists of poorly exposed nonresistant sandstone, silt, marl, and limestone. The present topography on the Minnelusa is generally very gentle in contrast to the steep valleys cut into the underlying Pahasapa limestone.

The lower contact is unconformable and apparently follows a surface of considerable relief. Nowhere is the contact exposed, but in several places it can be located within a few feet vertically. Generally, beds above and below the contact are parallel or nearly so. The evidence for the unconformity is the local stratigraphic variation in the position of the contact and the changes in thickness of the lower sandstone unit of the Minnelusa. This basal sand ranges from about 10 feet to possibly as much as 80 feet in thickness, and appears to have been deposited over a surface having a relief of at least 50 feet. At this same contact in the Hartville uplift in Wyoming, the Hartville formation (equivalent to Minnelusa) is deposited over a karst surface on the Guernsey limestone (equivalent to Pahasapa) (N. M. Denson, oral communication, 1956).

The lower part of the Minnelusa sandstone is generally not brecciated or broken above areas of solution and collapse in the Pahasapa, and, though fragments of the basal Minnelusa sandstone are embedded in the Pahasapa limestone near the south contact of the large outlier just east of Long Draw (pl. 21), there is no other evidence of extensive collapse of cave roofs. The subsidence probably was gradual and similar to that postulated by Bretz (1950, p. 830) as the origin of filled sinks in Missouri. The solution must have been in large part prior to the deposition of Oligocene rocks because these locally cover the deformed basal contact of the Minnelusa sandstone along Pleasant Valley. No evidence was found for the lowering of the Oligocene rocks.

The lower contact also has been modified by solution activity after burial. Part of the evidence for this is that the location of topographically low areas on the contact do not necessarily coincide with the location of thick sections of the basal sandstone, as would be expected if the low areas were at the original erosion surface. Additional evidence is that in places the dip of the Minnelusa is controlled by structural features formed in the underlying rocks by solutions. Along Pleasant Valley, for example, the contact of the Minnelusa with the Pahasapa limestone dips inward toward the valley edges (fig. 97), and beds in the overlying parts of the formation dip so steeply that they obviously were warped after initial deposition. Many of the more steeply dipping parts of the lower contact have flexures that appear to be controlled by solution along joints in the underlying limestone. Flexures in the lower contact form a trough along Pleasant Valley (fig. 97).

The Minnelusa has a total thickness of about 850 feet in the quadrangle south of the Fourmile quadrangle (Garland, Gott oral communication, 1957), but only about 350 feet is exposed in the Fourmile quadrangle. The exposed part has been divided into 3 main units on plate 21. These are a lower sandstone unit, a middle unit of limestone and chert, and an upper unit of sandstone, siltstone, and thin beds of limestone.

The lower sandstone unit is reddish brown to white and medium grained. Its thickness in many places is difficult to estimate because the relief on the underlying Pahasapa limestone is unknown. On the west side of Pleasant Valley (pl. 21, section *B–B'*) it is at least 50 feet thick and may be nearly 80 feet thick. Southeast of Long Draw the sandstone in general thins and locally only a few tens of feet of reddish soil mark the contact. In some poorly exposed areas, the unit may have pinched out completely.

Lenses of shale occur in the lower part of the unit and distinctive siliceous concretions are characteristic of the upper part of the unit. These concretions range from 6 inches to 4 feet in diameter and are generally oblate and concentrically banded. Concretions of this same type have been found in this unit in the Fanny Peak quadrangle, about 20 miles west-northwest of the Fourmile quadrangle (D. A. Brobst, oral communication, 1957), and at the south end of the Argyle quadrangle, which borders the Fourmile quadrangle to the south. Darton and Paige (1925, p. 9) reported that these concretions are characteristic of the lower sandstone unit elsewhere in the Black Hills.

This basal sandstone unit is apparently equivalent to the Fairbank formation (Bates, 1955, p. 1990–1993) and division VI of the Hartville formation (Denson and Botinelly, 1949).

The limestone and chert unit above the basal sandstone consists predominantly of pale-lavender to bluish-gray limestone containing numerous lenses and fragments of multicolored chert and a few beds of calcareous sandstone. It ranges in thickness from 20 to about 80 feet. The thicker sections consist of massive limestone, but most of the limestone in the thinner sections has been removed by solution and the unit consists mainly of beds of broken chert that are several feet thick.

Some of the more massive beds in the limestone cannot be distinguished from beds of the Pahasapa, but in a few exposures the limestone is distinctly granular and coarse-grained. In general, multicolored chert that contrasts with the brown and gray chert of the Pahasapa is diagnostic of this limestone unit.

The remaining exposed part of the Minnelusa consists of sandstone, siltstone, and thin limestone and dolomite beds. The general character of these beds is evident in the following generalized section:

Generalized stratigraphic section of the lower part of the Minnelusa sandstone (secs. 3, 10 and 11, T. 5 S., R. 4 E.)

	Thickness (feet)	Cumulative thickness (feet)
Pennsylvanian:		
Minnelusa sandstone:		
Sandstone, siltstone, and limestone unit:		
Siltstone and sandstone, interbedded, yellowish-brown to cream-colored, medium- to thick-bedded; cementing material slightly calcareous; a few thin calcareous and fossiliferous beds; angular chert fragments common in lower part of interval	65	193–258
Sandstone, tan, fine-grained, thick-bedded; a few feet of thinly bedded very impure fossiliferous limestone locally at bottom of interval	22	171–193
Sandstone, dark-brown, medium-grained; some crossbedding; largely calcareous cement and a clay-rich matrix	6	165–171
Limestone, grayish-tan to pale-buff, thin-bedded, very fossiliferous; rich in chert beds and lenses; some beds near bottom of interval sandy and siliceous	12	153–165
Sandstone, grayish-brown, medium-grained, thick-bedded; contains thin chert beds and lenses; locally a few quartzite beds	28	125–153
Conglomerate, tan, medium-bedded; pebbles 0.3 to 0.8 inch diameter, well rounded; composed of limestone and chert; matrix is calcareous siltstone and sandstone	5	120–125
Sandstone, light-colored, fine-grained, thick bedded; locally well cemented with silica and transitional to quartzite; elsewhere has calcareous cement	35	85–120
Limestone and chert unit:		
Limestone, gray, hard, fine-grained; cherty; contains about 20 percent gray, tan, white, red, and mauve chert in lenses and fragments; fossiliferous; locally extremely variable in appearance	5	80–85

*Generalized stratigraphic section of the lower part of the Minnelusa sandstone (secs. 3, 10 and 11, T. 5 S., R. 4 E.)—*Continued

	Thickness (feet)	Cumulative thickness (feet)
Pennsylvanian—Continued		
Minnelusa sandstone—Continued		
Limestone—Continued		
Limestone, bluish-gray and pale-lavender, fine-grained, thin- to massive-bedded; chert lenses and beds common; locally recrystallized into a medium-grained white limestone; section only partly exposed; thickness ranges from 20 to 60 feet	30	50–80
Basal sandstone unit:		
Sandstone, dusky-red to brownish-red, medium- to coarse-grained, medium-bedded to massive; locally exposures are entirely white or mottled red and white; section poorly exposed near bottom; large siliceous concretions in upper part part of sandstone; probably contains some maroon shale layers; thickness ranges from 15 to 80(?) feet	50	0–50
Total thickness (approximate)	258	
Mississippian:		
Pahasapa limestone: Light-gray, massive, limestone	--	----

An additional 100 feet or more of the Minnelusa formation lies above this section in the extreme southwest corner of the quadrangle. There are virtually no exposures, but float indicates that this part of the formation is composed largely of siltstone and sandstone. The total thickness of the Minnelusa within the quadrangle may be as much as 350 to 400 feet.

Fossils are not very abundant in the lower part of the Minnelusa. Darton and Paige (1925, p. 9) listed seven different species of probable Pennsylvanian age. The limy beds in the sandstone, siltstone, and limestone unit, where exposed in a cliff west of Pleasant Valley and just south of the quadrangle, contain numerous fossils of Pennsylvanian age. Specimens were identified by Helen Duncan, MacKenzie Gordon Jr., and L. G. Henbest (written communication, 1956) as *Chaetetes milleporaceus* Milne-Edwards and Haime, *Linoproductus* sp., *Composita subtilita* (Hall), several species of *Fusulinella*, and one of *Fusulina* (or possibly a late form of *Fusulinella*). The *Linoproductus* is somewhat similar to *L. nodosus* (Newberry).

Fossils were not found in the basal sandstone unit nor were there any diagnostic fossils in the limestone and chert unit.

WHITE RIVER GROUP

The Black Hills contains scattered deposits of sand, gravel, and volcanic ash that were assigned originally by Darton to the White River formation of Oligocene age (Darton and Paige, 1925, p. 15). Some of these deposits were dated by fossils, but many others, consisting largely of sand and gravel, can be assigned to the White River only on the basis of the similarity of their lithology and physiographic position. Darton (1951) later placed these deposits in the White River group, and that usage is followed in this study.

Several of these deposits of poorly exposed sand, gravel, and volcanic ash in and along Pleasant Valley were mapped by Darton and Paige (1925, geologic map of central Black Hills region). Many additional sand and gravel deposits of the same type are shown on plate 21 as White River group. Some of these small deposits may be younger terrace gravel consisting largely of reworked White River material. This is especially true of the upper few feet of such deposits.

The sand and gravel of the White River are mostly along the west side of Pleasant Valley in an old stream channel that is approximately parallel to the present course of Pleasant Valley Creek. The bottom of this old stream channel is about 100 feet above the present valley floor but cannot be accurately located because exposures are poor. The alluvial deposits in this channel are locally as much as 80 feet thick and may be as much as 140 feet thick. The channel apparently follows a solution zone in the Pahasapa limestone shown by structure contours in figure 97.

The gravel consists of sub- to well-rounded boulders set in matrix of sand or grit. Some of the boulders are as much as 3 feet across but most are about 4 inches. The boulders are of quartzite, pegmatite, amphibolite, and calc-silicate concretions, all readily identifiable as coming from nearby Precambrian rocks. Locally a few of the boulders are of Pahasapa limestone. The gravel also contains gold, cassiterite, and columbite-tantalite.

The only well-exposed deep opening in these deposits is in the gravel pit about 1,700 feet east of Griffis well in Griffis Canyon (pl. 21). A sand deposit that consists of quartz grains in a matrix of clay has no noticeable bedding although it is exposed for at least 8 feet in the face of this pit. This sand is overlain by flat-lying coarse gravel beds that form a vertical, crosscutting contact on one side of the sand deposit. The sand is white to tan, medium grained, subrounded, and exceptionally well sorted. Along joints and fissures it has been converted to clay. The deposit is lithologically similar to those in typical White River exposures described by Darton and Paige (1925, p. 15).

The deposits of volcanic ash are poorly exposed and the boundaries are very indefinite. Several shallow old prospect pits in the ash are located a few hundred yards southwest of Sevenmile school in Pleasant Valley. Some grayish-white partly altered ash in these pits has fragments of biotite, plagioclase, volcanic glass, and a few needles of amphibole(?). Biotite is easily visible in hand specimens and forms 2 to 8 percent of the total rock. The indices of refraction of the plagioclase are slightly above 1.55. The index of refraction of the glass ranges from 1.502 to 1.506±.003. W. E. Chisholm (oral communication, 1956) found that the heavy minerals are similar to those in other ash deposits from known White River localities in nearby Wyoming. The mineralogy of this ash fall is identical with that of an ash deposit about 6 miles south of the Fourmile quadrangle where Darton (Darton and Paige, 1925, p. 15) found White River fossils. On the other hand, the mineralogy of the ash does not closely resemble that of younger ash beds in Nebraska and adjacent areas described by Swineford, Frye, and Leonard (1948, 1955); thus, it seems likely that this ash is a part of the White River.

The extent and thickness of the ash is not known because it is so poorly exposed. Darton and Paige (1925, geologic map of central Black Hills region) showed White River deposits extending a mile eastward along Hay Creek from the junction with Pleasant Valley Creek. There is at the present time no conclusive evidence known for such an extension, but possibly early excavations which were later filled when the valley was farmed, contained exposures that cannot now be seen. Another exposure of ash just south of the entrance of Ninemile Draw is similar to the ash deposit near Sevenmile school.

These deposits containing volcanic ash are very close to the bottom of Pleasant Valley Creek, and indicate that in early Tertiary time this part of the valley had been eroded almost to the same level as today. The ash deposit 6 miles south of the quadrangle is about 150 feet above the adjacent Pleasant Valley floor.

QUATERNARY SEDIMENTS

Many of the stream valleys contain alluvium of presumably Recent age, and terraces on the sides of some of these valleys have deposits of sand and gravel that may have formed in the Pleistocene. Boulders in these deposits consist mainly of vein quartz, pegmatite, calc-silicate ellipsoids, and rarely schist. The calc-silicate ellipsoids are so resistant that they have been only slightly modified from their original shapes.

Gravel deposits are common along upper Lightning Creek, where they have been exposed to depths of 20 feet by placer mining. Some of these gravel beds are as much as 40 to 60 feet above the present valley. There is no precise evidence for their age, but they are believed to be Quaternary. A large terrace gravel deposit in the northwestern part of the quadrangle at the junction of secs. 2 and 3, T. 4 S., R. 3 E., with secs. 34 and 35, T. 3 S., R. 3 E., extends an even greater distance up on the side of the valley. Consequently, it may be of older age, possibly even of White River age. If the latter age is correct, the present drainage along Lightning Creek is largely inherited from early Tertiary time, as is true for Pleasant Valley.

Terrace gravel deposits along upper Hay Creek contain blocks and boulders of Paleozoic rocks as much as 4 feet in diameter. These are several miles from the present limit of Paleozoic rocks and must have formed when the Paleozoic rocks were much closer. They could well be residual gravel formed as early as the Oligocene, but it is likely that they have been somewhat modified in more recent times.

The alluvium beneath the valley floors is mostly covered by soil. Deposits consisting largely of gravel are exposed mainly along Lightning Creek and lower Pleasant Valley Creek. Most of these deposits are probably less than 20 feet thick.

STRUCTURAL GEOLOGY

A large southward-plunging open syncline is the major structural feature in the Precambrian metamorphic rocks. Most of the quadrangle is on the east limb of this feature, and bedding trends northwest and dips southwest. Paleozoic sediments, slightly warped and faulted, cover the western part of this syncline and dip gently southwest to west.

Axial plane and bedding plane schistosities, as well as a later northeast-trending schistosity, are common in the metamorphic rocks. The northeast-trending schistosity may be related to the intrusion of pegmatite and granite to the northeast.

Most pegmatites are parallel to the predominant foliations, but in some areas they are controlled mainly by joints.

PRECAMBRIAN ROCKS

FOLDS

The largest fold of the metamorphic rocks is a syncline whose axis passes close to the Tin Mountain mine (pl. 21). The axial plane of this fold strikes northward and dips almost vertically as indicated by the steep symmetrical dips on opposite limbs. Most of the western limb of the fold is concealed by Paleozoic sedimentary rocks, but bedding trends in the extreme northwestern part of the area suggest that the fold is relatively open. No single bed can be followed around the nose of the fold in the Fourmile quadrangle, and the location of the axis is based largely on a reversal of dips and on the north-northeast strikes west of the Tin Mountain mine. In the Berne quadrangle to the north, the Crow formation can be traced across the fold axis, but it is covered to the west by Paleozoic rocks (fig. 75). Graded bedding in metagrit at several localities near Fourmile and at the New York mine indicates that beds are upright and that the synclinal axis is southwest of these points. Minor folds and the attitudes of their limbs suggest that the plunge of the major fold is 40° to 50° S.

Minor folds are most abundant in the northeastern corner of the quadrangle, but only rarely can they be followed except by tracing the Crow formation. Some of these folds in the northern part of sec. 32, T. 3 S., R. 4 E. are open, gentle, and southward plunging; others in the extreme northeastern part of the quadrangle are small, tight, and isoclinal, and have a well-developed axial plane schistosity that strikes northwest and dips southwest. The open folds and the isoclinal folds have about the same plunge, and there is no field evidence that they were formed during different periods of deformation. Local thinning on the limbs of the folds suggest shear folding, but thicknesses of the Crow formation in the E½ sec. 33 and the W½ sec. 34, T. 3 S., R. 4 E., indicate considerable plastic flowage during folding. The intensity of folding apparently decreases to the west as indicated by the relatively straight contacts of the Crow formation. Farther west and north along the Crow formation in the Berne quadrangle, the dips of the axial planes of small isoclinal folds gradually steepen and ultimately become vertical near the main synclinal axis. The folds are therefore fan folds.

Minor folds are sparse in the central part of the quadrangle between the Crow formation and the synclinal axis. A single bed of metaconglomerate northwest of Fourmile can be traced for 1½ miles without any significant folding.

Along lower Lightning Creek many minor open folds are indicated by changes in strike and dip (pl. 21, section *A–A′*). Discontinuous outcrops of quartzose schist about 1,000 feet southwest of the Red Spar mine on Lightning Creek have a pattern that suggests isoclinal folding but the folds cannot be outlined with assurance. The observed minor folds near the major synclinal axis are open and have nearly vertical axial planes. To the north these folds become tighter and in part more nearly isoclinal. Thin amphibolite sills southeast of the Tin Mountain mine follow some of the open folds.

Very small chevron folds are fairly common, especially in incompetent rocks. These generally plunge southward parallel to the larger folds and presumably resulted from late shearing movements of the same deformation that produced the larger folds. Minor crenulations of bedding and foliation in the area along lower Lightning and Fourmile Creeks plunge more to the southwest than the other folds, and are at the intersection of a late northeast-trending foliation with northwest-trending bedding and foliation.

A northeast-trending cross flexure produces a change in the regional trends from northerly in the southeastern part of the quadrangle to northwesterly in the north-central part of the quadrangle. The continuation of this flexure produces a bend in the strike of the Precambrian rocks about 1 mile southeast of Custer (fig. 75). There are no known large folds related to this feature, but it may be associated with the late northeast-trending schistosity and have been caused by the intrusion of granite and pegmatite to the northeast.

FOLIATION

There are five types of foliation in metamorphic rocks in the Fourmile quadrangle. The oldest consists of rarely observed segregations of biotite into layers or pseudobeds that cross beds. The two main schistosities are younger; one is parallel to bedding planes and the other to axial planes of folds. Inasmuch as these cannot always be distinguished from each other in the field, they are shown by the same symbol on plate 21. The axial plane and bedding plane schistosities over most of the area strike north-northwest and dip southwest, as does most of the bedding. A less evident schistosity, younger than the two main types, strikes northeast and dips mostly southeast, shows no consistent relation to bedding, and crosscuts bedding everywhere except in the northeast-trending limbs of some small folds. The fifth type or youngest foliation occurs rarely and only around pegmatite bodies.

PSEUDOBEDDING

The oldest foliation, or pseudobedding, consists of thin layers of biotite-rich schist, 0.02 to 0.5 inch thick, that are discordant to true bedding (fig. 98) but similar in size and composition to many of the thin schist beds. A single biotite-rich layer may cross several beds, and small stringers of similar biotite-rich material join different layers together. The structure was recognized in only four outcrops, the best of which is shown in figure 98. The strike of the pseudobedding is more northerly or northeasterly than the strike of bedding in the same outcrops, and its dip is steeper. In two of the exposures part of the biotite is oriented parallel to the plane of the pseudobedding, but more commonly it is parallel to the main schistosity. Bedding is not deformed by the pseudobedding, and it seems likely that this foliation is in the plane of an early fracture cleavage or jointing along which biotite-rich material was segregated during metamorphism.

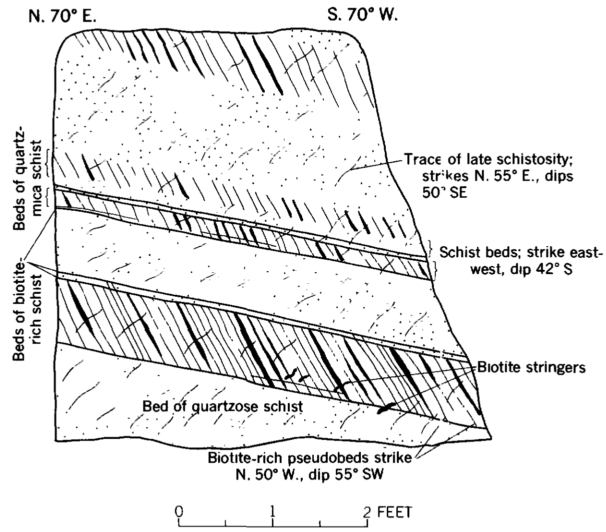

FIGURE 98.—Sketch of vertical face of outcrop showing pseudobedding in Mayo formation, SW¼NE¼ sec. 23, T. 4 S., R. 3 E. Biotite-rich layers or pseudobeds are at angle to true bedding and late schistosity.

AXIAL-PLANE AND BEDDING-PLANE SCHISTOSITIES

Schistosities parallel to axial and to bedding planes predominate over all other schistosities. They are caused mainly by the parallelism of micas but to a lesser extent by the orientation of elongated quartz and feldspar grains. In many outcrops axial and bedding plane schistosities are nearly parallel; in outcrops of thin micaceous beds interlayered with thin competent beds, bedding is by far the best slippage plane, and axial plane schistosity does not develop. In other outcrops where bedding cannot be recognized, only one schistosity is recognizable and one cannot tell which of the two varieties it is.

The schistosity, like the bedding, varies in dip from east to west and forms a fanlike pattern. Over most of the central and eastern parts of the area, the major schistosity strikes northwest and dips southwest, approximately parallel to bedding. In these same areas axial plane schistosity can be only locally recognized, especially on fold noses where it transects bedding at a high angle. Near the axis of the major syncline, axial plane schistosity strikes almost north and dips almost vertically. It is most conspicuous in the northern part of the area near the Tin Mountain mine and northward in the Berne quadrangle along the extension of the major synclinal axis. Where the synclinal axis follows Lightning Creek, axial plane schistosity is rarely found, and schistosity is mainly parallel to bedding in open folds. West of the synclinal axis, in the extreme northwestern part of the quadrangle, both bedding and schistosity strike northeast and dip southeast. The dip gradually becomes more gentle away from the synclinal axis. In only two places in this part of the area does the predominant schistosity have a dip steeper than that of the bedding.

Though the two main schistosities cannot be distinguished from each other in many outcrops, the slight angle between them can be recognized in petrofabric diagrams (fig. 99 *A*, *B*, S_1 and S_2). Rarely, as in figure 99 there are three maximums: one parallel to the bedding plane and two approximately parallel to the axial plane. The two axial plane maximums cannot be precisely interpreted. They may represent stages in the convergence of bedding and axial planes as folding became increasingly intense. A second possibility is that they represent the loci of slightly different shear planes produced by differences in competency and homogeneity of adjacent beds or larger stratigraphic units that were involved in the folding.

LATE SCHISTOSITY

A late northeast-trending and southeast-dipping schistosity is found locally throughout the quadrangle. A separate symbol is used on plate 21 to show its attitudes. This younger schistosity is the most prominent one in a few places in the southwestern part of the Precambrian area, especially along lower Lightning Creek. In these places the platy minerals are well oriented in the later schistosity, and the older bedding and axial plane schistosities are completely obliterated. The new schistosity crosses beds indiscriminately; it is parallel to bedding only in a few places on limbs of folds. In most places the late schistosity is poorly developed or not visible at all and the dominant schistosity remains parallel or subparallel to bedding. The dip of the late schistosity is variable, but is generally quite steep to the southeast.

The only large fold parallel to this schistosity is the major northeast-trending broad regional flexure that passes southeast of Custer (fig. 75). The northeast-trending schistosity is most conspicuous near the axis of this flexure, and the two features may be related in origin. Probably the schistosity was formed by northeast-trending shear movements. It may be related to the intrusion of granite and pegmatite near Harney Peak (fig. 75), but for the present this relation is conjecture. The schistosity preceded and controlled the emplacement of some of the northeast-trending pegmatites.

INDUCED FOLIATION AROUND PEGMATITES

The youngest foliation was induced in the wallrock along the contact of some of the pegmatite bodies. It consists of closely spaced fractures with some parallelism of micaceous minerals along the fractures. It is

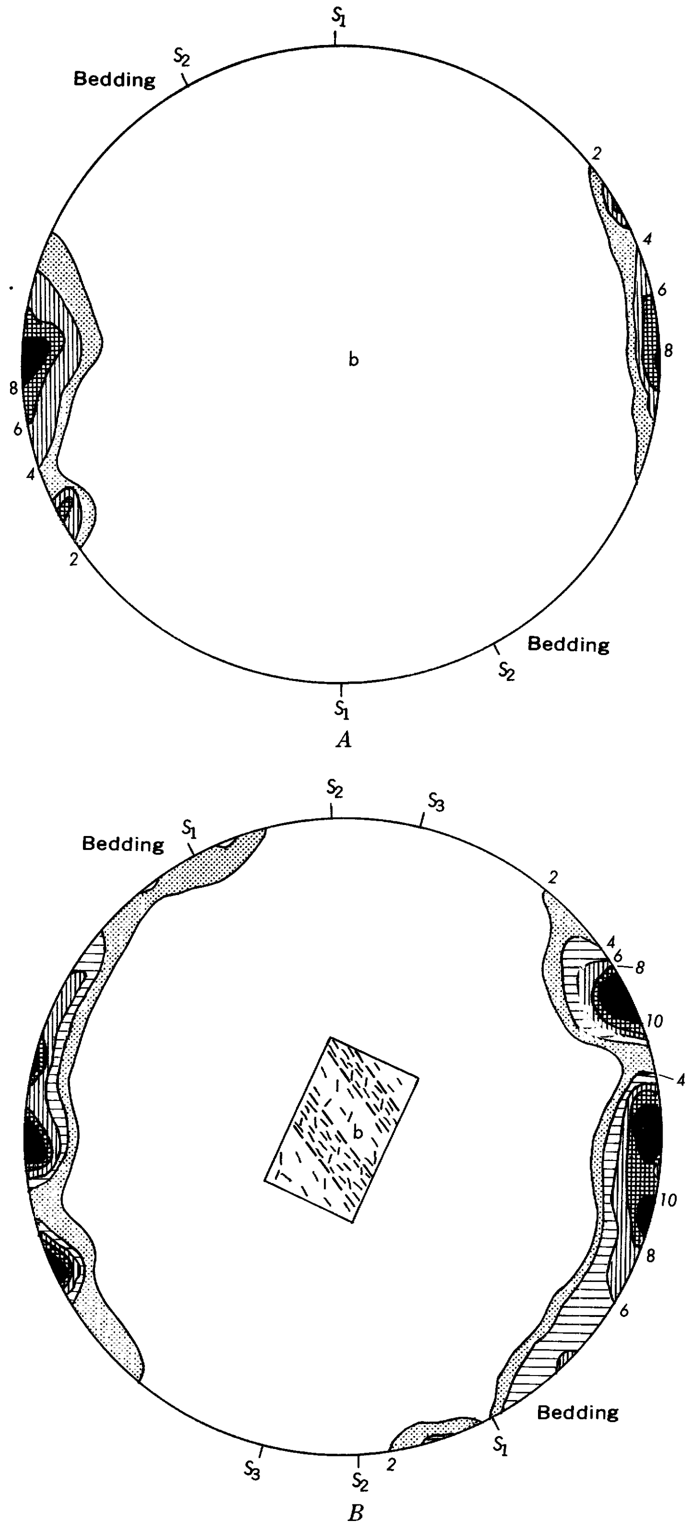

FIGURE 99.—Diagram showing poles of mica in schist of Bugtown formation (SW¼ sec. 19, T. 3 S., R. 4 E). *A*, Poles of 200 mica cleavages. Contours at 2 percent intervals. S_1 parallel to axial plane of folds, S_2 parallel to bedding. *B*, Poles of 200 mica cleavages. Contours at 2 percent intervals. S_1 parallel to bedding. S_2 and S_3 approximately parallel to axial plane of folds.

only observable where the pegmatite contact is discordant as in plate 22 (loc. C). Along the other contacts of pegmatites, the pre-existing schistosity of the country rock has been deflected into parallelism with the contact, but there is little evidence that it was reformed or that recrystallization occurred during this deflection.

LINEATION

Linear elements in the Precambrian rocks in the quadrangle include folds, mineral parallelism (in part due to intersection of foliation surfaces), elongate pebbles, calc-silicate ellipsoids, and elongate mineral aggregates (sillimanite-quartz knots). The small fold axes, and the lineation produced by intersection of bedding plane and axial plane schistosities are parallel and plunge 30° to 52°, S. 20° E. to S. 15° W. This is probably the primary lineation throughout the area. Elongate pebbles, calc-silicate ellipsoids, and knots of sillimanite and quartz generally plunge parallel to this primary lineation.

A secondary lineation is locally formed by the intersection of the late northeast schistosity and the generally northwest-trending bedding or axial plane schistosity. This lineation generally plunges 25° to 65° S. to S. 38° W. Locally where the late schistosity is well formed, small crinkle folds and elongated sillimanite-quartz knots plunge parallel to the late lineation. In areas of relatively little folding, such as northeast of the New York mine, the secondary lineation is dominant over the primary lineation.

The apparent elongation of the calc-silicate ellipsoids is considerably greater than the elongation of the pebbles. The ratio of the longest to the shortest axis of the ellipsoids is about 15:1; it is only 3:1 for pebbles consisting of fine-grained minerals and 2:1 for pebbles of coarser grained minerals. Even greater elongation of the ellipsoids occurs in the Berne quadrangle to the north.

If both the calc-silicate ellipsoids and the pebbles were almost spherical prior to deformation, their present elongation parallel to fold axes *b* rather than to the direction of greatest tectonic transport (generally inferred to be normal to fold axes and in the *a* direction) requires explanation. Fairbairn (1936, p. 678) pointed out that if such structures are rotated around *b*, any single part of the body will be alternately stretched along *a* and then shortened along *c*, but it will be continuously lengthened along *b*. Though this may apply in part of the Fourmile quadrangle, the very great elongation of the calc-silicate ellipsoids in comparison with the pebbles suggests elongation by plastic deformation that did not affect the more competent parts of the rock. It is also possible that the ellipsoids never were equidimensional, but that instead they were formed during deformation. Their shape is similar to the shapes of the sillimanite-quartz knots. Furthermore, the minerals of the ellipsoids are not deformed and presumably crystallized after folding. On the other hand, the intermediate axes of a few of the ellipsoids are folded, especially along a small ridge a few hundred feet north of the Tin Mountain mine, and thus at least part of the deformation was after the formation of the ellipsoids.

FAULTS, JOINTS, AND BRECCIA ZONES

A few faults of Precambrian as well as of later age disrupt the metamorphic rocks. The two largest faults are post-Paleozoic and probably are of early Tertiary age. One is subparallel to Pleasant Valley and the other, in the western part of the area, is parallel to U.S. Highway 16. Both displace the Paleozoic formations. These faults are largely covered by alluvium and cannot be followed accurately across the area of Precambrian rocks. The fault along Pleasant Valley continues northeast along Fourmile Creek but does not noticeably displace the Crow formation. In the sedimentary rocks near Sevenmile school, however, the displacement is as much as 320 feet. The late northeast schistosity commonly is well formed near this fault and probably controlled its attitude.

A similar fault about 400 feet east of the main fault in Pleasant Valley, and parallel to it, has displaced metagrit beds just west of the New York mine. The apparent vertical displacement near the New York mine is 100 to 175 feet. Inasmuch as there is no noticeable displacement of the pegmatites along this fault, the fault is older than these pegmatites and is of Precambrian age.

Small faults of a few feet displacement were recognized along Hay Creek but are not shown on the map (pl. 21). These faults trend northeast parallel to the northeast-trending schistosity and to small faults in the Paleozoic rocks. One of these faults displaces the New York pegmatite. These features indicate a post-Paleozoic age, and suggest a possible association with the Tertiary doming of the Black Hills

The contacts of a few scattered pegmatites with the country rock are offset a few inches or a few feet by small faults a few tens of feet long that strike nearly parallel and dip perpendicular to the contacts. It is significant that such faults do not cross the pegmatites, but do offset the opposite walls. The movement was such that the hanging wall of each fault has moved outward with respect to the footwall so that the pegmatite is thicker above the fault. The faults commonly contain fracture fillings. Several are found in pegmatite 14 (pl. 22, section A–A') and pegmatite 124. These small faults probably were formed by forceful intrusion of pegmatitic material; they may be miniature equivalents of marginal thrusts such as may accompany the intrusion of large granitic masses (Billings, 1947, p. 305–306).

Three different joint sets are prominent in the area. One set strikes about northwest parallel to bedding and dips steeply northeast; the second set strikes north-northeast and dips vertically or steeply southeast; the third strikes northeast to east and dips vertically. Many more pegmatites are emplaced approximately parallel to the second and third sets than to the first set, but on the other hand joints of the second and third sets cut pegmatites whereas joints of the first set do not. The frequency distribution of the second and third sets corresponds with the distribution of poles to discordant pegmatites shown in figure 82. The second set is especially well formed west of Pleasant Valley, where it is characterized by surfaces coated with fine-grained quartz or hematite. The first and third sets are fairly abundant throughout the area, but the third is most common in the southeastern part of the quadrangle.

Some pegmatites in the general area west of Fourmile Creek follow north-northeast joints and locally bedding, thereby creating the "right-hand" jogs previously mentioned (fig. 80G). The lack of "left-hand" jogs suggests that the pegmatites were emplaced along joints that were opened by a shear couple resulting from a southward and possibly upward movement of rocks to the east relative to rocks to the west. These joints therefore may be related to the emplacement of the large mass of granite and pegmatite northeast of the quadrangle.

Several pegmatite bodies contain quartz-rich fracture fillings which are apparently along joint planes that are about normal to the plunge of the enclosing pegmatite body. The schist country rock has flowed somewhat into the breaks in the pegmatites to produce a structure similar to boudinage (figs. 100 and 101). In places the partings between boudins of undisturbed pegmatite are as much as 30° off normal to the plunge of a pegmatite body, but some of these are parallel to a strong joint set in the country rock. The schist that enters the fracture between boudins is distorted for a distance of only 2 to 3 feet away from the pegmatite contact, and therefore deformation must have been largely plastic. Partings between boudins, appearing at intervals of 30 to 40 feet in many pegmatites, indicate tensional stress normal to the plunge at a time near the end of the period of pegmatite crystallization, when the rock was sufficiently competent to break and when confining pressure was still adequate for plastic deformation of the country rock.

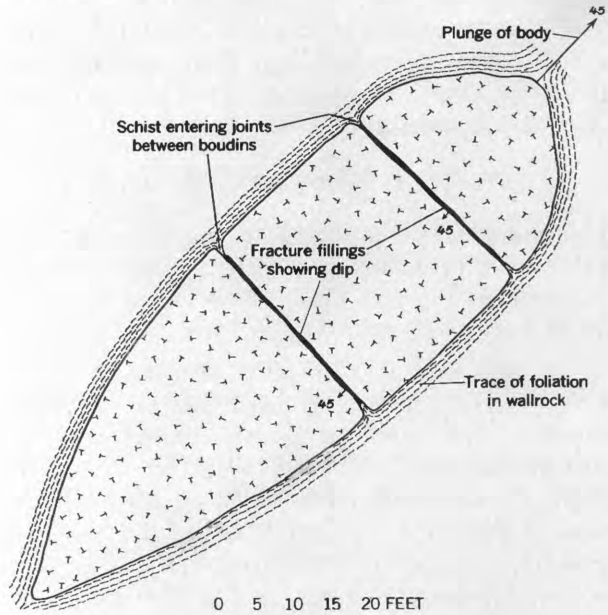

FIGURE 100.—Diagram showing boudinage development and orientation of filled cross joints in vertically dipping pegmatite body.

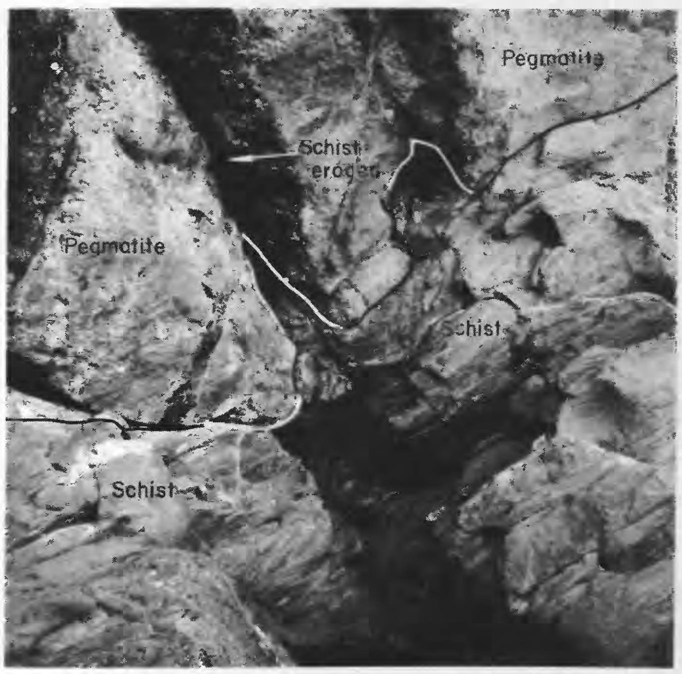

FIGURE 101.—Boudinage structure in pegmatite. Pegmatite contact is parallel to foliation of country rock. Schist penetrates pegmatite approximately 2 feet. Pegmatite in SE¼ sec. 10, T. 4 S., R. 4 E. Photograph by Philip Orville.

Thin breccia zones are exposed about a mile northeast of Fourmile as shown on plate 21. They are concordant with the schist, are only a few feet thick, and consist of angular to rounded fragments of schist cemented by hematite or other minerals. Some of the cement in microbrecciated parts consists entirely of albite. Though the fragments are concordant, and their movement was parallel to bedding, their schistose nature indicates an emplacement origin after the main period of folding and metamorphism.

PALEOZOIC AND YOUNGER ROCKS

The structure of the Paleozoic and younger rocks in the Fourmile quadrangle is closely related to the uplift and erosion of the Black Hills. These rocks have a westward and southwestward regional dip of 1½° to 2°.

Other small structural features are discernible from the contours on the base of the Pahasapa limestone in figure 97. The most conspicuous feature is a small southwest-plunging anticline parallel to Pleasant Valley. On the south flank of this anticline in NW¼ sec. 35, T. 4 S., R. 3 E., a small dome has about 20 feet of closure, probably as a result of differential compaction over topographic highs on the Precambrian surface. Another small structural high is just west of the quadrangle along Layton Canyon. The upper 8 feet of the Deadwood in this area is extremely well rounded sandstone suggesting winnowing on a slight topographic high that existed during or shortly after the Cambrian. If the block of basal Deadwood formation is actually in place in the northeastern part of the quadrangle, its presence suggests a southwest-plunging syncline along upper Fourmile Creek.

Small-scale structural features indicated by contours on the top of the Pahasapa limestone are not shown by contours on the base of that formation (fig. 97). As stated previously, these differences probably were caused both by selective solution of the upper part of the formation and by erosion at the unconformity between the Pahasapa and the overlying Minnelusa sandstone. The largest apparent solution feature or flexure is the trough parallel to Pleasant Valley. Other somewhat similar flexures appear in figure 97. The boundaries at most of these flexures are along straight lines and resemble small monoclines. Two of the north-trending monocline-like flexures become faults when traced southward. One of these flexures in the upper contact dips 40° east on the north side of S & G Canyon, but about 300 feet to the south along the flexure axis the contact is clearly a fault. The pattern and trend of the flexures suggest that solution of the Pahasapa was controlled by joints. Pleasant Valley follows the largest of these solution features, and presumably is controlled in part by it. The distribution of White River deposits suggests that there may also have been a valley in the same place in the Oligocene, and thus the solution effects may be largely pre-Oligocene.

The varied strikes and locally moderate dips of beds in the Minnelusa sandstone do not outline a consistent fold pattern and are believed to be caused by solution and collapse folding of underlying rocks.

The east-west fault parallel to U.S. Highway 16 consists of 2 branches. The vertical displacement across these 2 branches is at least 190 feet at the west edge of the quadrangle, and probably is about 300 feet, as suggested by the displacement of the contours on the base of the Pahasapa limestone in figure 97.

The large northeast-trending fault parallel to Pleasant Valley displaces the Pahasapa limestone approximately 320 feet near Sevenmile school. The displacement decreases to the southwest (fig. 97).

Origin during the doming of the Black Hills is suggested by the fact that all these faults have been upthrown on the north side. Lack of displacement of White River deposits by the fault near Sevenmile school indicates a pre-Oligocene age for that fault. Presumably the other faults are of the same age.

METAMORPHISM

REGIONAL METAMORPHISM

Most of the metamorphic rocks of the Fourmile quadrangle are in the sillimanite zone; only those in the extreme northwestern corner of the quadrangle are in in the staurolite zone. The highest metamorphic-index mineral in most of the quartz-rich schist is garnet, and only the more aluminous beds contain sillimanite and staurolite. The sillimanite isograd trends northeast across the Fourmile quadrangle and the southern part of the Berne quadrangle. Farther north it swings northward and then continues more or less parallel to the contact between the granitic rocks and the schists (fig. 78).

Relict structure and textures, especially bedding, are well preserved in many places. Grains and pebbles of recrystallized quartz and feldspar can be distinguished in quartz-feldspar-mica gneiss formed from grit and conglomerate. Mafic intrusive rocks now consisting of amphibolite contain no recognizable relict minerals, and relict structures have been found only in areas outside of the Fourmile quadrangle. Hornblende, oligoclase-andesine, biotite, garnet, and quartz were formed in the amphibolites during metamorphism.

The optical properties of the biotite, feldspar, and other minerals of the schists do not change appreciably with increasing metamorphism from the staurolite to the sillimanite zone.

STAUROLITE ZONE

Rocks in the staurolite zone are exposed only in the extreme northwestern corner of the quadrangle, where schists of the Mayo formation characteristically have staurolite, garnet, and biotite. The same rocks contain sillimanite along strike to the southeast. The rocks of the lower part of the Mayo formation, of the Bugtown formation, and of the Crow formation are not exposed in the staurolite zone in the Fourmile quadrangle, but they are in the Berne quadrangle to the north.

The schists in the Mayo in the staurolite zone consist mainly of quartz, biotite, muscovite, and plagioclase. Microcline is rare and generally restricted to metagrit and metaconglomerate. The rocks are medium grained, and although the micas can be easily identified in hand specimens, quartz and feldspar cannot be distinguished from each other. Generally, in the more micaceous beds (especially the biotite-rich ones), medium- to coarse-grained garnet and coarse-grained staurolite form conspicuous porphyroblasts in the schist. The outer parts of some of the staurolite crystals are replaced by muscovite, and rarely there are complete pseudomorphs of muscovite after staurolite. The most common mineral assemblages for these schists are listed in table 11.

Kyanite was also found in a single bed in the lower part of the Mayo formation in the Berne quadrangle. In the Fourmile quadrangle it is associated with sillimanite in a quartz vein in the NW1/4 sec. 6, T. 4 S., R. 4 E., near the sillimanite isograd. The general absence of kyanite is believed to be caused largely by the premetamorphic composition of the sedimentary rocks. The rocks of the Mayo formation ordinarily contain only enough aluminum to form the micas and feldspars as indicated by table 3. The small amount of sillimanite in the norms of specimens JR-1-54 and JR-2-54 is not likely to be found in thin sections, and only slight changes in the mineral compositions or errors in the analyses would eliminate it from the norms. The more aluminous beds are also richer in iron, and therefore contain garnet and staurolite instead of kyanite.

In the staurolite zone in the Berne quadrangle the Crow formation is characterized by such minerals as cordierite, hornblende, tremolite, diopside, calcite, and scapolite which indicate quite a different original composition from that of the Mayo formation. The minerals of the Crow formation do not change greatly from the staurolite zone to the sillimanite zone. In the amphibole-rich rocks in the Berne quadrangle, the hornblende gradually becomes lighter green toward the northwest, farther away from the sillimanite isograd. Tremolite also becomes abundant 2 to 3 miles from the sillimanite isograd; it apparently takes the place of diopside in magnesium-rich rocks, or it represents the equivalent of hornblende in the amphibole schist of the sillimanite zone. In the quartz-mica-feldspar schist, kyanite occurs instead of sillimanite.

TABLE 11.—*Metamorphic mineral assemblages of the Fourmile quadrangle*

[Minerals arranged in order of approximate decreasing abundance. The relative abundance varies locally]

Assemblage	Staurolite zone	Sillimanite zone
	Schists from Mayo and Bugtown formations	
1	Quartz, muscovite, biotite ± plagioclase ± microcline.	Same as staurolite zone ± sillimanite.
2	Quartz, biotite, plagioclase, and garnet.	Do.
3	Quartz, biotite, muscovite, and garnet ± feldspar.	Quartz, biotite, garnet, and feldspar ± sillimanite.
4	Quartz, biotite, staurolite, and garnet ± muscovite ± plagioclase.	Quartz, biotite, staurolite, and garnet ± sillimanite ± plagioclase.
	Calc-silicate gneiss from Mayo and Bugtown formations	
5	Quartz, calcite, plagioclase, hornblende, garnet, biotite and microcline ± clinozoisite ± diopside.	Same as staurolite zone.
	Amphibolite	
6	Hornblende and plagioclase ± garnet ± quartz ± biotite.	Same as staurolite zone.
	Schists and gneisses from the Crow formation	
7	Not exposed	Biotite, cordierite, microcline, plagioclase, and sphene.
8	do	Microcline, biotite, and muscovite.
9	do	Biotite, hornblende, diopside, microcline, plagioclase, calcite, and quartz ± scapolite.
10	do	Biotite, hornblende, quartz, and plagiclase.
11	do	Hornblende, plagioclase, calcite, and quartz.
12	do	Calcite, diopside, plagioclase, hornblende, and clinozoisite.
13	do	Quartz, muscovite, biotite, plagioclase, and sillimanite.

The Bugtown formation in the staurolite zone has quartz, plagioclase, microcline, muscovite, biotite, and rarely garnet. It contains insufficient aluminum or iron to form staurolite, except in the very lowest part of the formation, which does not cross into the Fourmile quadrangle.

The calc-silicate ellipsoids are much coarser grained than the associated schists. They contain the minerals listed as assemblage 5 (table 11) in both the staurolite and sillimanite zones. Garnet, hornblende, and diopside are coarse grained and the other minerals are medium grained. Although all the minerals in assemblage 5 may occur in a single thin section, hornblende and biotite are most abundant in, or limited to, the outer parts of ellipsoids, where apparently water was available for their formation. The centers of the ellipsoids on the other hand, contain the water-poor mineral clinozoisite, associated with the "dry" minerals, plagioclase, microcline, diopside, garnet, and calcite. The latter minerals are equant, and the biotite and hornblende in the outerparts have characteristic tabular habit. The coarse-grained minerals are extremely poikilitic, are not granulated, and evidently grew after the ellipsoids had attained their present shape.

Runner and Hamilton (1934, p. 60–64), as a result of a study of calcareous ellipsoids, postulated more than one metamorphism in the Black Hills. They stated that the biotite and hornblende were formed by recrystallization in the interiors of the ellipsoids of the earlier formed metamorphic minerals. It seems much more likely that the biotite and hornblende represent either an original difference in composition in zoned concretions or a difference produced by migration of material into or out of calcareous concretions during metamorphism. The layer of hornblende and biotite could well represent the limit of migration of water into the concretions.

Amphibolite in the staurolite zone is generally medium to coarse grained, uniformly textured, and devoid of relict structures or textures. It consists mostly of hornblende and plagioclase (oligoclase-andesine). The plagioclase has strong reverse zoning, twinning is rare, and it seems evident that all the plagioclase of the original igneous rock has been recrystallized. The presence of biotite, garnet, and quartz in the outer parts of some bodies suggests metasomatic introduction of material from the country rock during metamorphism. The biotite in the outer parts of amphibolite bodies indicates addition of potassium; furthermore, the abundance of garnet in adjacent schists indicates that iron left the amphibolite, either during the original intrusion or during subsequent regional metamorphism. Abundant actinolite and plagioclase in a schist inclusion of the Mayo in amphibolite may reflect contact metamorphic effects that were later obscured by the regional metamorphism.

SILLIMANITE ZONE

The sillimanite isograd crosses the northwestern part of the quadrangle (pl. 21), and all rocks to the southeast are the sillimanite zone. At the isograd there are no changes in the appearance of most of the rocks, and only a few beds in scattered outcrops contain sillimanite. To the southeast along strike, however, sillimanite gradually becomes more abundant, until in the extreme southeastern part of the quadrangle, it occurs in nearly every outcrop of the Mayo formation. An accompanying increase in grain size is most evident in biotite, but knotlike aggregates of sillimanite and quartz increase from a millimeter to a centimeter or more in diameter.

Unaltered staurolite persists into the sillimanite zone at least as far southeast as the MacArthur mine (pl. 21, pegmatite 137, G–4). On the other hand, staurolite pseudomorphs consisting of retrograde muscovite and biotite can be found throughout the length of the quadrangle. In the northwest these pseudomorphs commonly have fresh cores of staurolite surrounded by a thick layer of muscovite, which in turn is rimmed by biotite. Where pseudomorphism is complete, especially in pegmatite-rich areas in the southeast, the biotite is absent, and presumably the iron of the original staurolite is dispersed in the surrounding rock.

No evidence was found that any staurolite has been converted to sillimanite. Muscovite, however, decreases in abundance as sillimanite increases, and much of the muscovite that does appear in the higher grade rocks is in porphyroblasts that cross schistosity and presumably formed as a retrograde product during decreasing metamorphism. Thus it may be supposed that much of the sillimanite formed from muscovite. Near the sillimanite isograd, however, muscovite does not decrease noticeably in abundance, and the small amount of sillimanite in rocks near the isograd may have formed from fine-grained, micalike minerals that appear in lower grade rocks. On the other hand, in the very high grade rocks in the southeast part of the quadrangle, aggregates of sillimanite invade biotite in a fashion that leaves little doubt that biotite was the parent mineral, and there is no clear evidence that muscovite ever existed during increasing metamorphism (although it probably did).

The conversion of mica to sillimanite releases potassium, and according to Turner (1948, p. 85) produces potassic feldspar. However, the few microcline-rich beds in the Mayo formation show no tendency to be associated with sillimanite schist. Furthermore, the pseudomorphs of mica after staurolite are much too sparsely distributed to contain the necessary potassium. The only alternative possibility is that potassium migrated away from the rocks during metamorphism, but more complete sampling and chemical analyses are needed to substantiate this hypothesis.

The Crow formation, which in the Fourmile quadrangle is entirely in the sillimanite zone, contains a variety of mineral assemblages listed in table 11. Most of the rocks do not differ in appearance from those in the staurolite zone, inasmuch as the major difference is the change from tremolitic amphibole in the staurolite zone to either diopside or actinolitic amphibole in the sillimanite zone. If the tremolite changes to actinolite, metasomatic changes in the Fe-Mg content are indicated. The few thin aluminous schist beds in the formation acquire abundant sillimanite in the extreme northeast part of the quadrangle. In the Berne quadrangle, sillimanite extends farther to the northwest in some of the schist beds of the Crow formation than it does in the Mayo and Bugtown formations, in spite of the fact that both of these formations have highly aluminous beds that seem just as favorable for the formation of sillimanite as the sillimanite-bearing parts of the Crow formation. Perhaps relatively abundant CO_2 in the Crow formation was responsible for accelerating the metamorphic reactions.

The absence of garnet in the metamorphic mineral assemblages of the Crow formation is unexpected and difficult to explain. Calc-silicate rocks in the Mayo formation and cordierite-bearing rocks in the aluminous schist of the Berne quadrangle contain garnet as expected. Yoder (1952, p. 623) stated that garnet should form in a water-deficient environment and that cordierite is stable in the presence of excess water above approximately 400° C. Thus, an excess of water in the rocks of the Crow formation may be the reason for the presence of cordierite and the absence of garnet. If the rocks of the Crow formation were originally calcareous as seems likely, they may have been relatively impermeable compared to the overlying and underlying rocks. Therefore, garnet may have grown in the quartz-mica schist where greater permeability permitted water to migrate outward; this migration would have left the rocks deficient in water. Another possibility is that the physical conditions of pressure and temperature were not favorable for a magnesium-rich garnet, but were favorable for the iron-rich garnet.

The presence of cordierite in a potassium-rich environment is also somewhat unexpected. Yoder (1952, p. 623) pointed out that cordierite is relatively rare in regionally metamorphosed rocks and suggested that this sparcity results from an excess of potassium that causes the growth of micas instead of cordierite. Thin sections of the Crow formation contain cordierite grains surrounded by rims of biotite and embayed by microcline. A few grains are so completely replaced by microcline that only the biotite rims and a few relicts of cordierite remain as evidence that cordierite was ever present. Thus it appears that some of the potassium may have been introduced after cordierite formed. The potassium could have come from adjacent potassium-rich beds.

The amphibolite and the calc-silicate rock in the sillimanite zone are identical in appearance and mineralogy to those in the staurolite zone.

Apparently no significant differences were caused by metamorphism in the composition of the ubiquitous plagioclase or biotite in the sillimanite and the staurolite zones. Plagioclase ranges from An_{20} in some schist of the Mayo to An_{90} in some of the calc-silicate rocks, but this range undoutedly reflects differences in the original composition of the sediments. The range for most of the schist is approximately from An_{25} to An_{33}. The determination of the β index of biotite on about 50 samples of schist from throughout the area indicates no systematic variations.

CONTACT METAMORPHISM

The effects of contact metamorphism are generally limited to minor recrystallization and metasomatism of wallrock for a distance of a few feet outward from pegmatite contacts. The wallrocks adjacent to discordant pegmatities are more highly altered than those bordering concordant ones, and hanging-wall contacts commonly have a greater amount of alteration than footwall contacts. Tourmaline, apatite, feldspar, and mica are the predominant new minerals, and locally they are very abundant. Some contact-metamorphosed rock contains as much as 60 percent fine-grained tourmaline. At the Warren Draw pegmatite, discordant inliers of schist on top of the pegmatite contain abundant coarse-grained tourmaline, apatite, and muscovite as far as 4 feet away from the contact, and locally the county rock adjacent to the contact is almost completely replaced by tourmaline. An analyzed inclusion of schist at Calamity Peak contains an extraordinary amount of boron as shown by table 8.

Large microcline porphyroblasts as much as 1 inch in diameter locally constitute as much as 60 percent of the schist close to the contacts of several pegmatites near the White Spar pegmatite (pl. 21, G–1). The porphyroblasts are generally above the hanging walls of the pegmatites and mostly in relatively coarse-grained beds that probably are more permeable than adjacent beds. In several places the porphyroblasts are concentrated along the upper side of moderately well defined healed joints that dip at low angles toward the hanging wall of a pegmatite. Higazy (1949) stated that, in the Black Hills, schist with porphyroblasts grades into pegmatite formed by metesomatic processes, but no such features have been reported by anyone else. Microcline porphyroblasts in the schist contain as much as 25 percent inclusions, quite unlike the microcline in the pegmatite. Although the centers of some of the porphyroblasts contain fewer inclusions than the outer parts, no gradation to microcline or perthite like that in the pegmatite was seen during the present study.

The chemical analyses shown in table 12 substantiate the obvious conclusion that potassium was added to the schist. These analyses were made of samples collected about half a mile east of the White Spar mine where porphyroblasts are common in schist adjacent to pegmatite. A single bed of quartz-mica-feldspar schist was sampled at 2 places 30 feet apart: (a) above the hanging wall of a pegmatite where it contained microcline porphyroblasts and (b) 10 feet beyond the end of the pegmatite body where no porphyroblasts exist. The analyses showed these 2 samples (table 12) are very similar except that the sample from above the pegmatite contains about 2 percent more K_2O and 2 percent less iron oxide than does the sample from beyond the pegmatite. Because the two samples are from the same bed and only 30 feet apart, there is little doubt that potassium has been added and iron removed.

TABLE 12.—*Chemical analyses of samples from a quartz-mica-feldspar schist bed (Bugtown formation) showing effects of potassium metasomatism near pegmatite*

[Rapid rock analyses by: Harry Phillips, J. M. Dowd, Katrine White]

	Sample 1; 10 ft from end of pegmatite (lab. no. 53-44CW)	Sample 2; 30 ft from sample 1 and 2 ft directly above hanging wall of pegmatite (lab. no. 53-45CW)	Difference in terms of sample 2 (percent)
SiO_2	72.2	71.5	−0.7
Al_2O_3	14.3	15.7	+1.4
Total Fe as Fe_2O_3	4.6	2.6	−2.0
Fe_2O_3	1.0	1.1	+.1
FeO	3.3	1.4	−1.9
MgO	1.3	.66	−.64
CaO	.90	.58	−.32
Na_2O	2.9	2.8	−.1
K_2O	3.1	5.2	+2.1
TiO_2	.53	.25	−.28
P_2O_5	.14	.22	+.07
MnO	.07	.07	0
Ignition [1]	1.1	.90	−.2
Total (rounded)	101	100	

[1] Includes gain due to oxidation of FeO.

The apparent localization of the porphyroblasts along fractures and permeable layers suggests that transfer was by fluid rather than by solid diffusion. Similarly, the amount of water, boron, and fluorine in such minerals as muscovite, tourmaline, and apatite in the wall rock adjacent to other pegmatites suggests fluid transfer.

Amphibolite is generally converted into a biotite-quartz-epidote-garnet-chlorite rock for a few inches—rarely several feet—from contacts with pegmatite. Biotite is the most abundant mineral, and presumably was derived from decomposition of hornblende and the introduction of K_2O from the pegmatite. The border zones of the pegmatites commonly are rich in biotite and apatite, and they contain more iron and calcium than the border zones in pegmatites intruded into schist. There has been some reciprocal transfer of chemical constitutents across the contacts, but both the pegmatite and the amphibolite retain their identity.

The alteration of staurolite and sillimanite to aggregates consisting mainly of muscovite, biotite, and chlorite is mainly in pegmatite-rich areas. Though there is at present no clear proof of any changes in bulk chemical composition during this retrograde activity, it is likely that the potassium and water necessary for the development of micas from staurolite and sillimanite were derived from the pegmatitic magma.

TECTONIC HISTORY

The main period of Precambrian folding in the area was probably caused by compressive forces acting on deeply buried sedimentary rocks consisting largely of graywacke and subgraywacke. These forces possibly were directed along a line trending approximately N. 80° E. In response to the compressive forces, a large relatively open south-plunging syncline was formed in the western part of the quadrangle, and an anticline formed northeast of the quadrangle. Small open upright folds having an axial plane cleavage formed near the synclinal axis, and inclined drag folds formed on the limb of the large anticline to the east. The axial planes of these folds formed a fanlike pattern. Metamorphism presumably accompanied the folding, and with increasing metamorphism the sedimentary rocks were converted to medium- to high-grade schists and gneisses. Shearing between adjacent beds led to the development of bedding plane schistosity during the folding and metamorphism. Furthermore, as the folds became tighter, axial plane cleavage became more nearly parallel to bedding. Small bodies having the composition of gabbro or diabase were intruded late in the time of folding, some parallel to folded beds, others parallel to axial plane foliation, and all were metamorphosed to amphibolite. Quartz veins were formed somewhat later than most of the shearing, yet metamorphic intensity was sufficiently high to produce in the veins the typical metamorphic minerals. Some massive quartz veins containing gold may have been formed during a period of Precambrian hydrothermal mineralization.

After the completion of the major folding and possibly as early as the intrusion of the gabbro, but while temperatures were still high, a northeast-trending flexure and late northeast-trending schistosity formed. The only evidence that the schistosity formed as early as or prior to the gabbro is in the northeast trend of two bodies of amphibolite about half a mile west of the New York mine in the west-central part of sec. 18, T. 4 S., R. 4 E. The cause of the forces that formed the northeast-trending structural features is unknown. Possibly the deforming forces were related to the emplacement of granite and pegmatite in the vicinity of Harney Peak. This northeast-trending schistosity formed only locally and probably began as a slip cleavage. In places, it obliterated the earlier northwest-trending schistosity.

Granite and pegmatite were emplaced after the main folding, but the higher grade metamorphism and some of the folding may have been an advance manifestation of the period of intrusive activity. Emplacement probably began northeast or east of the quadrangle and spread gradually southwest and west across the Fourmile quadrangle. The emplacement of the large mass of granite and pegmatite around Harney Peak produced a general doming of the surrounding rocks, and probably caused the relatively flat dips in the country rock in the northeast part of the Fourmile quadrangle. Northeast-trending joints were formed and many pegmatites were emplaced along such joints as well as along the northeast-trending foliation. Some pegmatites west of Pleasant Valley followed north-trending joints that may have been caused by upward and southward movement of the eastern part of the area.

The granite and pegmatite were emplaced approximately 1.6 billion years ago, following the major orogeny. There is no record of later activity in the Precambrian rocks other than that of uplift and subsequent deep erosion to a moderately rolling surface before late Cambrian time. Deep semitropical or tropical weathering occurred before the deposition of the Upper Cambrian Deadwood formation. In the southern Black Hills, there is a hiatus between the Deadwood formation and the overlying Englewood limestone of Mississippian age. Some uplift at the close of the Mississippian resulted in erosion and leaching of the Pahasapa limestone. Further deposition took place at various times from the Pennsylvanian to the Tertiary period.

In early Tertiary time the Black Hills area was uplifted and eroded, and the Precambrian core was exposed. In much of the Fourmile quadrangle, the Paleozoic rocks were stripped back almost to where they are today. Solution occurred in the upper part of the Pahasapa limestone along joints, and valleys were probably controlled by these solution channels. In Oligocene time the White River group was deposited along these old valleys and on the plains outside the Black Hills. Moderate erosion has removed most of the Oligocene deposits, and the upper parts of others have probably been reworked.

ORIGIN OF THE PEGMATITE AND THE PEGMATITE BODIES

The major hypotheses for the origin of pegmatite are magmatic crystallization, metasomatic replacement, lateral secretion, and combinations of two or more of these processes (Kemp, 1924; Johannsen, 1932; Jahns, 1955). In the 1920's several writers advocated a magmatic origin for the simple quartz-feldspar pegmatites and a hydrothermal replacement origin for some of the minerals in complex pegmatite bodies (Schaller, 1925; Landes, 1925; Hess, 1925). More recently, extensive studies of the internal structures of zoned pegmatites suggest that the pegmatites are of magmatic origin and that they crystallized in a nearly closed system, rather than in an open system in which hydrothermal solutions played a dominant role (Cameron and others, 1949; Page and others, 1953; and Jahns, 1955). Other authors believe that metasomatism plays a major role, if not the only role, in the formation of pegmatites (Ramberg, 1952, p, 238–270; Barth, 1952, p. 22–24).

The field evidence in the Fourmile quadrangle suggests that the metamorphic rocks were intruded by many separate bodies of fluid that subsequently crystallized to form pegmatite. Furthermore, the evidence suggests that magmatic crystallization rather than replacement or multiple intrusion best explains the various internal structures of each pegmatite body. Pegmatite crystallized at relatively low temperatures, and in general the different types of pegmatites represent the crystallization-differentiation products of a large mass of granitic rocks. Variations in the original compositions and volatile pressures of each body modified to some extent the types of internal structures.

EVIDENCE FOR INTRUSION

The pegmatites of the Fourmile quadrangle have crosscutting contacts, inclusions of country rock, contact aureoles, and relatively fine grained border zones, all of which are commonly accepted as evidence of intrusion. These same features, however, are regarded by some geologists as being in accord with the theory of replacement or metasomatic crystal growth along zones of weakness (Ramberg, 1952, p. 248–56), and thus it is necessary to examine the field evidence in more detail.

Both concordant and discordant pegmatites have some features in the adjacent country rock that indicate forceful injection of the pegmatite. However, because of inadequate exposures, single beds generally cannot be traced completely around a pegmatite body; thus, proof is lacking that a space was opened by the pegmatite rather than that the pegmatite occupies space previously taken by schist. Adjacent to the concordant bodies the foliation of the wallrock bows around the nose of the pegmatite, much as it does around the ends of rhyolite dikes in the northern Black Hills described by Noble (1952; see especially fig. 1). At the noses of some of the pegmatite bodies there are minor crenulations in the wall rock, similar to those illustrated by Noble (1952, p. 35–41), that probably were caused by the compressive effect of the intrusive pegmatitic fluid. Along a very few of the discordant pegmatite bodies, such as pegmatite 71 (pl. 21, C–4), exposures of wallrock in the third dimension. hTe inclusions may be placed a distance approximately equivalent to the thickness of the discordant pegmatite.

Inclusions of country rock are rare, but where their orientation is at an angle to that of the wallrock, as in figure 86, there can be no doubt that they are true inclusions rather than unreplaced relicts of the country rock. Where supposed inclusions have the same orientation as the wallrock, it has commonly been shown by mining that they are connected to the wallrock in the third dimension. The inclusion may be fresh or greatly altered. The most common alteration is abundant tourmalinization in which biotite and muscovite are destroyed both in the early and the later stages. These mineralogic changes have little resemblance to any supposed process that can convert the inclusions into pegmatite, and the same holds for wallrock alteration along the outer borders of pegmatites. The formation of a biotite-rich layer in amphibolite adjacent to pegmatite and a plagioclase- or apatite-rich border zone in such pegmatite is clearly an example of reaction between the introduced pegmatite and the amphibolite wallrock. Apparently material has left the pegmatite, but there is no field evidence of a widespread replacement process that converts wallrock to pegmatite. Pegmatite in amphibolite does not differ in any significant way from pegmatite in the various types of schist.

The relatively fine grained character of border zones and their generally similar composition from one pegmatite body to the next suggest that they are early, rapidly crystallized products from fluids of essentially similar composition. The slight differences in border zones in different types of wallrock may be ascribed to slight contamination. In a sense they are equivalent to chilled contacts.

The physical evidence for distension and displacement, the fine-grained border zones, and the absence of large-scale replacement features indicate that the pegmatites have been intruded into the country rock. The scarcity of inclusions and the concordancy of most of the pegmatite bodies show that incorporation of country rock by stoping or other replacement processes was limited. Evidence for intrusion is necessarily of such a nature that it does not preclude the possibility that the pegmatites were formed in zones of weakness in the country rock by a "metamorphic-metasomatic" process of growth (Ramberg, 1952, p. 215–220, 248–256); this aspect of the problem is discussed with the origin of the zoned structures (p. 266–267).

TEMPERATURES OF FORMATION

The temperatures of crystallization must be known before the character of the pegmatite fluid and the course of crystallization can be discussed at length. Fortunately, the approximate temperatures at which the pegmatite crystallized can be deduced from the laboratory work of several investigators.

Weis (1953) studied the temperatures at which fluid inclusions disappear at atmospheric pressure in various minerals from zoned pegmatites in the Black Hills. He found that the temperatures vary widely, ranging from 216° to 515°C., but that most of them are between 300° and 400°C. These temperatures should be raised 200° to 300°C. when corrected for pressure (Weis, 1953, p. 694–695).

Grootemaat and Holland (1955) determined the sodium and potassium content of muscovite from various zones in the Peerless pegmatite, Keystone, S. Dak., and the temperature of crystallization from the solvus in the muscovite-paragonite system. Their data indicates temperatures of crystallization of about 500°C. for the outer zones and 350°C. for the core of the Peerless pegmatite. The sodium content of muscovite from the nearby Hugo pegmatite indicates temperatures of approximately 500° to 550°C. for the outer zones and as low as 250°C. for the inner zone (J. J. Norton, oral communication, 1957).

Bowen and Tuttle (1950, p. 495–496) showed that in the system $NaAlSi_3O_8$–$KAlSi_3O_8$–H_2O the highest point on the solvus is 660°C. and that above this temperature homogeneous feldspar is formed. Though the small amount of calcium in the plagioclase from the pegmatites would increase the temperature somewhat and though other variables would also change the relations, the 660°C. temperature is a fair approximation of the conditions in the Fourmile pegmatites. The previous description (p. 117–122) of the textural and structural relations shows that 2 separate feldspar minerals did crystallize in these pegmatites and thus that the temperature was less than 660°C.

The solvus also provides a means of determining temperatures of crystallization of the perthite if it can be assumed that the solvus, as determined experimentally on high temperature forms of albite and orthoclase, can be applied directly to the natural material. The uniform distribution of the plagioclase lamellae throughout the large perthite crystals, regardless of the nature of the associated groundmass minerals, as well as the slightly lower anorthite content of the lamellae as compared with the plagioclase of the groundmass, indicates that the perthite was initially homogeneous potassic feldspar, and that the albite formed by exsolution. The only recorded analyses of perthite from pegmatite of the Black Hills were made by Higazy (1949, p. 559–561). The sodium in Higazy's average perthite (based on 15 analyses) is equivalent to 22.8 percent albite and indicates a temperature of about 525°C. on the experimental solvus. The highest sodium content of the perthite specimens analyzed by Higazy indicates a temperature close to 600°C.

Perthite in the inner zones of some lithium-rich pegmatites in the Black Hills shows a marked decrease in the plagioclase component as compared to the perthite in the outer zones, and in some of these pegmatites the plagioclase lamellae are virtually absent. Higazy (1949, p. 559) found 3.19 percent Na_2O (equivalent to 28.8 percent albite) in perthite from an outer zone and 1.21 percent Na_2O (equivalent to 10.9 percent albite) from nonperthitic microcline from an inner zone of the Hugo pegmatite, Keystone, S. Dak. If sodium was present in excess when the nonperthitic microcline was formed, one can infer that the temperature of crystallization was significantly lower than that of the potassic feldspar of outer units.

Spodumene in some of the zoned pegmatites is also of value as a temperature indicator. Roy, Roy, and Osborn (1950, p. 159) found that the upper stability temperature of alpha spodumene (which occurs in the pegmatites) is 500°C, or somewhat higher at high pressures. They stated that the crystal form of any of the higher temperature polymorphs is so different from that of the alpha polymorph that there can be little doubt that the large spodumene crystals in the pegmatites formed initially as alpha spodumene at a temperature lower than 500°C.

Some of the pegmatites in the region south of Harney Peak and east of Custer contain sillimanite instead of muscovite. Although muscovite is in all the Fourmile pegmatite, it appears that some of the Black Hills pegmatite crystallized above the upper stability temperature of muscovite, which has been determined by Yoder and Eugster (1955, p. 263) to be 650°C at 15,000 psi water pressure (675°C at 30,000 psi). Above these temperatures, sillimanite and orthoclase (or microcline) are formed instead of muscovite. The absence of sillimanite in the pegmatites of the Fourmile area indicates crystallization below 675°C.

A number of independent lines of evidence thus demonstrate that most of the pegmatite crystallized at temperatures below 660°C, and probably at temperatures of about 500° to 600° C. Some of the inner zones in lithium pegmatites crystallized at temperatures below 500°C and possibly as low as 300°C. Furthermore, it seems likely that the inner zones of zoned pegmatites formed at lower temperatures than did the outer zones.

SIGNIFICANCE OF COMPOSITION

The pegmatites are rich in minerals containing SiO_2, Na_2O, and K_2O; they are poor in calcium and ferromagnesian minerals. The Al_2O_3 content is moderate and not greatly different from most granite. The composition is generally similar to that of granite; probably the major chemical differences are the greater abundance of volatiles—especially transient volatiles such as water that escaped during final crystallization—and the minor elements. Although precise chemical data are not available, the content of muscovite, tourmaline, and apatite suggests that the pegmatites were rich in water, boron, and possibly fluorine. In some of the zoned bodies such rare elements as Li, Be, Cb, Ta, and Cs are locally concentrated.

The effect of volatiles is largely one of lowering crystallization temperatures (Bowen, 1956, p. 302), and the inferred high content of volatiles is in accord with the low crystallization temperatures. Furthermore, the composition of the pegmatites, as shown in figure 94, falls in the low temperature part of the albite-orthoclase-quartz system.

There are all gradations in composition among the layered, homogeneous, and zoned pegmatites. The zoned pegmatites generally correspond in composition to the fracture filling units in layered and homogeneous pegmatites, so it may be assumed that the zoned pegmatites are the product of a late stage of crystallization. The SiO_2 content, higher in the zoned pegmatites than in the layered and homogeneous pegmatites, indicates that the quartz-feldspar field boundary (fig. 94) moved toward the quartz corner of the diagram as crystallization proceeded at progressively lower temperatures. This shift in the boundary may in part reflect the influence of chemical constituents other than quartz and feldspar, but it may also have been in response to changes in pressure, especially water pressure (Stewart, 1957).

ORIGIN OF ZONED AND LAYERED BODIES

The zoned pegmatites in the Fourmile quadrangle are characterized by a similar shape, mineral composition, and grain size from one zoned body to another. Previous detailed studies of many zoned pegmatites have shown that zones crystallized necessarily from the outer contact of the pegmatite inward to the core (Page and other, 1953, p. 18–24; Cameron and others, 1949, p. 98–105; Jahns, 1955, p. 1086); only a few aspects of the problem associated with zoned structures warrant review here. The most clearly defined evidence of the crystallization inward of the Fourmile quadrangle pegmatites is the transection of outer zones by fracture fillings that correspond in composition to inner zones. Tapered crystals, like that illustrated in figure 91, appear in many of the pegmatites. In general, the grain size of the pegmatite increases from the contact inward. Outer zones are generally rich in plagioclase-quartz and muscovite; these are succeeded by zones rich in perthite, quartz, and rarely amblygonite or spodumene; additional inner zones, if present, consist mainly of quartz, but may also contain plagioclase, perthite, spodumene, or lepidolite. This general sequence is constant not only throughout the Fourmile quadrangle and elsewhere in the Black Hills, but also in other pegmatite districts (Cameron and others, 1949, p. 61). This sequence is so universal that it is one of the more important parts of the body of evidence for origin of zoned pegmatites by progressive crystallization differentiation of a fluid, without significant addition or subtractions of material after intrusion.

Ramberg (1956, p. 210–211) argued that zonal structure is formed by metasomatic migration of material into joints and low pressure areas where the relative mobility of the chemical constituents would control their precipitation and that no large reservoir of fluid would be needed. He thought that silica and the alkalies are the most mobile material in the metamorphic rocks and that as some of the joints and other openings appear they are filled with quartz and feldspar. The process continues as the walls of the opening become further separated, but ultimately the available alkalies are exhausted and silica is the only material that continues to be introduced, and it then forms quartz cores.

A pegmatite formed in this way would have many of the structural characteristics of an intrusion, and it would also have gradual changes in composition from the outer to the inner parts that depend on the relative mobility and availability of the various constituents. Though it can be argued that this theory is in accord with many features that would otherwise be regarded as evidence for an intrusive magmatic origin, the pegmatites of the Black Hills have many characteristics that, taken as a group, are hard to explain. The Edison pegmatite (Page and others, 1953, p. 111), for example, has alternating wall zones and cores; it is much easier to explain as the product of a series of separate intrusions than as a product of metamorphic-metasomatic activity. The origin of similarly layered pegmatite bodies appears unamenable to explanation by the Ramberg proposal. It is difficult to understand the similarity of pegmatites that are in country rock as different in composition as amphibolite and quartz-mica schist; nor is it apparent why the distribution of the various types of zoned pegmatites in the southern Black Hills is so closely related to an apparent igneous center and yet has no significant relation to the stratigraphy of the metamorphic rocks and only a partial relation to the metamorphic intensity. The supposed parent fractures are not preserved beyond the ends of the pegmatites; instead, an induced schistosity wraps around the pegmatites. On the other hand, marginal faults resembling faults elsewhere at the borders of igneous intrusions are present. There are clear indications that temperatures of crystallization decrease inward in zoned pegmatites. The boudinage features that are normal to the plunge of pegmatites suggest abnormal tension in the plane of the pegmatites. Even if a body like the Tin Mountain pegmatite, containing many zones of different mineralogy, formed in accordance with the Ramberg proposal (assuming that the various elements do have the appropriate mobility and availability), this pegmatite has such irregular outlines that the maintenance of a consistent pattern of pressure gradients would be difficult. This problem becomes even more acute for highly irregular branching bodies like pegmatite 143 (pl. 21, F–4). It seems evident that the Ramberg proposal, despite its apparent plausibility, cannot adequately explain the origin of the pegmatites of the Black Hills.

The tendency for perthite to be concentrated in the upper parts and along the hanging walls of individual bodies, especially in perthite "hoods" (Page and others, 1953, p. 23) is hard to explain in any satisfactory manner. Somewhat similar concentrations of perthite have been noted in other pegmatite districts by Staatz and Trites (1955, p. 26–27), Thurston (1955, p. 37–38), and Jahns and Wright (1951, p. 28–30). In the Fourmile quadrangle such concentrations appear not only in zoned pegmatites but in other types of pegmatite, especially those having flat dips. Though this concentration of potassium-rich material in the upper parts and sodium-rich material in the lower parts of pegmatites suggests a gravitational control, there is not sufficient evidence to postulate liquid immiscibility or crystal floating or sinking.

The Hugo pegmatite, Keystone, S. Dak., contains evidence that this transfer of potassium- and sodium-rich materials took place in the fluid prior to crystallization (J. J. Norton, oral communication, 1957). This pegmatite is separated into two segments that are directly in contact with each other in only a small area; thus they must have crystallized virtually independent of each other. Though each has a perthite hood, potassium is much more abundant in the segment that lies at a higher elevation and sodium is more abundant in the lower segment. The movement of these materials must have taken place in the liquid before the avenue connecting the two segments was blocked by crystallizing material. Though the reason is not clear, it is evident that upward enrichment in potassium is more characteristic of zoned pegmatites than other types. The fluid from which zoned pegmatites formed may have been especially rich in volatiles. Movement of volatiles could have carried potassium ions (whose specific volume is less than sodium ions) to the upper part of the pegmatite, and in a few places could have carried potassium into the schist, as indicated by table 10 and by the retrograde alteration of staurolite and sillimanite to mica.

The main differences in structure between the layered and the zoned pegmatites are (a) that the layers do not extend continuously around a single body, but gradually disappear near the ends of lenticular bodies and (b) that instead of changing composition in a continuous sequence, the layers contain many repetitions of mineral assemblages. The changes in mineral content between fine-grained and coarse-grained layers, however, are very similar to the changes in the outer zones of zoned pegmatites. The fine-grained layers belong mostly to assemblages 1 and 2 and the coarse-grained layers the assemblages 3 and 4 (table 6). Ordinarily a fine-grained layer is adjacent to the wallrock and is followed by a coarse-grained layer, so that the sequence at the contact is the same as in a zoned pegmatite, but for some reason the sequence in the layered pegmatites is then repeated many times. Line rock superimposed on this larger rock layering is also characterized by the repetition of layers of similar composition.

The repetition of the layers must have resulted from some process which periodically changed the pressure, temperature, or the composition of the crystallizing fluid. The number of the layers and the uniformity of the layering in many different pegmatites of various sizes and shapes offers little support to the theory that some repetitive mechanism affecting temperature or composition caused the layering. It is true that some part may have been played by (a) local repeated injection of new material, thus upsetting equilibrium, or (b) by failure of convection or diffusion to maintain equilibrium at the border of fluid and crystalline material. Rhythmic layering caused by failure of diffusion to transfer constitutents fast enough to the crystal-liquid boundary cannot be disproved, but the writer knows of no rock whose layering has been proved to have been caused by such a process.

Repetitive changes in volatile pressure may have caused some of the observed relations. The local alteration of the wallrock by minerals rich in volatiles indicates that volatiles were able to leave the pegmatite system. The volatile pressure may even have exceeded the confining pressure inasmuch as Morey (1956, p. 722) has shown that water pressure can increase to 2,200 bars (equivalent to a rock load of 8 to 9 km) with continued crystallization and cooling in the system H_2O-Na_2O-Si_2O_5. Although volatile pressure greater than confining pressure might provide an excellent method for rhythmic loss of material (like a safety valve on a steam engine) the excess pressure is not needed if the volatiles can escape into the permeable wallrock where volatile pressure is presumably much lower. As the volatile pressure alternately increases and then decreases as volatiles are expelled, the equilibrium relations of the fluid to the mineral phases may be expected to change. Bowen and Tuttle (1950, p. 496–498) found, for example, that the minimum melting point on the system orthoclase-albite-water moves toward the albite end as water pressure increases. Similar changes in the composition at the minimum melting point have been found in the various water-feldspar-quartz systems (Stewart, 1957, p. 215); and the eutectic composition of the diopside-anorthite system is greatly shifted by added water pressure (Yoder, 1954, p. 107).

The effect of a change in volatile pressure on the crystallization of a layered pegmatite can be illustrated in a hypothetical simple eutectic system containing components A and B and a volatile. Suppose A and B are crystallizing together at the eutectic, under high volatile pressure, and the volatile pressure is suddenly lowered, causing the eutectic to shift away from end point A. Component A will now crystallize rapidly until, by subtraction of A from the liquid, the composition is moved to the new eutectic, and B resumes crystallization. Layers of different composition will result. During the period when only A was being precipitated, crystals of B would be out of equilibrium and would or could in part be resorbed by the liquid.

Though the natural system is much more complex than the simple version presented here, abundant evidence suggests that layering is related to the behavior of the volatiles. Layers in line rock that are rich in minerals containing volatiles such as muscovite and tourmaline alternate with layers containing anhydrous minerals. Layers of coarsely crystalline material that suggest a low viscosity and high volatile content alternate with fine-grained material that suggests dry conditions or rapid crystallization to restore equilibrium.

Single layers, whether of the thin line rock type or of the thicker layers, extend for only short distances, rather than for the entire length of a pegmatite. Consequently the mechanism here suggested would operate only locally and would not establish a cycle that would apply throughout the crystallizing part of the pegmatite. Possibly very slight changes in the volatile content greatly influence the rate of diffusion of material through the liquid to the crystal-liquid interface. At any place along the contact between crystals and liquid, equilibrium conditions would be constantly changing, and these changes would be accompanied by changes in the crystalline phases deposited. The resulting layers could be short like those in the line rock, or very long, as in some of the more coarsely layered material.

The constant changing of volatile pressures during crystallization of layered pegmatites would then be the main genetic difference between layers and zones. Zones would crystallize under more uniform conditions in which successive mineral assemblages of different composition could form in the sequences indicated by table 6.

Other layered igneous rocks may have been produced by changes in volatile pressure. Ussing (1912, p. 361) believed that layers in the Ilimausak batholith in Greenland were caused by changes in pressure resulting from volcanic activity. Each presumed change in pressure resulted in crystals which were then separated by gravity into layers. Layered structural features, both planar and folded, occur in the Skaergaard intrusive body in Greenland, but are considered by Wager and Deer (1939) to have been originated by convection currents and flow banding combined with gravitational sinking of crystals rather than by any significant change in volatile pressure. It is interesting that Ussing and also Wager and Deer concluded that the intrusions studied by them had had unusually low viscosities. Viscosities relatively low because of high volatile content are also commonly inferred for pegmatite.

SIGNIFICANCE OF REGIONAL DISTRIBUTION OF TYPES OF PEGMATITES AND OF PEGMATITE MINERALS

Regional distribution of various types of pegmatites has been recognized in many pegmatite districts in the world (Heinrich, 1953; and Varlamoff, 1954; 1955). Ordinarily the pattern is not clearcut, and it may depend on many factors other than distance from the source magma (Cameron and others, 1949, p. 81).

In the Black Hills there is a regional distribution of the types of pegmatites and of pegmatite minerals, though this regional pattern is, in many respects, vaguely defined. Layered pegmatites are mostly in areas of abundant pegmatite, homogeneous pegmatites tend to be farther out, and zoned pegmatites are largely in the outer parts of the pegmatite region. All these types of pegmatites, however, occur together as in many parts of the Fourmile quadrangle.

Among the various types of zoned pegmatites, a fairly well defined pattern can be recognized from the distribution of the mines shown in figure 78. Sheet mica mines are mainly in areas of abundant pegmatite, especially on the southwest side of the Harney area; feldspar mines are farther out; and the lithium and beryl-scrap mica mines are near the edge of the pegmatite-bearing area.

A similar pattern can be recognized in the distribution of the individual minerals, as indicated by table 13 and figure 95. Zoned pegmatites in the outer part of the pegmatite area contain most of the cleavelandite, lithium minerals, beryl, lithiophilite-triphylite, and columbite-tantalite. In a few places lithiophilite-triphylite appears in homogeneous pegmatites and beryl in homogeneous and layered pegmatites in the outer part of the area. Quartz, plagioclase (except cleavelandite), potassic feldspar, mica, and tourmaline have a more nearly ubiquitous distribution. It is notable, however, that among the various types of potassic feldspar, there is a trend from graphic perthite to nongraphic

TABLE 13.—*Distribution of minerals, based on pegmatite distribution shown in figure 78*

[Present in: A, all types of pegmatites; B, generally all types of pegmatites; C, mainly layered pegmatites, but locally in other types of pegmatites; F, fracture fillings in homogeneous and layered pegmatites (*F*, rarely); H, Homogeneous pegmatites (*H*, rarely); L, layered pegmatites (*L*, rarely); Z, zoned pegmatites.]

Mineral	Pegmatite abundance				
	More than 50 percent of area	Number per square mile			
		200–300	100–200	50–100	0–50
Quartz	A	A	A	A	A
Plagioclase	A	A	A	A	A
Muscovite	A	A	A	A	A
Perthite (graphic)	B	B	B	B	C
Garnet	B	B	B	C	C
Biotite	B	C	C	C	C
Tourmaline	A	A	A	A	B
Perthite	Z, F	Z, F	Z, F	Z, F, H	Z, F, H, *L*
Beryl	Z, *F*	Z, *F*	Z, *F*	Z, F, H	Z, F, H, *L*
Lithiophilite-triphylite	Z, *F*	Z, *F*	Z, *F*	Z, *F*	Z, *F*, *H*
Amblygonite	-	-	Z, *F*	Z, *F*	Z, *F*, *H*
Cleavelandite	-	-	Z	Z	Z, *F*
Columbite-tantalite	-	-	Z	Z	Z, *F*
Spodumene	-	-	Z	Z, *F*	Z, *F*
Microcline (nonperthitic, coarse-grained)	-	-	-	Z	Z
Lepidolite	-	-	-	Z	Z
Cassiterite	-	-	-	Z	Z

perthite to nonperthitic microcline from the inner to the outer part of the pegmatite area.

Trends indicated by mineralogic changes from layered through homogeneous to zoned pegmatites have many similarities to trends in the sequence of sheet mica, feldspar, beryl-scrap mica, and lithium pegmatites and also to trends from the outer to the inner parts of zoned pegmatites. Thus, table 14 shows that minerals that are more abundant in layered than in other types of pegmatites are also more abundant in sheet mica pegmatites than in other types of zoned pegmatites and are concentrated in the outer zones of zoned pegmatites. Conversely, minerals that are characteristic of inner zones are also likely to be most abundant in the lithium-rich zoned pegmatites and the zoned pegmatites as a type rather than in the homogeneous or layered types. A few minerals like sheet muscovite that are concentrated in outer or intermediate zones may also be more abundant in zoned pegmatites as a group than in layered or homogeneous pegmatities, but this obviously is to be expected.

The similarity of the trends suggests that the distinctive features of the various types of pegmatites are controlled by chemical and physical gradients. The temperature of crystallization may be very important as suggested in table 14. Compositional changes in the original fluid must be very important and one can infer that layered pegmatites were generally formed early in fractionization, homogeneous pegmatites somewhat later, and zoned pegmatites last. This sequence cannot be based firmly on crosscutting relations, and there is certainly no evidence for a difference in time in the emplacement of different pegmatites. On the other hand, the similarity of zoned pegmatites to fracture fillings in homogeneous and in layered pegmatites and of homogeneous pegmatites to some fracture fillings in layered pegmatites does not leave much room for doubt as to the sequence, although there may be considerable overlap. The presence of such minerals as beryl, amblygonite, and lithiophilite-triphylite in homogeneous and layered pegmatites in the outer parts of the pegmatite area suggests that these bodies may have formed at a fairly late stage in fractionation, perhaps later than some of the zoned pegmatites. Previous discussion has already shown that layered pegmatites may owe their internal structures to conditions that permitted the periodic escape of volatiles. Thus the distribution pattern exhibited by the various types of pegmatites may reflect the interaction of many chemical and physical forces and it is not surprising that this pattern is somewhat vaguely defined.

The latest differentiates have the lowest temperatures of crystallization and probably had a low viscosity. They are able to move farther outward and upward from the parent source before crystallization begins. Subsequent erosion produces a peripheral pattern of the zoned pegmatites and rarer minerals.

The trend of crystallization differentiation in some zoned bodies suggests the development of a silica-rich end product or core, notwithstanding the lack of supporting experimental evidence. Therefore, the correlation between zoning in individual pegmatites and regional distribution indicates that a belt of quartz pegmatite bodies should appear outside the belt of zoned pegmatite bodies. Some of the larger massive white quartz veins in the northwestern part of the quadrangle may represent late-stage differentiates corresponding to pegmatite (see also Jahns, 1955, p. 1104–1106; Andersen, 1932, p. 34–42).

ECONOMIC GEOLOGY

The chief mineral production of the Fourmile quadrangle, as in the rest of the southern Black Hills, has been from pegmatites. Some gold has been recovered

TABLE 14.—*Relative distribution of individual minerals in types of pegmatites, types of zoned pegmatites, and pegmatite zones*

[Distribution: A, abundant; MA, most abundant; NO, not observed; R, rare.]

		Inferred relative temperatures of crystallization			Mineral distribution									
						Muscovite		Plagioclase		Potassic feldspar				
		High	Intermediate	Low	Biotite	Sheet	Scrap	Blocky	Platy (cleavelandite)	Graphic perthite	Perthite	Nonperthitic microcline	Lithium minerals	Quartz
Pegmatite types	Layered	×			MA	NO	A	A	NO	MA	R	NO	NO	A
	Homogeneous		×		R	R	A	A	R	A	A	NO	R	A
	Zoned			×	R	MA	A	A	MA	R	MA	MA	MA	MA
Types of zoned pegmatites.[1]	Sheet mica	×			MA	MA	A	MA	NO	MA	R	NO	NO	A
	Feldspar		×		R	A	A	A	R	R	MA	R	R	A
	Beryl-scrap mica		×?	×?	R	A	MA	A	R	R	A	R	A	A
	Lithium			×	R	R	A	R	MA	R	A	MA	MA	MA?
Pegmatite zones	Outer	×			MA	MA	A	MA	R	MA	R	NO	NO	A
	Intermediate		×		NO	R	A	A	R	NO	MA	R	R	A
	Inner			×	NO	NO	R	R	MA	NO	R	MA	MA	MA

[1] From figure 78.

from placer and lode deposits. Pegmatite mining in the area started about 1879–1880 when the New York pegmatite was worked for mica. Since then, sheet and scrap mica, feldspar, beryl, columbite-tantalite, pollucite, and lithium minerals have been mined. Mica mining increased greatly during World War II but declined shortly thereafter. It increased again in the early 1950's as a result of the establishment of a Government purchasing depot at Custer. Production of beryl and the lithium minerals increased after the start of World War II. The mining of potash feldspar, scrap mica, and rarer minerals doubtless will continue, although the finding of new deposits will be more difficult than in the past.

PEGMATITE DEPOSITS

The homogeneous and layered pegmatites are of only slight economic interest at present because all are too low grade to be profitably mined, despite their content of industrial minerals. Neither feldspar nor mica is segregated into high-grade deposits as they are in some zoned pegmatites. Many homogeneous and layered pegmatites have been propected along fracture fillings and coarse-grained units, but with unsatisfactory results. The potential resources of potash feldspar and scrap mica in these are large, but resources of sheet mica are small. Beryl resources in these pegmatites are probably relatively large compared to present United States production, but their recovery will be difficult because of the very low grade and the small size of crystals.

Most of the present minable material is in the zoned pegmatites. The main products have been sheet mica and potash feldspar, but scrap mica, beryl, and lithium minerals have also been important. Table 15 shows that the mica is predominantly in wall zones, potash feldspar in intermediate zones, and lithium minerals in inner zones. Beryl appears in a variety of units. All these minerals mined in the Fourmile quadrangle have been concentrated only by hand sorting.

TABLE 15.—*Summary of mineral production from the major pegmatite mines in the Fourmile quadrangle*

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

Mine (pl. 21)	Minerals and zones from which mined						Comments on size and ore grade of deposits
	Potassic feldspar	Sheet mica	Scrap mica	Beryl	Lithium minerals	Cb-Ta minerals	
Big Tom			X(WZ)	X(WZ,IZ)			Small; low grade.
Corky	X(C)		X(WZ)	X(WZ,C)			Small; could have sheet mica at depth.
Gayle-Royal Flush	O(IZ,C)		O(WZ,C)	X(WZ,C)			Small.
Helen Beryl	X(IZ)		O(WZ,IZ)	X(WZ,IZ,C)	X(IZ,C)	O(IZ)	Large; large beryl resources but difficult to recover because of small grain size.
Lithia Lode				X(C)	X(C)		Small.
MacArthur		X(WZ)					Small; low grade.
Michaud Beryl				X(WZ,IZ)		O(IZ)	Small; grade and size decrease at depth.
New York	O(IZ)	X(WZ)		O(WZ)			Large; but mostly mined.
Pleasant Valley	X(IZ)		O(WZ)	X(WZ,IZ)			Small; thinning at depth.
Punch		X(WZ)					Small; low grade.
Rainbow No. 4	X(IZ)						Small.
Red Spar	X(C)						Small; mostly mined.
Rock Ridge	X(C)						Medium size; should be more feldspar at depth.
Tin Mountain	X(IZ)		X(WZ)	X(WZ,IZ)	X(IZ,C)	X(IZ,C)	Large reserves.
Tip Top	X(C)						Mostly mined.
Volcano			X(WZ)	X(WZ,IZ)			Small; low grade.
Warren Draw	X(IZ)						Feldspar is iron-stained; possible unexposed lithium zones.
White Spar	O(IZ)	X(WZ)		O(WZ)			Deposit mostly mined.
Wright Mica Lode	X(IZ)		X(WZ)	X(WZ,IZ)			Small; north segment untouched; possible sheet mica at depth.

MICA

Sheet mica has been the most valuable product from the Fourmile quadrangle because of the large production from the New York mine. The production from the New York mine has been estimated at between 1,500 and 2,000 tons of crude mica (Page and others, 1953, p. 35). The New York is much the largest sheet mica mine in the Black Hills and is a large mine by world standards. Scrap mica is moderately abundant in all types of pegmatites, but is concentrated in the outer zones of zoned pegmatites where it can be profitably mined as a byproduct mineral in conjunction with the mining of feldspar, beryl, and other minerals.

In the Fourmile quadrangle sheet mica is produced entirely from wall zones of zoned pegmatites (table 15), but elsewhere in the Black Hills it does occur in other zones. The mica content of the wall zone deposits generally is between 10 and 15 percent, but the recovery of crude sheet mica during mining is rarely more than 5 percent. In some wall zones where the quality of the sheet mica is high, the recovery in a profitable operation may be as low as 1.5 percent.

The geology of the New York pegmatite and its sheet mica deposits has been described in detail by Page and others (1953, p. 163-170). The mine was unwatered and reopened in the fall of 1956 and several tons of crude mica that yielded fairly low quality sheet mica were mined on the 300-foot level from the main mica-bearing roll. The mine was still in operation in the fall of 1957.

Other mines that produced sheet mica during World War II include the White Spar, the Punch, and the MacArthur. The White Spar produced more than 10 tons of sheet mica during World War II and was a major producer (Page and others, 1953, p. 217–220). It has been inactive since 1945.

The Punch mine was reopened in 1952 and produced a little more than 3 tons of crude mica. The MacArthur mine described by Page and others (1953, p. 146–148) closed after World War II. Production of sheet mica from other pegmatites has been limited to a few tens or hundreds of pounds recovered during feldspar or beryl mining. The Corky and some of the other feldspar-bearing zoned pegmatites may contain sheet mica below the exposed perthite-rich zones.

Most of the zoned pegmatites of the area are potential sources of scrap mica as a byproduct. Scrap mica is largely in wall zones, but may also occur in intermediate zones or cores. Most of the scrap mica produced from the Warren Draw pegmatite occurs throughout the core in scattered tabular units of "bull mica," which are a few feet long and contain 40 to 90 percent small felted muscovite books that average 0.8 inch in diameter. The rock recovered during feldspar mining probably averages about 1 percent in scrap mica, but recoveries of 5 percent of the total mined rock are fairly common.

The New York mine still contains considerable reserves of scrap and sheet mica, although the best and most easily recovered mica has already been mined. In addition, the Tin Mountain and Warren Draw mines have large reserves of scrap mica that could be recovered as a byproduct in mining lithium minerals and feldspar. The inner part of the wall zone and outer part of the core of the Corky pegmatite are also unusually rich in scrap mica, but the pegmatite is a relatively small one. A few tens of tons of scrap mica probably could be recovered from most feldspar-bearing pegmatites in the area.

Although sheet mica has been the most valuable product from the area, future production is not promising because the easily recoverable reserves of the main source, the New York mine, have largely been depleted. Nor is the outlook promising for the discovery of any similar rich sheet-mica deposits in pegmatites now exposed in the area.

FELDSPAR

The greatest present mining activity in the Fourmile area is in the production of perthite, the "spar" of the local miners. The producing mines (some operated only intermittently) of the area are the Tip Top, Helen Beryl, Royal Flush, Corky, Red Spar, and Warren Draw as well as some of the smaller unnamed zoned pegmatites and fracture fillings within layered and homogeneous pegmatites. Most of the perthite occurs in intermediate zones or in the cores of larger pegmatites (table 15) where perthite crystals are as much as 10 feet in length. The upper parts of the intermediate zones of the Tin Mountain and Warren Draw pegmatites consist of nearly pure perthite "hoods." The feldspar from the various mines is medium to coarse grained and is easily recovered by hand cobbing. Scrap mica and beryl are byproducts.

The feldspar is of good quality and most of it is acceptable at the local mills, though some from the Warren Draw pegmatite is relatively rich in iron and cannot be sold. The graphic perthite in most of the homogeneous and layered pegmatites brings a lower price than nongraphic perthite and ordinarily cannot be mined profitably.

The Tip Top, Tin Mountain, and Warren Draw pegmatites have moderately large reserves of feldspar. Feldspar can also be produced from most of the small zoned pegmatites of the area, but the reserves of most of these pegmatites do not exceed a few thousand tons. A few of the larger unzoned bodies, especially a homogeneous one like pegmatite 80 (pl. 21, E-3), which contains 50 percent perthite, may ultimately be mined. The differences between these and zoned pegmatites are gradational in nature and the feasibility of mining them depends more on economic factors than on their geologic characteristics.

BERYL

Beryl is in nearly all the zoned pegmatites, but only in a small percentage of the other two types. The beryl in layered and in homogeneous pegmatites is relatively fine grained and commonly occurs only in fracture fillings or in coarser grained units. Some beryl, however, is disseminated in homogeneous pegmatite in the western part of the quadrangle. It is almost impossible to make estimates of the grade of the beryl in the homogeneous and layered pegmatites, but none of these pegmatites are believed to contain more than 0.1 percent beryl. Samples from homogeneous and layered pegmatites northeast of Custer (table 7) average about 0.001 percent BeO in spectrographic analyses. This would be equivalent approximately to 0.01 percent beryl, providing the BeO is not held in other minerals. Recovery of beryl that may exist in even the richest of these pegmatites would have to be by milling and not by hand methods.

Beryl from zoned pegmatites is commonly coarse grained and concentrated in certain zones. It is thus more amenable to mining and hand cobbing than that from other pegmatites. Most of the crystals that are mined are less than 8 inches in diameter and 1 foot in length. Nearly all the beryl is in the wall and the intermediate zones (table 15), but some has been found in all zones (table 7). The grade of a particular zone is seldom uniform, and rich concentrations are separated by nearly barren rock.

The largest and richest deposits of beryl occur in the inner parts of mica-rich wall zones and extend to the outer part of the adjacent inner zones. Beryl rarely occurs in quartz cores but is common in perthite-quartz or quartz-perthite cores. Fine-grained beryl occurs in border zones but is of no economic importance. The lithium-bearing intermediate zones of the Tin Mountain pegmatite (table 7) contain abundant minable beryl. In general the beryl crystals increase in size inward from the outer pegmatite contact and are largest and most easily mined in coarse-grained intermediate zones and in cores.

Determination of the grade of the beryl in the different pegmatite deposits is difficult because of the large grain size and the small amounts of beryl that are present. Probably one of the better methods of determining grade is by measuring the area of visible crystals in a representative exposure of a zone or several zones. Percentages may also be computed from production. It is rare for individual zones to contain 1 percent beryl; the average for an entire pegmatite body is generally less than 0.2 percent. The mined parts of the Michaud Beryl pegmatite, mostly from the wall zone, contained about 2 percent beryl. However, several diamond-drill holes that crossed the pegmatite at depths of 50 feet or more found only a trace of beryl, and it is questionable whether the actual grade of the entire pegmatite body is as much as 0.5 percent. Most of the beryl in the eastern part of the Warren Draw pegmatite (pl. 23) is confined to a 2-foot thick unit along the inner edge of the wall zone and the outer edge of the perthite-rich inner zone. Although the grade in this unit is estimated at 0.5 percent beryl, the grade for the entire pegmatite is estimated at less than 0.05 percent.

Refraction indices of the beryl from the Fourmile area (using a graph by W. T. Schaller and R. E. Stevens, written communication, 1948) indicate an average content of 12.5 percent BeO. Impurities that are not removed during hand cobbing tend to lower the BeO content of the marketed beryl concentrates to 11.5 to 12.0 percent.

Most of the larger and richer beryl deposits, such as those in the Helen Beryl and Tin Moutain pegmatites, have already been described (Page and others, 1953). The Helen Beryl pegmatite has the largest potential reserves of beryl in the area. Most of the beryl, however, is fine grained; the grade is only about 0.3 percent, and recovery would require some method of milling. Much of the beryl in the Tin Mountain pegmatite, on the other hand, is sufficiently coarse grained to be recovered by hand cobbing. The wall and intermediate zones of this pegmatite (table 7) are relatively rich in beryl and contain the largest reserves of beryl recoverable by mining methods presently used in the area. The New York pegmatite has a fairly large quantity of beryl in the wall zone, but it is too low grade to warrant mining for beryl alone.

Most of the other zoned pegmatites have a low beryl content and only a small tonnage of beryl-bearing rock. Pegmatites such as the Big Tom, Corky, Gayle and Royal Flush, Michaud Beryl, Red Spar, Wright Mica Lode, and Warren Draw have inferred beryl reserves of several tons or several tens of tons each.

LITHIUM MINERALS

Amblygonite is the most widespread lithium mineral as shown in figure 95. It is mainly in intermediate zones and in cores. Amblygonite in the Lithia Lode, Royal Flush, Corky, and some of the other small pegmatites is of no economic significance because of the small size both of the crystals and of the host pegmatite. At the Tin Mountain pegmatite considerable ambylgonite occurs as relatively large and easily recoverable crystals in the second intermediate zone (table 7). The amblygonite in the fracture fillings of the Warren Draw pegmatite is of importance only if it indicates concealed zones of other lithium minerals.

Spodumene has almost the same areal distribution as amblygonite, and most of it occurs within amblygonite-bearing units in zoned pegmatites. The Tin Mountain mine is the only one with large reserves of spodumene.

Lepidolite is in inner zones of a few pegmatites but only in the core of the Tin Mountain pegmatite is it rich enough for mining.

OTHER MINERALS

Other minerals of possible economic interest include columbite-tantalite, microlite, tapiolite, pollucite, and cassiterite, but all these are rare. Only columbite-tantalite is of economic significance in mines other than the Tin Mountain pegmatite. A small amount of columbite-tantalite is recovered as a byproduct from feldspar mining, but production from most of the mines is no more than a few tens of pounds.

Small bladed crystals of columbite-tantalite commonly are in aggregates that rarely exceed 2 inches in maximum diameter. These are commonly not recovered by hand cobbing; they are most common in the inner parts of wall zones, in intermediate zones, or along core margins. Grade and reserve estimates on the small quantities of columbite-tantalite are extremely difficult to make. The small Michaud Beryl pegmatite yielded approximately 2 pounds of columbite-tantalite per ton of rock during the mining of the beryl-rich upper part of that pegmatite, but the grade of the entire pegmatite is much lower. The grade elsewhere is less than a pound per ton of pegmatite, even in the Tin Mountain pegmatite.

The Tin Mountain mine has been the source of several tons of pollucite, and it also contains microlite and cassiterite. Microlite in very small crystals is locally concentrated around spodumene crystals. Cassiterite is largely in the wall zone.

GUIDES FOR EXPLORATION OF PEGMATITE DEPOSITS

Exploration should generally be restricted to zoned pegmatites, especially if a large or long-continued mining operation is anticipated. Layered and homogeneous pegmatites are a risk though a small operator may mine for a short period from a large fracture filling or segregation in either of these types.

Most of the zoned pegmatites that have been found in the past are in the outer part of the pegmatite-bearing area, and it may be safely assumed that the same will be true of new discoveries. Even mica pegmatites that lie within the 100 isopleth of figure 78 are mainly in those areas less than a square mile in size where pegmatites of all kinds are relatively sparse.

Zoned pegmatites tend to occur in clusters. They are especially abundant, for example, in the vicinity of the Tin Mountain mine. Furthermore, the zoned pegmatites in any small area may have distinctive structural relations. In the northwest part of the Fourmile quadrangle, 19 of the 28 pegmatites cutting amphibolite are zoned. These are near the outer limit of the pegmatite area, however, and the concentration in amphibolite may not be valid elsewhere. Another example is in an area of about 1 square mile around the Corky pegmatite, where nearly all the pegmatites that strike northwest and dip steeply northeast are zoned and those that have other attitudes are not zoned. In a broad sense, oval, teardrop, and pipelike shapes are more characteristic of zoned pegmatites than of unzoned pegmatites throughout the area.

Zoning or its absense is readily observed where a pegmatite is well exposed, but not where only the outer parts of a pegmatite are exposed; there other evidence must be sought. Layered outer units surrounding zoned pegmatites can be especially deceptive, as at the Warren Draw mine. In general, all minerals in zoned pegmatites are coarser grained than the same minerals in layered and homogeneous pegmatites. Most of the zoned pegmatites are richer in quartz, plagioclase, and muscovite near their contacts with country rock than are the homogeneous and layered pegmatites. Plagioclase grains an inch or more in average diameter are especially common in a sheet mica-bearing pegmatite. The presence of cleavelandite strongly suggests a zoned body. The occurrence of beryl or lithium minerals, apart from small crystals in fracture fillings, is also indicative that a body may be zoned.

Fracture fillings are a helpful clue as to what minerals may be encountered within inner zones. At the Tin Mountain pegmatite, for example, the spodumene-rich inner units were not exposed at the surface prior to mining, but the outcrop did contain a spodumene-bearing fracture-filling that could have been used as evidence that lithium-bearing inner units might be found. Spodumene-bearing fracture fillings at the Warren Draw pegmatites can be interpreted in the same way.

Some of these guides for exploration and others presented by Page and his associates (1953, p. 43–44, 53, 57–58), were used in 1952 to evaluate the potentialities of the Beecher No. 3 pegmatite, south of Custer. Many of the pegmatites in the surrounding area are zoned. The Beecher No. 3 pegmatite itself, although poorly exposed, had a well-developed wall zone rich in medium-grained muscovite, plagioclase, and quartz and an inner zone rich in perthite and spodumene-bearing fracture fillings. Beryl was found to be relatively abundant in float but had been overlooked previously because of its heavy iron staining. Subsequent exploration and mining in 1952–1954 led to the production of more beryl than has been obtained from any other mine in the Black Hills except the Peerless and the Bob Ingersoll. The outer edge of a spodumene-bearing zone was apparently being uncovered by mining at depth in 1957.

GOLD DEPOSITS

Gold has been mined in the Fourmile quadrangle both from placer and lode deposits. The original discovery of gold in the Black Hills (Newton and Jenney, 1880, p. 16) was made in 1874 on French Creek east of the Fourmile area. After the Black Hills area was opened for prospecting in 1877, placer claims along most of the streams were staked out. Later, lodes were located and a few small mills were erected late in the nineteenth century. The total production was not large and mining soon declined. It was revived somewhat in the 1930's. Allsman (1940, p. 132–138) described several properties in the area that were being mined in 1938. Since the 1940's, exploration has been limited mainly to assessment work on unpatented claims. The value of gold produced in the Fourmile quadrangle is not accurately known, but is surely small.

Placer deposits are limited to present-day stream channels and to gravels of the White River group along Pleasant Valley. The latter are low grade, so far as known, but have a volume of several million yards. Most of the placer mining has been along Lightning Creek and Warren Gulch near U.S. Highway 16, probably in the 1930's, but at an earlier date there may have been work along Pleasant Valley and Hay Creeks.

Most of the lode deposits are in the north-northeast set of quartz veins, where they cut either amphibolite or schist. Those in schist are commonly bordered by graphite and tourmaline (fig. 77). Exposures of the veins and ore in dumps are oxidized but presumably the unoxidized veins contain iron sulfides and arsenides.

The May gold mine, west of Lightning Creek (pl. 21, B–4), is in small east-trending quartz veins that cross amphibolite, but the workings are now inaccessible. The gold was taken from a small high-grade quartz stringer (Allsman, 1940, p. 133). The production is not known.

The Ruberta mine, northeast of Fourmile (pl. 21, E–2), is on a north-northeast-trending quartz vein that is several feet thick and less than 200 feet long. The mine workings are inaccessible. The wall rock along the vein is somewhat brecciated and altered. Graphite and tourmaline are locally concentrated along the edge of the vein and in the wallrock. According to Allsman (1940, p. 139), the mine originally had an amalgamation and cyanide mill. The value of total production of gold was probably between $10,000 and $20,000 from ore having a grade of less than $10.00 per ton.

Similar small north-northeast-trending quartz veins cut by pegmatite occur about 0.3 mile due east and at 0.4 mile southeast of Fourmile. A claim at the latter workings, called the Monatomic, is owned by Mae Nevins of Custer, S. Dak. These veins are similar to those at the Ruberta, but are smaller in size. There has been no known production.

The Caledonia mine, northwest of Castle Rock (pl. 21, C–2), has extensive workings. Little is known of the history of the mine and the main drift was locked; however, a plat of U.S. Mineral Survey 2134 indicates that the main drift is nearly 500 feet long and has 400 feet of drifts and crosscuts. There are numerous surface prospects and trenches. Material in the dump indicates that the workings are entirely in amphibolite. Small quartz veins are visible and presumably contain low-grade gold ore. The production is not known.

Many small quartz veins in amphibolite bodies near Warren Gulch and north of Tin Mountain mine contain gold. Arsenopyrite is locally disseminated in the amphibolite near these veins and stringers Many of these have been prospected, but no production has resulted.

OTHER DEPOSITS

Deposits of sand that are possibly suitable for the sand-frac process in the petroleum industry occur in sandstone in the upper part of the Deadwood formation. The quartz grains in the sandstone are exceptionally well rounded and free from impurities. According to Mr. Montgomery Heumphreus (oral communication, 1957), preliminary tests on the physical characteristics of the screened sand were favorable and the upper 10 to 15 feet of the Deadwood formation was being explored in Layton Canyon just west of the Fourmile quadrangle.

The well-rounded upper sandstone is largely concealed in the Fourmile quadrangle, but there are blocks and float from the unit in several places. Natural exposures are inadequate to determine the thickness and extent of the sandstone. This sandstone is 15 to 20 feet thick, however, on the west side of the Berne quadrangle and is relatively well exposed in valleys bottomed in the upper part of the Deadwood formation in the area immediately west of the Berne quadrangle.

Some copper minerals were noted in an old prospect pit located 3,600 feet N. 42° W. from Sevenmile school (pl. 21). Dump material contained malachite, calcite, and hematite filling cavities in brecciated garnetiferous schist from the Mayo formation.

An occurrence of copper and uranium was noted in a single piece of float found approximately 500 feet southeast of the intersection of Lightning Creek and Pleasant Valley Creek. Malachite, cuprite, and uranophane(?) filled interstices in sandstone from the Deadwood formation. The specimen contained 1 percent uranium (analysts: C. G. Angelo and H. H. Lipp). A brief survey of the immediate area with scintillometer failed to show where this float specimen had originated, but there was considerable cover over the Deadwood formation.

PEGMATITE MINES AND PROSPECTS

Most of the larger and more productive pegmatite mines of the Fourmile quadrangle have been described in detail. Descriptions are not repeated here except where more detailed examinations than those previously reported were made.

The Gayle-Royal Flush, Helen Beryl, Michaud Beryl, New York, Pleasant Valley, Rainbow No. 4, Tin Mountain, MacArthur, Tip Top, and Punch pegmatite mines were described by Page and others (1953). A much more detailed description of the Helen Beryl and Tin Mountain pegmatites is being prepared by M. H. Staatz and others. The Tip Top pegmatite was also described by Fisher (1942, 1945). The mines and claims described in the following pages are located by number on plate 21.

BIG TOM MINE (PEGMATITE 64)

The Big Tom mine, in the NE¼ sec. 12, T. 4 S., R. 3 E., is developed by 2 shallow shafts and a narrow opencut 150 feet long on a thin pegmatite. The past production has apparently been restricted to mica and beryl, but the amount is not known.

The pegmatite is tabular and strikes N. 50° W., parallel to the foliation of the quartz-biotite-sillimanite schist wallrock. The dip of the pegmatite, however, is about 80° SW., considerably steeper than that of the foliation in the country rock. The pegmatite is about 200 feet long and 15 feet thick, although it thins to a few feet at either end.

Five zones are distinguishable: a border and a wall zone of quartz-plagioclase-muscovite pegmatite, a first intermediate zone of cleavelandite-muscovite-quartz pegmatite, a second intermediate zone of cleavelandite-quartz, and a quartz core. The mineralogy and texture of these zones are given in table 7. The first intermediate zone is locally very rich in pale green beryl. Small areas of 20 to 30 square feet have as much as 2 percent beryl in crystals about 1½ inches in diameter. Beryl also occurs in the second intermediate zone. Cleavelandite of these zones characteristically forms rosettes as much as 1 foot in diameter.

Most of the exposed mica books are relatively large (3 to 6 inches across), but "A" structure and inclusions of other minerals are so prominent that little sheet mica can be recovered. Beryl crystals are small and difficult to separate from the rocks by hand. The percentage of beryl in a cross section of the pegmatite was estimated at 0.2 percent, on the basis of areal measurements of beryl crystals. The beryl-rich zones contain between 1 and 2 percent of beryl.

There is no evidence that the pegmatite body increases in size with depth. Future mining therefore will probably be on a small scale.

CORKY MINE (PEGMATITE 58)

The Corky is a typical small feldspar-beryl mine in the NE¼ sec. 12, T. 4 S., R. 3 E. The mine was operated intermittently during 1947–1957. It was owned in 1957 by Mr. Montgomery Heumphreus of Custer. A tape and compass map, made on December 17, 1950, is shown as figure 102.

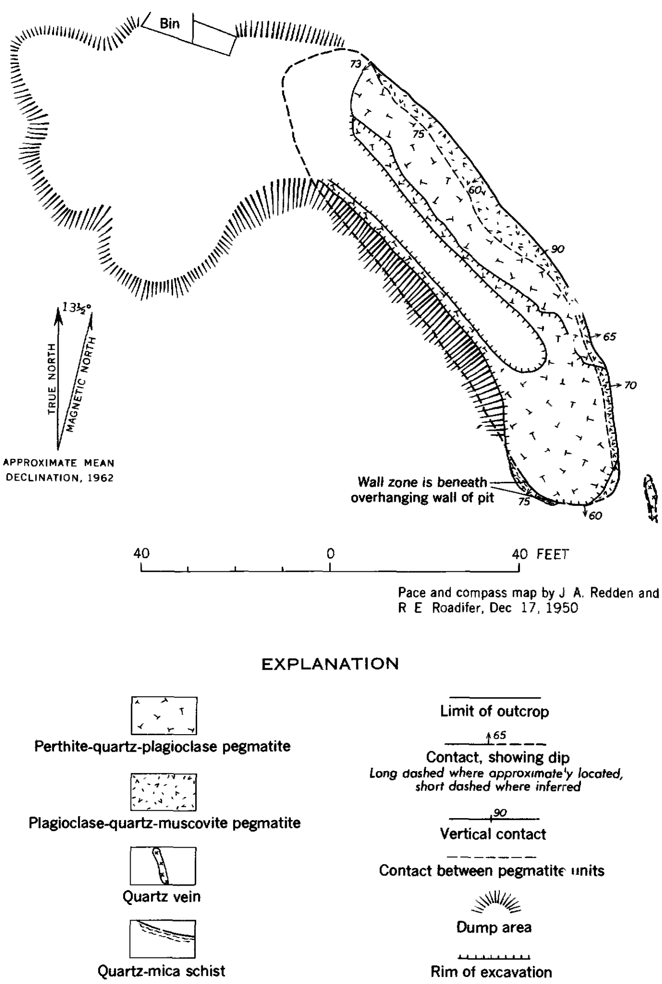

FIGURE 102.—Geologic map of the Corky pegmatite, Custer County, S. Dak.

The pegmatite is a tabular body striking about N. 40° W. and dipping 70° NE. to 70° SW. Its length is between 110 and 120 feet. The maximum thickness is about 28 feet, and the average is about 25 feet. Only the upper part of the pegmatite is exposed.

Two zones are distinguishable: a wall zone of plagioclase-quartz-muscovite and an apparent core of quartz-perthite-plagioclase. The wall zone ranges from about 4 inches to 3 feet in thickness; it averages 1.5 feet. The inner zone ranges from 12 feet to 24 feet in thickness. The mineral content and textures of the two zones are presented in table 7.

Most of the muscovite is segregated on the inner side of the wall zone in crystals from 1 to 8 inches across. It is pale yellowish green to clear. Pale green beryl and lithiophilite-triphylite are associated with muscovite.

The core consists of perthite crystals, as much as 6 feet long, which are set in a matrix of massive quartz and fine-grained aggregates of quartz and plagioclase. Medium- to coarse-grained green beryl is associated with the massive quartz; fine-grained pale-green to colorless beryl occurs with the quartz-plagioclase aggregates. Rarely, white amblygonite crystals, 6 inches to 2 feet in diameter, are found with the perthite and massive quartz.

Nearly all the past production has been feldspar and beryl. The tonnage ratio of feldspar to beryl is about 30:1. However, the value of the beryl produced has been almost twice that of the feldspar.

The perthite is of good quality and the recovery has exceeded 30 percent of the mined rock. When the map (fig. 102) was made, the floor of the opencut was very rich in perthite. The outer contacts of part of the body diverge slightly downward, and it is reasonable to expect that the feldspar will continue to 2 or 3 times the present depth of the workings. The pegmatite is so small that the mine cannot become a major feldspar source.

The beryl is mostly medium to coarse grained and easily recoverable by hand sorting. The core and edge of the wall zone contains 0.9 to 1.0 percent beryl according to the past production. Probably a similar recovery can be expected at greater depths.

All the observed mica contains "A" structure, inclusions, ruling, or other defects and is of scrap quality. The recovered scrap mica amounts to about 2.5 to 3.0 percent of the mined rock, and this recovery can be expected to continue without change during further feldspar mining. Mr. Heumphreus believed (oral communication, 1952) that a small amount of the mica could have been sold as sheet mica. Sheet mica may be more abundant at considerably greater depth in the pegmatite where the wall zones may thicken at the expense of the core or may coalesce to form a single deposit.

Other minerals such as amblygonite and columbite-tantalite are so scarce that any sizable production is unlikely.

GAYLE (DUBUQUE) AND ROYAL FLUSH CLAIMS (PEGMATITE 33)

The Gayle-Royal Flush pegmatite is on 2 claims in the SW¼ sec. 31, T. 3 S., R. 4 E. The Gayle claim, formerly known as the Dubuque claim, was owned in 1957 by Mr. G. Behrens of Custer. The Royal Flush is a patented claim adjoining the Gayle claim to the southeast; it was owned in 1957 by Mr. Joe Monger of Custer. These properties were described by Page and others (1953, p. 100–101) under the names Dubuque and Royal-Flush.

The mine workings in the northwest part of the pegmatite (fig. 103) on the Gayle claim consist of a 15-foot shaft and several small pits. On the Royal Flush claim, southeast of the map area of figure 103, an opencut nearly 80 feet long and as much as 12 feet deep exposes the southeast part of the pegmatite. There are other small prospect pits along the pegmatite between the main workings.

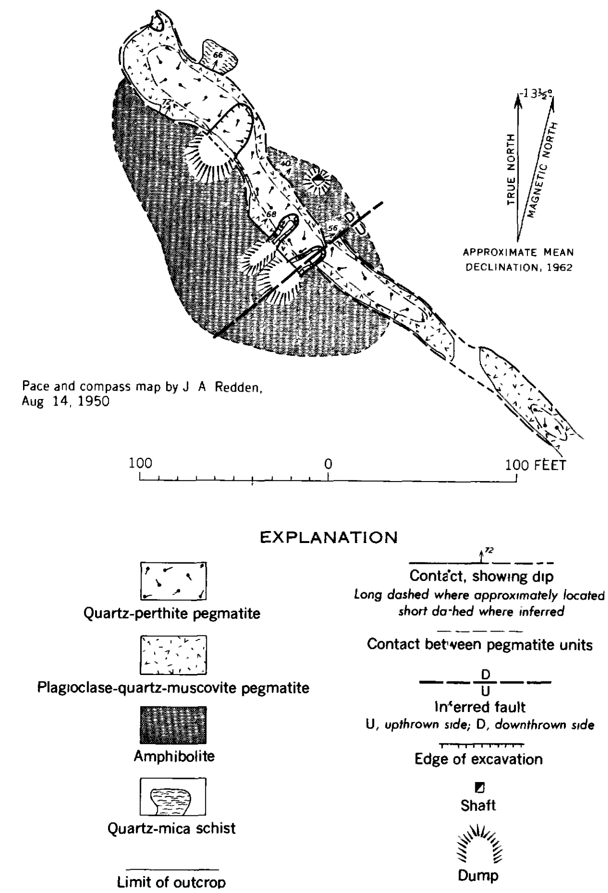

FIGURE 103.—Geologic map of part of the Gayle-Royal Flush pegmatite, Custer County, S. Dak.

Outcrops are not continuous, but apparently the exposed pegmatite is part of a single body nearly 1,000 feet long. The pegmatite is thin and dikelike; it strikes N. 40° W. and dips 35° to 75° to the northeast. The country rock is predominantly quartz-mica schist, but two bodies of amphibolite are at each end of the pegmatite. The middle of the pegmatite is only about 5 feet thick, but at the ends it is as much as 28 feet thick. Only the thicker parts are mined.

Two zones were mapped—a wall zone of plagioclase-quartz-muscovite and a core of quartz-perthite pegmatite. A border zone about 1 inch thick surrounds the wall zone which is 1 to 6 feet thick. If mapping were on a larger scale, the inner zone of the northwest part of the pegmatite could be subdivided into a quartz core surrounded by quartz-perthite. The core is discontinuous and is found only in the thicker parts of the pegmatite.

The mineralogy and texture of the Royal Flush and Gayle parts of the pegmatite are summarized in table 7. Most of the exposures of the border zone are very rich in quartz and muscovite, although plagioclase is locally abundant where the wallrock is amphibolite. The wall zone is richer in muscovite at the southeast end of the pegmatite than at the northwest end. The common accessory minerals in the different zones are tourmaline, apatite, and beryl. Some of the tourmaline in the wall zone is completely replaced by fine-grained muscovite.

The core of the southeast part of the pegmatite is richer in perthite and poorer in quartz than that of the northwest part. The outer core of the southeast part contains pale-green beryl, muscovite, tourmaline, and small crystals of amblygonite.

Beryl has been the largest source of income from the relatively small production of the Gayle-Royal Flush pegmatite. Some perthite and scrap mica has been produced, largely from the southeast opencut, but the relatively small size of the exposed pegmatite prohibits any large future production.

Much of the more recent mining has been for beryl and scrap mica. Mr. Behrens (oral communication, 1950) recovered 900 pounds of beryl in two crystals from the larger opencut (fig. 103). The previous owner is reported by Mr. Behrens to have recovered 1,500 pounds of beryl from an aggregate of several crystals that were exposed on the original outcrop. The percentage of beryl produced from this part of the pegmatite is high relative to the small amount of mined rock, but it is doubtful that the beryl is uniformly distributed and that the true grade is as high as is indicated by the production. About 0.6 percent of the mined rock from the southeast end of the pegmatite was recovered as beryl. This figure is probably close to the true grade of the beryl in the pegmatite.

All the observed mica has "A" structure or contains other defects so that it is of scrap quality. The scrap mica content of the pegmatite is relatively low, and such mica will be recovered only as a byproduct in the mining of feldspar or beryl.

The contacts diverge downward near the northwest end of the pegmatite, so the thickness is somewhat greater at depth than at the surface. It is not likely, however, that the increase in thickness continues to any great depth. On the southeast end, the contacts converge downward; hence the pegmatite becomes thinner at depth.

LINCOLN NO. 7 CLAIM (PEGMATITE 81)

The Lincoln No. 7 claim in the SE¼ sec. 5, T. 4 S., R. 4 E. (pl. 21, E–2) was owned in 1957 by the Consolidated Feldspar Department of the International Minerals and Chemical Corp. Several pegmatites are exposed on the claim, but only one is explored and that only by a small test pit. Apparently feldspar is the only mineral that has been recovered.

The pegmatite that has been explored is about 60 feet long and not more than 10 feet thick. It strikes N. 50° W. parallel to the strike of the country rock and dips 75° SW. The dip is somewhat steeper than that of the enclosing quartz-biotite schist. Exposures are poor, but the thickness probably does not exceed 10 feet and the length is less than 100 feet.

The zones are a border zone of quartz and muscovite, a wall zone of quartz, muscovite, and plagioclase, and an inner zone or core of perthite and quartz. The average thicknesses of these zones are 1 inch, 1 foot, and 6 feet, respectively. The mineralogy and textures are given in table 7.

Perthite is coarse grained and contains a few inclusions of tourmaline. The only beryl observed consisted of 2 yellowish-green crystals, approximately 0.5 inch in diameter, which were found in the wall zone. All the mica is of scrap quality. There is no evidence that the pegmatite increases in size at depth, nor is there evidence that its content of perthite, beryl, or mica will increase.

LITHIA LODE CLAIM (PEGMATITE 9)

The Lithia Lode claim is due north of the Warren Draw mine in the NW¼ sec. 36, T. 3 S., R. 3 E. (pl. 21, C–1). It was owned by the Maywood Chemical Corp. of Trenton, N.J., in 1955. Several small prospects expose a long thin pegmatite body. No production is known, but in 1950 there was a stockpile of approximately 40 pounds of amblygonite at one of the prospect pits.

The pegmatite body strikes northwest and dips 15° to 30° northeast. Its average thickness is 3 feet and the maximum is 7 feet. Most of the wallrock is amphibolite.

The border zone is of quartz, muscovite, and plagioclase, and the wall zone, which is 1 foot thick, consists of plagioclase, quartz, and muscovite. Wall zone rock contains approximately 0.5 percent yellowish-white beryl. The innermost zone of the pegmatite consists of these same minerals plus 6 to 10 percent perthite (table 7) and some associated amblygonite. Discontinuous centrally located pods of quartz are exposed in some of the prospect pits.

The amount of beryl, amblygonite, and perthite in the thin wall zone is small, and the limited size of the pegmatite precludes extensive mining.

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RED SPAR MINE (PEGMATITE 70)

The Red Spar mine in the NE¼ sec. 23, T. 4 S., R. 3 E., (pl. 21, B–5) is a source of feldspar, beryl, and scrap mica. It was owned in 1955 by the Consolidated Feldspar Department of the International Minerals and Chemical Corp. but was leased by various individuals in later years. Dale McDermand of Custer worked the mine intermittently in 1951; L. D. Pitts of Custer mined there for several weeks in 1952. The workings in August 1951 consisted of a small 2–level opencut (fig. 104) near the center of a tabular pegmatite. Most of the past production was feldspar, but about 10 tons of beryl and 20 tons of scrap mica also were produced.

The pegmatite body is conformable with the well-bedded quartz-biotite schist country rock, which locally contains sillimanite and staurolite. The hanging wall is exposed as a dip slope that strikes about N. 75° W. and dips 40° to 50° to the south (fig. 104). The attitude of the footwall is more irregular and curves around the thick central part of the body. Minor rolls on both the hanging wall and footwall plunge 39° to 50° S. 3° E. to S. 15° W. These plunges are similar in orientation to the regional plunges of folds in the country rock, and are assumed to be approximately parallel to the plunge of the pegmatite. The dips of the hanging wall and footwall indicate that the pegmatite thins at depth as shown in the section in figure 104.

Three units have been distinguished in the pegmatite: a quartz-plagioclase border zone, a quartz-plagioclase-muscovite wall zone, and a quartz-perthite core. The estimated percentages and sizes of the various minerals are given in table 7. The border zone is only a few inches thick despite its wide exposure as a dip slope. The wall zone is a few inches to 1 foot thick in the upper part (north end) of the pegmatite but thickens to 4 feet or more at depth. The core is approximately 20 feet thick at the elevation of the opencut.

The perthite has a deep reddish-salmon color, which has given rise to the name of the mine. Most of the perthite is in the inner quartz-rich zone, but it also forms as much as 10 percent of the wall zone. The feldspar is exceptionally free from impurities, but is limited in quantity and future production will be small.

The beryl-bearing unit, 1 to 4 feet thick, includes the inner part of the wall zone and the outer part of the core, where tourmaline, muscovite, and lithiophilite-triphylite are associated minerals. The beryl crystals range from 1 inch to 10 inches in diameter. One "mass" of beryl was reported to have weighed 1,500 pounds. The estimated grade of the beryl-bearing rock is 0.5 to 1.0 percent.

All the mica produced is of scrap quality and it is doubtful that sheet mica will be found. Scrap mica from the inner part of the wall zone and from the core is relatively coarse grained and easily cobbed.

A few pounds of columbite-tantalite has been produced and a similar amount may be expected in future mining.

ROCK RIDGE MINE (PEGMATITE 113)

The Rock Ridge mine, in the SE¼ sec. 4, T. 4 S., R. 4 E., has been primarily a potash feldspar producer. It was owned in 1957 by the Consolidated Feldspar Department of the International Minerals and Chemical Corp. Feldspar was mined in the 1930's but only one man was working there intermittently in 1951 when a tape and compass map was prepared (fig. 105). A large opencut with a maximum depth of 30 feet exposes the pegmatite.

Past production is not known, but the relatively large opencut and small size of the dump suggest that the production was considerable.

The exposed pegmatite forms a long tabular discordant body that strikes somewhat west of north. Exposures of the footwall contact are rare, but the body probably thins somewhat with depth. The plunge of the body is assumed to be parallel to that of neighboring pegmatites, which is about 40° to the southeast.

The pegmatite has two main zones: a quartz-plagioclase wall zone and a quartz-perthite core (fig. 105). A thin border zone of quartz and plagioclase is included with the wall zone. In most places the wall zone is about 8 feet thick, but in a few places it is as little as 2 feet thick.

The mineralogy and some textural data of the pegmatite are listed in table 7. Plagioclase in the wall zone is very coarse grained; single crystals of grayish-white plagioclase are more than 1 foot in diameter.

The volume of the present mine dumps as compared with the volume of the opencut indicates that at least 50 percent of the core was perthite. Single perthite crystals as much as 8 feet in diameter are still exposed in the end of the opencut, and probably large crystals are concealed beneath the floor of the opencut. Some of the perthite in the outer part of the pegmatite has graphic quartz but this is rare elsewhere. The contacts of the body converge downward indicating that the feldspar-rich zone thins at depth.

Beryl is found only in the outer 2 feet of the core, and even there it forms not more than 0.3 percent of the rock. The largest crystal observed had a diameter of approximately 10 inches. Some beryl could be recovered as a byproduct of feldspar mining.

The only rich concentration of muscovite occurs in a unit about 1.5 feet thick along the inner part of the wall zone. All the mica in the core is pale green, and occurs as wedge-shaped books as much as a foot in

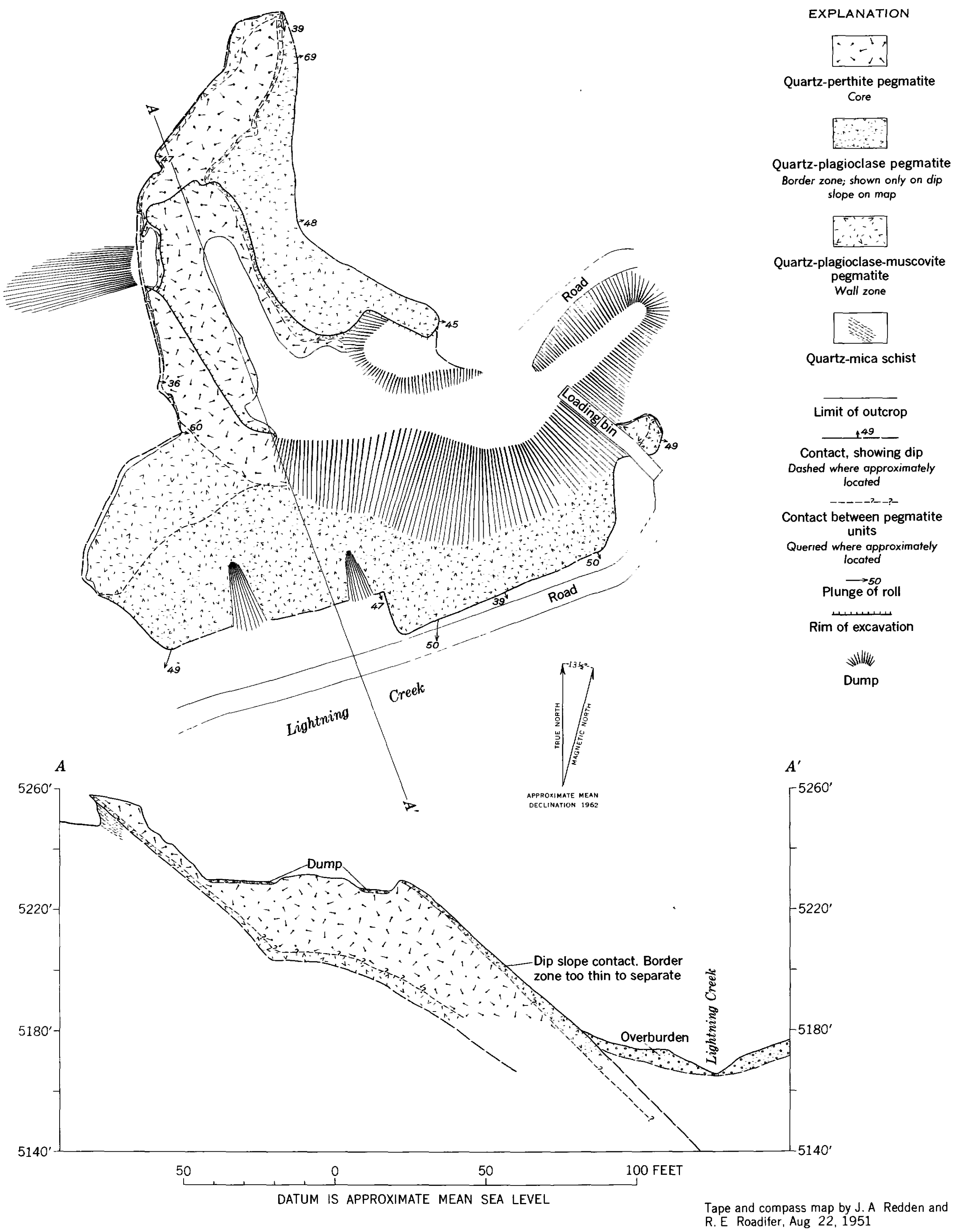

FIGURE 104.—Geologic map and section, Red Spar pegmatite, Custer County, S. Dak.

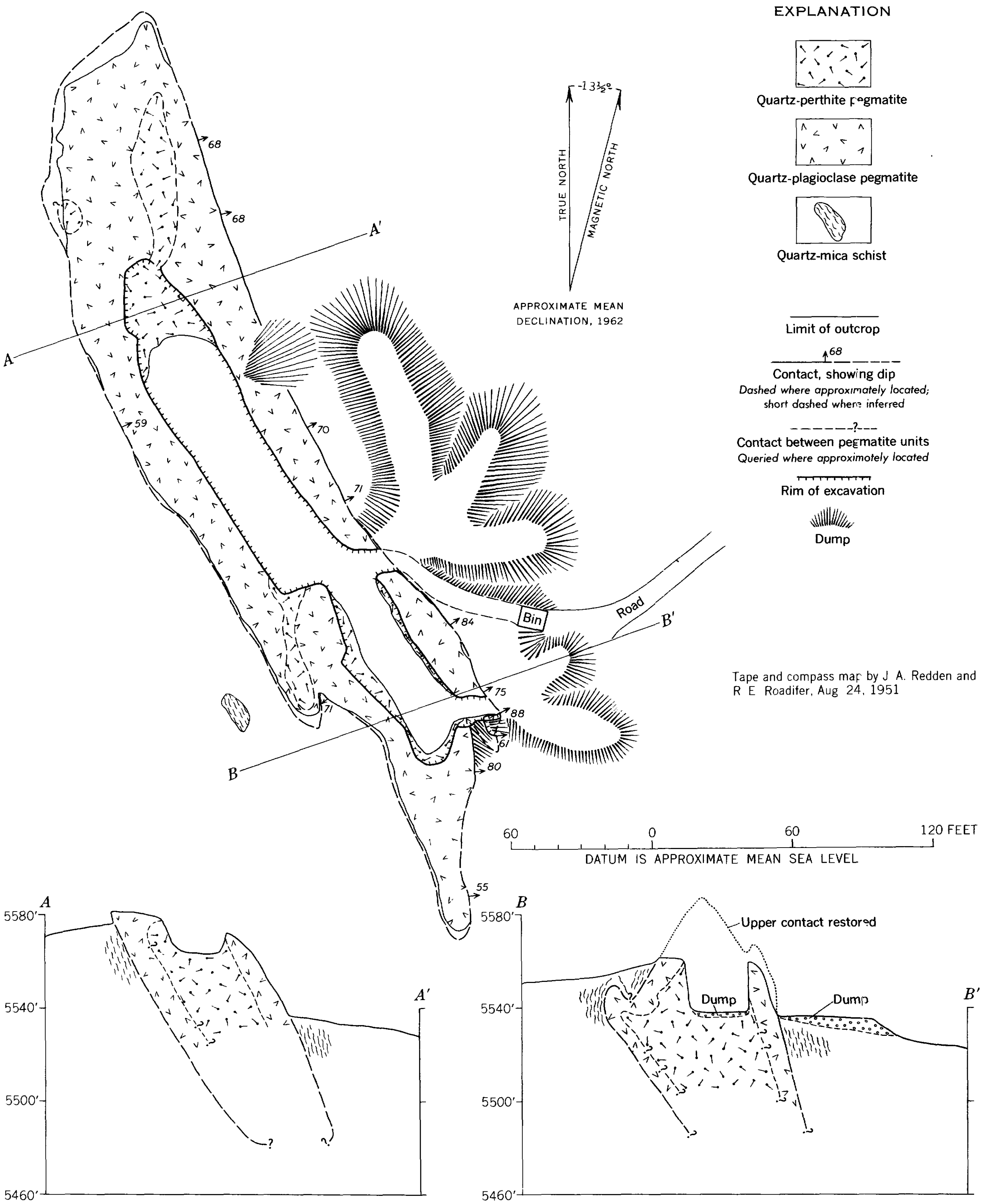

Figure 105.—Geologic map and sections, Rock Ridge pegmatite, Custer County, S. Dak.

length. The observed mica is of scrap quality and small in quantity. It can be recovered only as a byproduct of feldspar mining. No sheet mica is known to have been mined, although coarse-grained plagioclase like that in the wall zone commonly is an indication of rich sheet mica-bearing pegmatite (Page and others, 1953, p. 44).

A few altered crystals of spodumene, 2 to 10 inches long, were found in the bottom of the opencut. Fresh spodumene was not found, but some may exist at depth.

WARREN DRAW MINE (PEGMATITE 12)

The Warren Draw mine in the SE1/4 sec. 35 and the SW1/4 sec. 36, T. 3 S., R. 3 E., is one of the larger pegmatite mines in the Fourmile quadrangle. It was owned in 1957 by Mrs. Gladys Wells and Kem Koch of Custer; they have operated the mine intermittently since the late 1930's. The workings consist of 4 opencuts and 10 small prospect pits (pl. 23). A small mill on the property had been used for the recovery of scrap mica. The main product has been perthite and the byproducts are beryl and scrap mica.

GEOLOGY

The pegmatite crops out on a northwest-trending hill (pl. 23), the sides of which are covered with numerous blocks of pegmatite. Most of the exposed country rock is quartz-mica schist, but there are a few small bodies of amphibolite in contact with the pegmatite.

The shape of the pegmatite body is somewhat irregular. The central part of the body has a flat-lying discordant hanging wall that is approximately parallel to the surface of the ground, as shown in the sections (pl. 23). Two limbs, one trending eastward and the other north-northwest, extend outward from the central mass of pegmatite. The average dip of the main body is between 15° and 20° to the east, but the east-trending limb dips steeply northward and the north-northwesterly limb probably dips steeply eastward. The overall plunge is to the east, but rolls in schist inliers on the hanging wall of the central part of the body plunge gently northwest and southeast.

There are three distinguishable zones: (a) a very fine grained border zone of quartz-plagioclase-muscovite; (b) a wall zone containing two mappable subdivisions, one of which is quartz-plagioclase-muscovite in both layered and unlayered assemblages, and the other, sugary-grained plagioclase-quartz-muscovite; and (c) a perthite-quartz core. Thin fracture fillings cut all zones but are not shown on the map. North of the opencut that is near the mill (pl. 23), the outcrop is composed of both massive and fine-grained quartz with accessory muscovite. This outcrop may be part of a quartz-rich core, but was mapped as a phase of the wall zone.

QUARTZ-PLAGIOCLASE-MUSCOVITE PEGMATITE (BORDER ZONE)

The very fine grained border zone is about 2 inches thick and was mapped with the wall zone. Its composition varies; in a few places it has abundant plagioclase and in other places it consists only of quartz and muscovite. The accessory minerals are tourmaline, apatite, garnet, and beryl. The average grain size is 0.1 inch.

QUARTZ-PLAGIOCLASE-MUSCOVITE AND PLAGIOCLASE-QUARTZ-MUSCOVITE PEGMATITE (WALL ZONE)

The wall zone forms a 0.5- to 10-foot thick shell around the pegmatite. It has been separated into two subdivisions: one consists of very fine grained sugary-textured plagioclase-quartz-muscovite; the other is fine-grained quartz-plagioclase-muscovite. The latter phase locally contains line rock structure near the south contact of the pegmatite.

The fine-grained quartz-plagioclase-muscovite part of the wall zone is widely exposed. It consists of about 40 percent quartz, 40 percent plagioclase, 10 percent muscovite, 4 to 10 percent perthite, 2 percent tourmaline, and accessory apatite, beryl, and garnet. The quartz and plagioclase are about 0.5 inch in diameter. The muscovite occurs in plates averaging 1 inch in diameter, but locally the plates are as much as 8 inches across. Muscovite is more abundant along the inner side of this unit. Perthite is locally abundant as crystals ranging in size from 2 inches to 1 foot. Green euhedral beryl crystals, as much as 4 inches in diameter, are irregularly distributed along the inner part of the zone, and fine-grained greenish-white anhedral crystals are disseminated in the outer part of the zone. Tourmaline occurs throughout the zone in elongate crystals oriented sub-perpendicular to the outer contact.

On the south side of the pegmatite there are scattered exposures of line rock in this zone. The darker layers of line rock are rich in either tourmaline or muscovite; the light-colored layers are rich in plagioclase. The layers range from 0.1 inch to 10 inches thick; most of the individual minerals are less than 0.1 inch in diameter. This line rock is best exposed just north of the mill shed and west of the largest opencut. A few of the tourmaline- and muscovite-rich layers are crinkled and folded, but there may be virtually unfolded layers on either side (fig. 87). Small crystals of apatite and garnet are commonly less than 0.1 inch in diameter.

The very fine grained plagioclase-quartz-muscovite phase of the wall zone is exposed only in the northwestern limb of the pegmatite. The rock contains about 60 percent plagioclase, 30 percent quartz, 10 percent muscovite, and accessory garnet, apatite, amblygonite, and spodumene(?). The thickness is at least 4 feet, and it may be much greater. The average grain size is about 1 mm, but scattered through the rock are coarser grained stringers and pods as much as 3 feet thick composed mainly of fine-grained quartz, perthite, and muscovite, but also containing small beryl crystals. The spodumene(?) occurs in the very fine grained pegmatite in rounded masses averaging about 2 inches in diameter. It has been almost entirely altered, and now consists of a soft maroon powder that surrounds a center of white powder containing fragments of spodumene. In places the powder has weathered out, leaving cavities. The contact between this rock and the rest of the wall zone rock is gradational. The geologic relations are not entirely clear, but the trace of lithium minerals present suggests the very fine grained plagioclase-quartz-muscovite phase may be younger than the wall zone elsewhere in the pegmatite.

PERTHITE-QUARTZ-PLAGIOCLASE PEGMATITE (CORE)

The innermost exposed zone consists of perthite (50 percent), quartz (27 percent), plagioclase (20 percent), muscovite (2 percent), tourmaline, and beryl. Although perthite averages 50 percent, it ranges from 30 to 90 percent.

In the largest opencut the core contains layered structures that are not found elsewhere in the Fourmile quadrangle. Layers consisting almost entirely of perthite are parallel to the upper contact and to fracture fillings. The perthite crystals in these layers are lath shaped, are a foot or more long and 2 to 4 inches thick, and are oriented approximately perpendicular to the layering. Small segregations of fracture fillings of quartz and muscovite a fraction of an inch thick separate the nearly pure layers of perthite. Some of the perthite crystals contain thin layers of tiny inclusions of hematite, 0.5 inch to 4 inches thick, that are parallel to the thicker layers. Shears and joints are parallel to the hematite layers.

Elsewhere in this zone, the layers of perthite are absent and the textures are comparable to that of other pegmatite. Perthite occurs as flesh-colored subhedral to euhedral crystals as much as 4 feet in diameter. The matrix between the large perthite crystals consists of quartz grains an inch or less in diameter and a smaller amount of plagioclase. Muscovite occurs largely as small intergrown heterogeneously oriented plates in aggregates (called "bull" mica by the miners) that form rounded or tabular-shaped bodies, 1 to 4 feet thick, composed of about 60 percent muscovite and 40 percent quartz. Scattered wedge-shaped plates of muscovite are also mixed with quartz. Pale green beryl occurs in the outer part of the perthite-quartz-plagioclase zone adjacent to the beryl-bearing part of the wall zone. Crystals are as large as 8 inches in cross section.

QUARTZ-SPODUMENE-AMBLYGONITE-MUSCOVITE FRACTURE FILLINGS

Quartz-spodumene-amblygonite-muscovite fracture fillings, generally parallel to the layers in the main opencut, range from 2 inches to nearly 20 inches in thickness (fig. 83). Locally, spodumene is completely altered to a soft claylike substance, and amblygonite is coated with a soft white powder. Accessory beryl is associated with muscovite along the edges of the fracture fillings. These fracture fillings cut the core and wall zone and locally extend to the pegmatite contact. Joints and slip surfaces within the fracture fillings are parallel to their contacts. One fracture filling follows a small fault that displaces the outer contact about 6 inches. The fracture filling does not penetrate the country rock however.

MINERAL DEPOSITS

FELDSPAR

Perthite feldspar is the major product of the Warren Draw mine. The greatest production came from the core, which contains an estimated 70 percent perthite in the vicinity of the main pit. Elsewhere the perthite content of the core is about 40 percent. Some of the feldspar contains small hematite inclusions and is intergrown with quartz. These impurities are so abundant that some beneficiation or upgrading may be necessary in future mining. Reserves of potash feldspar are inferred to be at least 10 times the past production.

MICA

Almost all the mica mined from the pegmatite is of scrap quality. Mrs. Gladys Wells (oral communication, 1952) stated that "very little" sheet mica has been produced. Large mica books, as much as 1 foot long, were found in the core in all four opencuts, but all these large books were of scrap quality. Much of the mica is in books 1 inch or smaller, within "bull" mica aggregates in the innermost zone. Reserves of scrap mica are large.

BERYL

Most of the observed beryl occurs in a unit about 2 feet thick along the contact between the wall zone and the core. Measurements of beryl in typical exposures of this unit indicate a content of 0.4 percent beryl. Most of the crystals are less than 2 inches in diameter, and only about half of them could be recovered by cobbing.

A small amount of beryl is disseminated throughout the core in the large opencut. Most of the beryl is in

2-inch crystals; the largest was 8 inches in diameter. Small beryl crystals in the wall zone and in the border zone are less than 1 inch in diameter and would be difficult to recover. Approximately 400 pounds of beryl was stockpiled in 1952. The content of beryl through the entire pegmatite body is apparently less than 0.1 percent.

PEGMATITES 10 AND 11 (WARREN DRAW CLAIM)

Pegmatites 10 and 11 (pl. 21) are on the Warren Draw claim. Pegmatite 10 is about 200 feet northwest and pegmatite 11 is about 200 feet west of the Warren Draw pegmatite.

Pegmatite 10 consists of 2 separate bodies, one vertically above the other, in amphibolite wallrock. The upper body is developed by an adit, 60 feet long, that trends northeast and exposes a cross section of the pegmatite. The lower pegmatite was worked from 3 inclined shafts, 25 to 30 feet deep, the bottoms of which are joined in a stoped area. Scrap mica and probably some sheet mica have been produced from this pegmatite.

The upper body strikes northwest and dips 60° to the northeast at the mouth of the adit. A few feet to the northeast, the dip becomes horizontal and continues so to the end of the drift. The thickness is about 6 feet. The lower pegmatite strikes about N. 60° W., and dips 50° NE. It averages 2.5 feet thick and does not exceed 4 feet. The exposed length is 60 feet.

Each of these bodies has a quartz core, 1 to 5 feet thick, bordered by a wall zone and a border zone of plagioclase-quartz-muscovite. The quartz core in the lower, thinner body is exposed only in thick parts of the pegmatite; elsewhere the wall zone extends across the entire width of the pegmatite. The amount of muscovite in the wall zone is variable; locally it forms as much as 35 percent of the rock. The muscovite books are clear to pale yellow and most are small—less than 4 inches in diameter. Perthite and tourmaline were not found in either pegmatite.

Pegmatite 11 is bordered by quartz-biotite schist and amphibolite. Several small prospect pits extending entirely across the pegmatite have yielded a little scrap mica. The pegmatite strikes northwest for 120 feet and dips 15° to 30° northeast. Its maximum thickness is 3 feet. Two zones are distinguishable: a cleavelandite-muscovite-quartz wall zone and an apparent core of massive cleavelandite (table 7). The cleavelandite core contains vugs that are lined with hemispherical aggregates of cleavelandite as much as 6 inches in diameter. Tiny white crystals of apatite are perched on the cleavelandite crystals.

Possibly the apparent core is an intermediate zone and a true core is present in thicker, concealed parts of the body.

Pegmatite 10 is too small for extensive economic development. Scrap mica is fairly abundant, but scarcely any sheet mica could be recovered if the exposed mica is representative. It would be necessary to use more expensive underground-mining methods than would be warranted in the recovery of scrap mica alone.

Pegmatite 11 is so small that it is of no importance for mining. Scrap mica is the only mineral that might be recovered.

WRIGHT MICA LODE (PEGMATITE 22)

The Wright Mica Lode in the NE¼ sec. 36, T. 3 S., R. 3 E. is primarily a feldspar and beryl mine. In 1956 it was owned by the heirs of the late W. W. Wright of Custer. The mine workings consist of an opencut 8 to 10 feet deep and a small prospect trench (fig. 106). The mine was inactive in the fall of 1950, and probably had been inactive since World War II.

Two zoned pegmatites separated by a thin septum of schist are exposed on the property, as well as several bodies of homogeneous pegmatite. The southern zoned pegmatite is pipelike and plunges 35° to 40° S. 10° to 15° E. The narrow part of the opencut (fig. 106) follows the crest of the pegmatite down plunge along the south slope of a gentle hill. The northern zoned pegmatite is less well exposed but probably is similarly shaped and may join the southern pegmatite at depth. The plunge of both bodies is parallel to the plunge of folds in the country rock, but in detail both bodies are discordant with the quartz-mica schist country rock. In the opencut the schist adjacent to the contact with pegmatite is tourmalinized and contains porphyroblasts of muscovite. Muscovite porphyroblasts are also conspicuous in a small schist roll on the crest of the pegmatite in the narrow southern part of the opencut.

The southern zoned pegmatite contains a quartz-muscovite border zone, a plagioclase-quartz-muscovite wall zone, a perthite-quartz intermediate zone, and a quartz core (table 7). The border zone is 0.5 to 2.5 inches thick. Though it consists mainly of quartz and muscovite, it has abundant plagioclase locally. The wall zone ranges in thickness from 0.5 to 2.5 feet. Some exposures of the wall zone contain as much as 40 percent perthite, but in others it is entirely absent. Small 1-inch crystals of pale-green to yellow beryl are disseminated throughout the zone. Other accessory minerals are biotite and apatite. The quartz and plagioclase are in anhedral grains; muscovite forms small

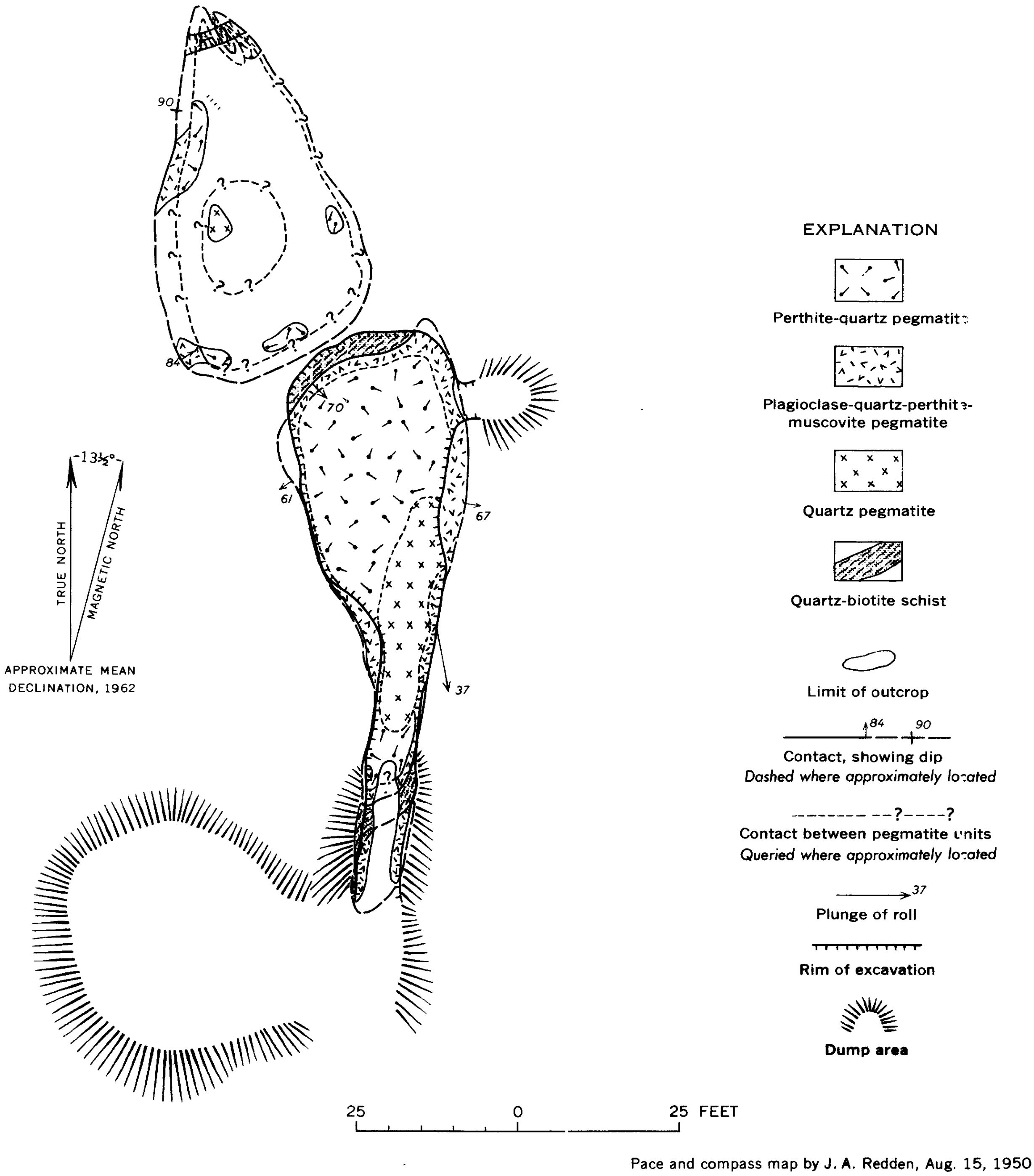

FIGURE 106.—Geologic map of the Wright Mica Lode, Custer County, S. Dak.

diamond-shaped books, and perthite occurs as subhedral flesh- to cream-colored crystals.

At the south end of the opencut, a transitional phase between the wall zone and the intermediate zone contains about equal proportions of plagioclase, quartz, and perthite, and 8 percent muscovite.

The intermediate zone is exposed in the wide part of the opencut where it forms a hood over the quartz core. Above the core the zone is about 5 feet thick, but along the side it commonly is only 2 feet thick and locally is absent. The zone is composed largely of perthite and quartz, but plagioclase and muscovite, are accessory minerals. Beryl crystals as much as 18 inches in diameter are found near the outer part of the zone. Most of the quartz is massive, and the perthite is in subhedral crystals as much as 4 feet long.

The quartz core exposed in the bottom of the pit is pod-shaped. Its maximum thickness is 8 feet, and its length is about 12 feet. The zones of the northern pegmatite are very poorly exposed, but apparently have the same composition as those of the southern one.

The wall zone of the southern pegmatite contains 5 to 12 percent scrap mica in books averaging 1 inch but as large as 4 inches in diameter. All the mica is ruled, weeved, or has "A" structure. About two-thirds can be hand cobbed as scrap mica.

Beryl occurs in the wall and intermediate zones of the southern pegmatite and in the wall zone of the northern one. Practically all the beryl has been removed from the opencut, but casts of beryl crystals as much as 18 inches in cross section are visible. The volume of the casts indicates that at least 500 pounds of beryl has been removed and that the mined rock contained at least 0.2 percent beryl. Probably 90 percent of the beryl in the intermediate zone is recoverable by hand cobbing. The beryl in the wall zone is finer grained and probably only 50 percent is recoverable.

About 75 percent of the perthite in the intermediate zone is recoverable by cobbing. The zone is apparently hood shaped, however, and much of the thicker part has already been mined.

VOLCANO MINE (PEGMATITE 53)

The Volcano mine is in the SE¼ sec. 1, T. 4 S., R. 3 E. and has produced mica, feldspar, and beryl from a thin tabular narrow pegmatite body. A narrow opencut, 100 feet long and locally as much as 15 feet deep, and 4 small prospect pits are the only workings.

The pegmatite strikes northwest parallel to the strike of bedding and foliation in the quartz-mica schist country rock. The dip of the contacts varies considerably, but the average dip is about 80° SW., which is steeper than the dip of the enclosing schists. The plunge is steep to the south. The pegmatite outcrop is approximately 220 feet long and has an average width of about 15 feet. The pegmatite is 25 feet thick at the northwest end and tapers to about 3 feet at the southeast end.

A thin border zone, consisting of quartz, plagioclase, and muscovite, and a fine-grained wall zone, 1 to 8 feet thick, of quartz and plagioclase, enclose the entire pegmatite body. The composition of the wall zone is given in table 7. At the southeastern end where the thickness is only about 3 feet, the wall zone consists of almost equal amounts of quartz, plagioclase, and tourmaline; the lack of muscovite is conspicuous.

A coarse-grained core of quartz, cleavelandite, and perthite is exposed in the thicker northwest end of the pegmatite. Muscovite, beryl, amblygonite, and lithiophilite-triphylite are accessory minerals. The beryl and the amblygonite contents of the core is estimated at 0.5 percent each.

Most of the visible mica is of scrap quality and the grade is low. Sheet mica may have been obtained from the thin southeastern part of the opencut, which is now filled with waste rock. Some potash feldspar could be produced from the core, but the relatively small total amount of feldspar and the small size of the pegmatite preclude any large production. The northwest end of the pegmatite thickens somewhat with depth and possibly this part of the pegmatite could be mined for feldspar, mica, and beryl.

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