

GEOLOGY AND MINERAL DEPOSITS  
OF THE ELLISTON MINING DISTRICT  
POWELL COUNTY, MONTANA

by

FORBES SMITH ROBERTSON

A thesis submitted in partial fulfillment of the requirements  
for the degree of

DOCTOR OF PHILOSOPHY

UNIVERSITY OF WASHINGTON

1956

Approved \_\_\_\_\_

Department \_\_\_\_\_

Date \_\_\_\_\_



## TABLE OF CONTENTS

	Page
LIST OF TABLES	x
LIST OF ILLUSTRATIONS	xi
ABSTRACT	xvii
INTRODUCTION	1
Purpose and Scope	1
History and Production	2
Previous Work	3
Acknowledgements	3
GEOGRAPHY	6
Location and Definition of District	6
Accessibility	6
Vegetation and Climate	7
Topography	7
GENERAL GEOLOGIC SETTING OF THE BOULDER BATHOLITH	8
Geologic History in Elliston District	11
SEDIMENTARY ROCKS	12
Madison Group (Mississippian)	12
Big Snowy (?) (Mississippian)	13
Amsden Formation (Mississippian-Pennsylvanian ?)	15
Quadrant Formation (Pennsylvanian-Permian)	16
Phosphoria Formation (Permian)	17
Swift Formation (Jurassic	17
Morrison Formation (Jurassic ?-Lower Cretaceous ?)	18
Kootenai Formation (Lower Cretaceous	18
Colorado Group ? (Middle Cretaceous ?)	22



## TABLE OF CONTENTS (continued)

IGNEOUS ROCKS	23
<u>PRE-BATHOLITHIC VOLCANIC ROCKS</u>	24
Introductory Statement	24
Basic Andesitic Breccias	24
Areal Distribution	24
Petrographic Types	29
Origin of Breccias	31
Latitic and Andesitic Rocks	32
Lithologic Types and Distribution	32
Petrography of Units in Little Blackfoot Creek Section	32
Andesitic Lapilli Tuff	33
Andesitic Tuff Breccia	33
Andesitic Tuff	34
Petrography of Units in Cliff Mountain Section	37
Andesitic Tuff	38
Andesitic Tuff Breccia	40
Latite Tuff	42
Andesitic Breccia	42
Latite Tuffs and Breccias	43
Other Latite Localities	46
Welded Tuffs of Andesite Composition	46
Chemical Analyses of Andesite and Latite	49
Emery Basalt	51
Porphyritic Basalt with Pyroxene Phenocrysts	55



## TABLE OF CONTENTS (continued)

Porphyritic Basalt with Pyroxene and Plagioclase Phenocrysts	56
"Oatmeal" Basalt	57
Amygdaloidal Basalt	57
Cliff Mountain Basalts	59
Basalt Breccia	60
Chemical Composition of Basalts	61
Volcanic-Derived Sediments	62
Age and Correlation of Older Volcanics	63
Upper Cretaceous Volcanics in Adjoining Regions	68
Conclusions	71
YOUNGER VOLCANIC ROCKS	71
Rhyolite	71
Older Rhyolite	72
Younger Rhyolite	72
Previous Work on Welded Tuffs of Rhyolitic Composition	76
Chemical Composition	77
Age of the Rhyolites	80
<u>INTRUSIVE ROCKS</u>	83
Introductory <u>Stagement</u>	83
Granogabbro	84
Bison Mountain Granodiorite	87
Quartz Monzonite	93
Mineralogy	95
Olivine	95



## TABLE OF CONTENTS (continued)

Pyroxene	95
Hornblende	97
Biotite	97
Plagioclase	100
Potash Feldspars	106
Orthoclase	106
Microcline	110
Anorthoclase	110
Microperthite	120
Quartz	124
Accessory Minerals	125
Textures	126
Coarse-Grained Quartz Monzonite	127
Medium-Grained Quartz Monzonite	132
Fine-Grained Quartz Monzonite	138
Aplite and Alaskite	140
Definition	140
Structural Control of Aplite-Alaskite	141
Types of Aplite and Alaskite	143
Mineralogy	150
Potash Feldspars	150
Tourmaline	150
Sulfides	151
Alaskite Occurrences	152
Ontario Mine Area	152
Telegraph Mountain Area	154



## TABLE OF CONTENTS (continued)

Magmatic Phenomena Leading to Metalization	156
Magmatic Differentiation	158
Deuteric Crystallization Stage	164
Introductory Statement	164
Definition	165
Deuteric Stage Minerals and Textures	167
Origin of Aplite-Alaskite	167
Hydrothermal Stage	172
Definition	174
Structures in Pre-hydrothermal and Hydrothermal Stage	174
Crystallization Histories of Minerals	180
Potash Feldspar	182
Sericite	183
Quartz	183
Molybdenite	184
Tourmaline	185
Pyrite	186
Sphalerite-Galena	187
Copper Minerals	188
Carbonate Minerals	188
Vein Structures	188
Wall-Rock Alteration	192
Geochemical Evidence of Base Metals in Areas of Strong Wall-Rock Alteration	194

## TABLE OF CONTENTS (continued)

Distribution of Base Metal Deposits	202
Conclusions	202
Dikes and Sills	206
Dikes in Volcanic Pile	206
Dikes in Batholithic Rocks	207
Contact Metamorphism	208
Hornfelsic Contact Metamorphism	208
Microaplitic Contact Metamorphism	210
Thermal Stage of Metamorphism	215
Age of the Batholithic Rocks	216
Chemical Composition of Intrusive Rocks	216
STRUCTURE	224
Structural Relations of Contacts of the Batholith	225
Roof of the Batholith	225
Floor of the Batholith (?)	226
Faults	228
Northeast Fault System	228
East-West Fault	230
Northwest Faults	230
North-South Faults	231
Joints	231
Veins	232
Veins in Quartz Monzonite	232
Veins in Volcanic Rocks	233



## TABLE OF CONTENTS (Continued)

MINERAL DEPOSITS	234
Mineralogy	234
Non-Metallic Gangue Minerals	234
Tourmaline	234
Quartz	235
Calcite	238
Ankeritic Carbonates	238
Chalcedonic Quartz	238
Arsenopyrite	239
Pyrite	239
Sphalerite	241
Galena	247
Chalcopyrite	247
Bornite	247
Millerite	249
Hematite	249
Magnetic Iron Ore	249
Chalcocite	249
Covellite	249
Oxidation Minerals	250
Paragenesis	250
Stages of Vein Formation	250
Criteria for Paragenetic Relations	251
Paragenesis Diagram	251
Pyrite-Arsenopyrite	252
Sphalerite	255

## TABLE OF CONTENTS (continued)

Galena	255
Chalcopyrite	255
Veins	257
Veins in Quartz Monzonite	259
Quartz Veins	265
Variations in Thickness of Veins with Changes of Strike and Dip	265
Veins in Volcanic Rocks	265
N. 65° E. Veins	267
Northeast Veins	268
Flat Veins	271
Veins in Sedimentary Rocks	271
Breccia Deposits	272
Replacement in Limestone	273
Wall-Rock Alteration	273
Veins in Quartz Monzonite	275
Volcanic Rocks	278
Origin of Wall-Rock Alteration	278
OUTLINE OF PHYSIOGRAPHIC HISTORY	283
Upland Surface	283
Glaciation	284
Older Stage	284
Younger Stage	287



## TABLE OF CONTENTS (concluded)

PETROGENESIS	289
The Volcanic Pile	289
Sills in Sedimentary Rocks	290
Structure Prior to Emplacement of the Batholith	292
Emplacement of the Batholith	292
Marginal Relations of the Batholith	292
Magmatic Differentiation	294
Chemical Petrology	295
Method of Emplacement of the Batholith	299
Dimensions of the Boulder Batholith	301
Magmatic Stopping	301
Dilation	305
Granitization	307
Origin of the Regional Fracture Pattern	307
Mineralization	308
Summary	309
CONCLUSIONS	313
REFERENCES	318

## LIST OF TABLES

Table	Page
1. Stratigraphic Section of Sedimentary Rocks	14
2 Measured Section of Phosphoria Formation	19
3 Composite Section Upper Cretaceous Volcanics	25
4 Little Blackfoot Creek Section	26
5 Cliff Mountain Section	27
6 Chemical Analyses of Andesites	53
7 Chemical Analyses of Latites	54
8 Chemical Analyses of Basalts	66
9 Elkhorn Mountain Volcanics	70
10 Chemical Analyses of Rhyolite--Quartz Latite	78
11 Analyses of Geochemical Soil Samples at Comet Mine	196
12 Analyses of Geochemical Soil Samples Across Barren Zone between Comet and Gray Eagle Mines	198
13 Analyses of Geochemical Soil Samples Collected over Strongly Fractured and Altered Zone between Comet and Gray Eagle Mines	199
14 Chemical Analyses of Microaplite	214
15 Zirzon and Monazite Age Determinations	217
16 Chemical Analyses of Aplite	218
17 Chemical Analyses of Quartz Monzonite	220
18 Vein Minerals	236
19 X-ray "d" values for Tetrahedrite and Tennantite, Third Term Mine	245
20 Chemical Composition of Volcanic Pile	291
21 Chemical Composition of Sedimentary Rocks and Theoretical Composition of Granitic Magma	304



## LIST OF ILLUSTRATIONS

Figure	Page
1 Index Map	5
2 General view of Nigger Mountain taken from the air	9
3 Outcrop of basal andesitic pyroclastic breccia	30
4 Photomicrograph of microlitic to felted texture in groundmass of basal andesitic breccia at Cliff Mountain	30
5 Photomicrograph of andesitic tuff	35
6 Photomicrograph of altered pyroxene phenocryst	35
7 Photomicrograph showing partly collapsed shards	36
8 Photomicrograph of plagioclase phenocrysts in nearly glassy groundmass of andesitic tuff	36
9 Photomicrograph showing collapsed and compacted shards in slightly welded part of andesitic tuff	39
10 Photomicrograph of lapilli tuff in massive part of Unit No. 2, Cliff Mountain	39
11 Photomicrograph of andesitic tuff breccia	41
12 Photomicrograph of welded crystal tuff of latitic composition	41
13 Photomicrograph of lapillus fragment in crystal tuff	44
14 Outcrop of welded latitic tuff at Cliff Mountain	45
15 Lapilli tuff from upper part of latitic welded tuff	45
16 Photomicrograph of compaction structures in latitic welded tuff. Note direction sense of movement.	50
17 Photomicrograph of compaction structures in latitic welded tuff. Note small drag folds.	50
18 Photomicrograph of plagioclase phenocryst in "Oatmeal" basalt	58
19 Photomicrograph of texture of leucocratic granitic inclusion in basalt	64

## LIST OF ILLUSTRATIONS (continued)

Figure		Page
20	Photomicrograph of cataclastic texture of inclusion in basalt	64
21	Photomicrograph of plagioclase phenocryst in devitrified rhyolite	73
22	Photomicrograph of partly resorbed sanidine phenocryst in obsidian vitrophyre	73
23	Photomicrograph of perlitic fractures in glass	75
24	Photomicrograph of welded and microbrecciated glass at base of welded rhyolite tuff	75
25	Photomicrograph of potash feldspar (anorthoclase ?) in granogabbro	86
26	Photomicrograph of diabasic texture in granogabbro	86
27	Photomicrograph of hornblende crystal with relic of pyroxene. Note dark patches of leucoxene	90
28	Photomicrograph of texture of Bison Mountain granodiorite showing flood of late quartz	90
29	Photomicrograph of plagioclase phenocryst with alteration to more albitic composition at margin of crystal and along fracture through the crystal	92
30	Photomicrograph of olivine in basic quartz monzonite at base of sill-like body in Pauper area	96
31	Photomicrograph of relic olivine in altered pyroxene	96
32	Photomicrograph of hornblende in quartz monzonite	98
33	Photomicrograph of hornblende altered to chlorite, iron oxide, leucoxene and epidote	98
34	Photomicrograph of nearly automorphic biotite and ragged hornblende in quartz monzonite	99
35	Photomicrograph of biotite altered to muscovite with release of magnetite	99
36	Photomicrograph of plagioclase phenocrysts in microgranitic groundmass in quartz monzonite	102
37	Photomicrograph of central core of plagioclase phenocryst with inclusions of pigeonite, augite, apatite, biotite and magnetite	102



## LIST OF ILLUSTRATIONS (continued)

Figure	Page
38 Photomicrograph of plagioclase phenocryst with central core. Inclusions altered, compare with Figure 37.	103
39 Photomicrograph of same plagioclase in Figure 38	103
40 Photomicrograph of plagioclase phenocryst strongly altered deuterically	105
41 Photomicrograph of plagioclase crystal with albitic rim	105
42 Photomicrograph of large potash feldspar porphyroblast	107
43 Photomicrograph of small porphyroblastic potash feldspar	107
44 Photomicrograph of potash feldspar porphyroblast with well cleared core and poikiloblastic rim	109
45 Photomicrograph of poikiloblastic texture in anorthoclase-like potash feldspar porphyroblast	109
46 Phase diagram of alkali feldspars after Winchell	114
47 Probable phase diagram of alkali feldspars that have cooled very slowly, after Laves	114
48 Phase relations of material cooled rapidly enough to suppress exsolution, after Laves	114
49 Phase relations of alkali feldspars after Laves	114
50 Photomicrograph of anorthoclase partly replaced by orthoclase-micropertthite	117
51 Photomicrograph of plagioclase crystal partly replaced by anorthoclase	117
52 Photomicrograph of anorthoclase with relics of plagioclase	118
53 Photomicrograph of anorthoclase-like mineral partly replaced by another potash feldspar	118
54 Photomicrograph of plagioclase phenocryst partly replaced by potash feldspar	119
55 Photomicrograph of anorthoclase which replaced plagioclase and shape of original crystal retained	119
56 Photomicrograph of coarse micropertthite and quartz in alaskite	122
57 Photomicrograph of replacement granophyre in potash feldspar	122



## LIST OF ILLUSTRATIONS (continued)

Figure		Page
58	Photomicrograph of coarse-grained quartz monzonite with minor amount of fine-grained groundmass	128
59	Photomicrograph of coarse-grained quartz monzonite with no fine-grained groundmass	128
60	Photomicrograph of plagioclase phenocryst in aplitic groundmass	133
61	Outcrop of medium grained quartz monzonite on southeast slope of Jericho Mountain	136
62	Outcrop of quartz monzonite porphyry near Pauper claim	136
63	Intrusive aplite dike in quartz monzonite	145
64	Photomicrograph of texture of aplite in intrusive aplite dike near Julia mine	147
65	Photomicrograph of saccharoidal texture in intrusive aplite	147
66	Outcrop of quartz monzonite showing thin aplitic stringer	148
67	Pod of replacement aplite in quartz monzonite	148
68	Photomicrograph of plagioclase partly replaced by anorthoclase	155
69	Photomicrograph of olivine and magnetite in cluster of pyroxene in basal part of quartz monzonite porphyry sill in the Pauper area	160
70	Thin section showing genesis of magmatic quartz vein in alaskite (from U.S.G.S., P.P. 55 Pt. XXII)	170
71	Hand specimen showing tourmaline-pyrite segregation in quartz monzonite	176
72	Hand specimen showing tourmaline-pyrite segregation pod with leucocratic potash feldspar-rich selvage zone	177
73	Tourmaline-pyrite "sunbursts" on joint surface	177
74	Vein-like structures in deuteric stage, and quartz veinlet in hydrothermal stage	179
75	Paragenesis diagram of the ore minerals	181
76	Photomicrograph of quartz segregation stringer in quartz monzonite	190



## LIST OF ILLUSTRATIONS (concluded)

Figure	Page
96 Photomicrograph of tourmaline partly replaced by pyrite	254
97 Photomicrograph of brecciated arsenopyrite cemented by quartz	256
98 Camera lucida drawing of chalcopyrite replacing galena	258
99 Camera lucida drawing of chalcopyrite veinlet in and replacement of tetrahedrite	258
99a Sketch showing black tourmaline-quartz vein matter fractured and filled with white quartz veinlets which became brecciated and cemented with sphalerite, galena and quartz	263
100 Map of Lilly mine	261
101 Map of Blue Bird mine	269
102 Geochemical soil sampling grid at the Clark prospect	274
103 Photomicrograph of sericite wall rock alteration next to the Lilly vein	277
104 Older stage glacial drift in road cut on U. S. Highway 10 N about ¼ mi east of Elliston	286
105 Glacial erratic of andesitic tuff on Continental Divide northeast of Ontario mine	286
106 Air photo showing cirques on Cliff Mountain	288
107 Generalized tectonic map showing distribution of igneous rocks in Idaho and Montana	293
108 Variation diagram of the igneous rocks of, and associated with the Boulder batholith	297
109 Variation diagrams of the igneous rocks of, and associated with the Boulder batholith	298
110 Variation diagram of the Boulder batholith showing relation of igneous rocks of Idaho and western Montana	300
111 Orientation of couple which may have acted upon the region of the boulder batholith to have produced the mineralized structures	308
Geologic Map and Structure Sections	In Pocket



## ABSTRACT

The Elliston mining district comprises an area of about 85 square miles in the forest-covered mountainous region about 20 miles southwest of Helena, Montana. Sedimentary rocks, Mississippian to Lower Cretaceous, crop out in a small area in the northwest part of the district.

Upper Cretaceous volcanic rocks occur throughout the northwestern two-thirds of the district where they rest unconformably on sedimentary rocks. The volcanic section is about 5,000 feet thick, and comprises basic andesitic breccias at the base, andesitic and latitic pyroclastics, including welded tuffs, above, and basalt flows at the top. Volcanic derived sediments are minor in amount.

The Boulder batholith intruded the volcanic rocks probably in very early Tertiary and is represented by granodiorite followed by main stage quartz monzonite with related aplite-alaskite. Three main textural facies, coarse, medium and fine, and three stages of quartz monzonite crystallization are recognized.

1. The early magmatic stage is defined by crystallization of andesine, olivine (rare), pyroxene, hornblende and biotite.
2. The late magmatic or deuteritic stage is defined as potash feldspar-stable stage. The essential mineralogy includes crystallization of potash feldspars and quartz, and alteration of plagioclase.
  - a. Potash feldspars crystallized (1) in xenomorphic grains in the groundmass, (2) porphyroblastic growths and (3) by replacement of plagioclase, first by anorthoclase-

like minerals and then by orthoclase. Potash feldspars are somewhat perthitic in normal quartz monzonite, and strongly perthitic in alaskite. Orthoclase microperthite possibly formed by reorganization of albite accompanying replacement of andesine by potash feldspars at a relatively low temperature.

- b. Quartz occurs in xenomorphic grains in the groundmass and in granophyric intergrowths with potash and rarely plagioclase feldspars.
- c. Early formed plagioclase is altered to clinozoisite, albite and flecks of sericite.

Aplite and alaskite formed during the deuteric stage, the former in replacement and dilation dikes and in segregation pods, the latter by metasomatic replacement of quartz monzonite at the roof of the batholith. Tourmaline and lesser amounts of pyrite commonly occur in aplite-alaskite.

Deuteric stage segregations, structurally controlled segregation pods, and joint coatings of tourmaline + pyrite are surrounded by a leucocratic potash feldspar-rich selvage zone from which most of the calcium, iron and sulfur for the tourmaline and pyrite were derived.

- 3. The hydrothermal stage is defined by feldspar destruction to sericite, with or without structural control. A complete sequence of vein-like structures has been recognized from the deuteric to the early hydrothermal stage. Likewise, a



complete sequence of crystallization, without obvious structural control, has been recognized from the early magmatic to the early hydrothermal stage.

Sericite alteration occurs along all veins during the hydrothermal stage, but is more intense along veins formed at the early hydrothermal stage prior to the introduction of base metals. Base and precious metal mineralization formed in the late hydrothermal stage and may be, at least in part, locally derived.

Intrusive quartz monzonite changed volcanic rocks close to the contact to incipient hornfelses above normal quartz monzonite, and to microaplite above quartz monzonite which exhibits strong endomorphic features, especially above alaskite.

The structural events in the district are: (1) very mild warping and erosion prior to the early volcanics, (2) westerly tilting of sedimentary and volcanic rocks,  $5^{\circ}$  to  $30^{\circ}$ , prior to and/or accompanying intrusion of the batholith, (3) intrusion of the Boulder batholith which has gentle west dips on the roof, and at two localities apparently lenses out into sill-like bodies at the margin of the batholith, followed by (4) faulting, jointing, and vein formation, all tectonically related.

Veins in quartz monzonite (strike N.  $65^{\circ}$  E.  $\pm$  dip steeply S.) formed by many stages of fracturing. Early replacement lodes were followed by successive fissure fillings. The common vein minerals are: (in order), tourmaline, quartz, pyrite, auriferous arsenopyrite,



sphalerite, argentiferous galena, argentiferous tetrahedrite, chalcopryrite and ankeritic carbonates.

Fissure veins in volcanic rocks are generally narrow and include: (1) steep N.  $65^{\circ}$  E. veins, (2) steep veins parallel to NE faults, and (3) "flat" veins. The mineralogy is the same as above except for tourmaline.

The mineralogical and petrographic features, as well as chemical variation diagrams, suggest consanguinity of the batholithic and Upper Cretaceous volcanic rocks. The volcanic pile is considered to represent the first magmatic episode which was followed by the intrusion of quartz monzonitic magma in an essentially concordant body of batholithic dimensions. The formation of aplite-alaskite and the mineralized veins are late magmatic (deuteric) and latest magmatic (hydrothermal) stages in the crystallization of the batholith.

Batholithic rocks were exposed by erosion in early Tertiary time. An upland surface was developed which is now at about 7,200 feet. Rhyolite welded tuffs were extruded onto the partly dissected surface. The age of the rhyolite is not known, but a late Tertiary age is possible. Two stages of glaciation, one possibly pre-Wisconsin and another probably early Wisconsin are recognized.

GEOLOGY AND MINERAL DEPOSITS  
OF THE ELLISTON MINING DISTRICT  
POWELL COUNTY, MONTANA

INTRODUCTION

Purpose and Scope

The original objective of this investigation was to map the Elliston district, in reconnaissance, to determine the relationship between geologic structure and metalization for the purpose of aiding prospecting. As the study progressed, the scope was enlarged to include a more detailed geologic map and to investigate the genetic relations of the mineral deposits of the district to rocks of the Boulder batholith, as part of an investigation of this general problem by the U. S. Geological Survey. Permission to use the area for a thesis was granted by Dr. J. R. Van Pelt, President of Montana School of Mines.

The report covers the areal geology and the mineral deposits of the district. Particular stress is placed on the petrology and petrography of the intrusive rocks and on late magmatic phenomena leading to vein formation.

About three months of field work were spread over the 1950 - 1952 field seasons. Plane table maps and property examinations were made at this stage. Because the early work was done without the benefit of adequate base maps, more detailed mapping was desirable when air photos were obtained in 1953 and advance topography for the southern half of the district became available for the 1954 field season. Parts of the 1954 and 1955 seasons were devoted to additional mapping.



## History and Production

The region was first opened to settlement by the construction of the Mullan Road between Fort Benton, on the Missouri River, and Walla Walla, Washington, by Capt. John Mullan from 1855 to 1862. The road crossed the Continental Divide northeast of Elliston and followed Little Blackfoot Creek across the northern part of the district.

The discovery of gold in Last Chance Gulch at the site of the city of Helena, in 1864, brought prospectors into the area. The first discovery of gold in the Elliston district was probably made before 1870. From 1890 to 1909, the district had its greatest period of activity, (Pardee and Schrader 1933, p. 262) at which time a number of small gold and silver mines were in production. There has been periodic activity in the district ever since, but no important mines have been developed. The total production to date is estimated at somewhat more than \$3,100,000.

From 1909 to 1954, about 7,600 ounces of gold, 149,000 ounces of silver, 98,000 pounds of copper, 1,560,000 pounds of lead, and 197,000 pounds of zinc have been marketed from the district. Some placer gold was produced in the very early days, and ever since there have been sporadic attempts to operate small placers.

Although there has been virtually no mineral output during the past few years, there have been three organized efforts to bring properties into production for the recovery of base and precious metals. Two verified occurrences of uranium minerals in the district have been a mild stimulus to exploration during the past field season.



### Previous Work

The first serious geological investigation was made by Knopf (1913) who described the general geology of the Helena Mining region and made observations on some mining properties in the district. Billingsley and Grimes (1918) investigated some of the mineral deposits in the district in their discussion of the ore deposits of the Boulder batholith. Pardee and Schrader (1933) outlined the general geologic setting of the district and described some of the properties then active in their bulletin on the greater Helena mining district. Only incidental reference to the district has been made by other investigators.

### Acknowledgements

In its early stages, the project was conducted under the supervision of Montana Bureau of Mines and Geology. The counsel of the late Dr. F. A. Thomson, Dr. Eugene S. Perry, and Professor Walter S. March, Jr. are acknowledged. Messers Higby Williams, Richard Marvin and Frank Coupal ably assisted in the 1949, 1950, and 1951 field seasons, respectively.

I am grateful for the time which the following have devoted to this thesis: to Professor G. E. Goodspeed under whose supervision the thesis was compiled. To Professor Goodspeed and Dr. J. Hoover Mackin for the time they spent in the field, and to these and Dr. Howard A. Coombs and Dr. Julian D. Barksdale for their suggestions relative to those parts of the manuscript which they read.

I am indebted to Dr. Montis R. Klepper and many of the geologists of the U. S. Geological Survey party who are mapping in the northern

part of the Boulder batholith, for many stimulating discussions of the geology of the region. From these discussions, two joint papers were compiled and prepared on late magmatic and vein formation phenomena with Dr. Klepper and Mr. Darrell Pinkney (1956). Mr. Edward Ruppel has made a copy of his map in the adjoining area east and south available to me. Dr. Randolph W. Chapman, formerly with the Survey, provided a chemical analysis from his unpublished manuscript. The attempt has been made to credit the sources of all ideas and specific information obtained from these sources.

The cooperation and good will of mine owners, operators, and interested persons in the Elliston district are much appreciated. Special thanks are due Messers David and Leo Newman and the late Charles Anderson. Commissioner Frank E. Dougherty, Dr. Fee, the Hopkins brothers, the Best brothers, the Ward brothers, Messers Lester Thomson, Elimer Lince, Henry Neumiller, "Casey" Jones, W. L. Henderson and others provided useful information.



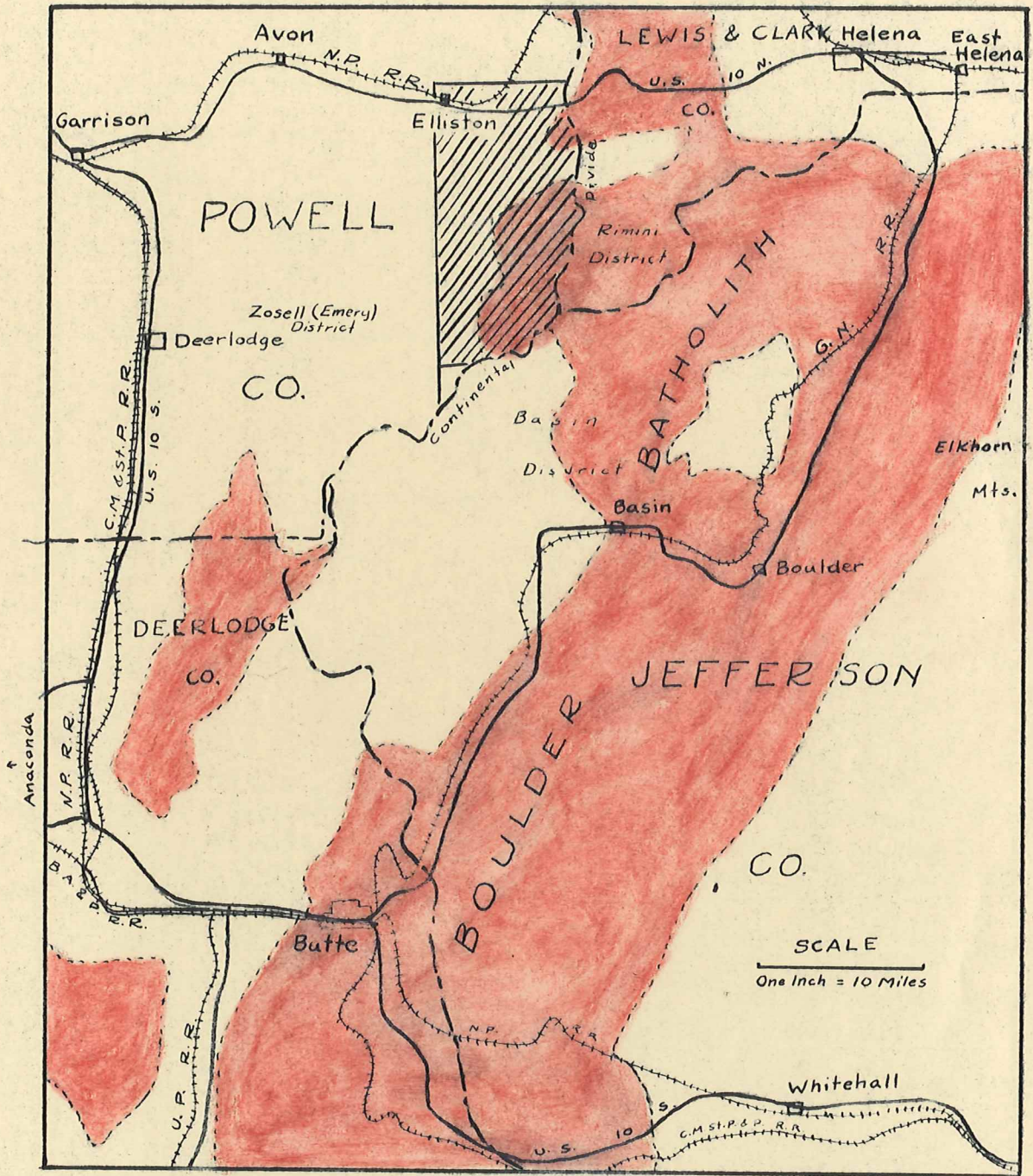


Figure 1. Index Map showing location of the Elliston district with respect to the Boulder batholith in southwestern Montana. (Boulder batholith in red)



active mines and prospects. **GEOGRAPHY** the mines and prospects have a downhill grade to the railroad.

### Location and Definition of District

**Vegetation and Climate**  
The Elliston district comprises an area of about 85 square miles in the southeastern corner of Powell County, Montana, as shown in the index map in Figure 1. The center of the district is at about latitude  $46^{\circ} 30'$  north, and longitude  $112^{\circ} 20'$  west, about 20 miles west southwest of Helena.

The district includes that part of a broad mineralized area located south of Elliston and west and north of the Continental Divide. East of the Divide, in Lewis and Clark County, lies the Rimini district, and south of the Divide in Jefferson County, is located the Basin district. Thus the Elliston district is defined by geographic rather than geologic boundaries.

### Accessibility

Elliston is located on U. S. Highway 10 N, 31 miles by paved highway west of Helena as shown on the index map in Figure 1. Likewise, the main transcontinental freight line of the Northern Pacific Railroad passes through Elliston.

The district is well situated for easy access to two smelters. The East Helena smelter of the American Smelting and Refining Company is about 35 miles east of Elliston, and the Anaconda smelter of The Anaconda Company is located about 80 miles southwest of Elliston.

Within the district, graded roads are maintained by the U. S. Forest Service up the Little Blackfoot and Telegraph Creeks. Numerous mine roads and jeep trails provide access to all active and recently

of northeasterly or northwesterly trending valleys which have moderately straight courses.

The highest point is on Bison Mountain at 8028 feet, and the bench mark at Elliston at 5046 feet is near the lowest point. Bison, Nigger and Treasure Mountains have a local relief of about 1800 feet.

Although the valley sides are steep, particularly along the Little Blackfoot Creek, the summit and divide areas are marked by gentle slopes. Above 7200 feet (plus or minus 200 feet), the topography is a gently rolling upland, particularly noticeable in the southeastern part of the district and along parts of the Continental Divide.

#### GENERAL GEOLOGIC SETTING OF THE BOULDER BATHOLITH

The Elliston district lies astride the northwestern border of the Boulder batholith (Figure 1). The batholithic rocks crop out in the eastern and southern parts of the district, and the pre-batholithic volcanics in the western and northern parts. Within the district, most of the main geologic episodes relative to the history of the Boulder batholith are recognized. The distribution of the principal rock types is shown on the geologic map.

In the region of the batholith, sedimentary rocks were deposited without evidence of strong structural deformation from late pre-Cambrian time throughout the Paleozoic and much of the Mesozoic. During the early Upper Cretaceous, the sedimentary rocks were strongly warped. Upper Cretaceous volcanic rocks were spread over an erosion surface which truncates the sedimentary beds from Mesozoic (Colorado group)





Figure 2. General view of Nigger Mountain taken from the air and looking northwest. Note timbered slopes and open parks. Dump at Big Dick shaft shows as white patch near top of mountain. Dump at the drainage level is the white patch near center of photo. The valley of Little Blackfoot Creek is behind Nigger Mountain in the left background. Note steep glaciated slopes visible over the top of the mountain.



to upper Paleozoic (Madison limestone). The thickness of the volcanic pile is about 5000 feet.

Subsequent to the outpouring of the volcanics the region was subjected to strong folding, followed by thrust faulting, followed by the emplacement of the Boulder batholith. The batholith was not intruded as a single body of magma at a given time, but was formed by a succession of intrusive episodes. Rocks of gabbroic and basic granodioritic composition are represented in the early stages, followed by the main body of quartz monzonite with related aplite and alaskite. Over an area of approximately 1500 square miles the batholith was intruded into a position relatively close to the earth's surface, in many areas probably less than 5,000 feet from the surface.

In the Radersburg area, sediments interbedded with pre-batholithic volcanics have yielded late Cretaceous or Paleocene fossils (Klepper, personal communication). Northwest of Helena, at Blossburg, lower Oligocene plants are reported by Knopf (1913, p. 36) in sediments which rest on eroded quartz monzonite. An early Tertiary age, possibly Eocene, is indicated for the batholith. From a completely different approach, the age has been determined to be about 61 to 72 million years, based on zircon age determinations. (Chapman 1953)

After the batholith was partly exposed by erosion, dacite and rhyolite were spread over parts of the area and rest unconformably on granitic and older rocks. Intrusive dikes and plugs of the same composition have been recognized cutting quartz monzonite, notably in the Butte area. In the Pleistocene, the region was glaciated. A small ice cap centered in an area just south of the Elliston district.



## Geologic History in Elliston District

Paleozoic and Mesozoic sedimentary rocks are exposed in the northern part of the district. The oldest formation is the Mission Canyon limestone of the upper Madison (Mississippian). The sedimentary rocks were only slightly warped prior to the outpouring of the volcanics which rest unconformably on upper Kootenai and lower Colorado beds. The amount of warping and erosion in this area prior to the extrusion of the volcanics was not as great as in the area east of the batholith.

The volcanic rocks comprise a series of flows, pyroclastics, and some volcanic-derived sediments. The composition changes from the base upward, from basic andesite to andesite to latite to basalt. The thickness of the volcanics is probably about 5,000 feet.

Following the accumulation of the volcanic rocks, the region was slightly folded prior to the emplacement of the Boulder batholith. The mild deformation in this district is in striking contrast to the strong deformation on the east side of the batholith.

Two intrusive stages of the batholith are recognized in the district, and in the immediately adjacent areas east and west, two additional stages are found. Alaskite and small bodies of aplite are related to late magmatic crystallization of the quartz monzonite.

The mineral deposits of the district are found principally in veins, the dominant trend of which is about N. 65° E. in the batholithic rocks. Gold, silver, lead, zinc and minor amounts of copper, are the metals which have been produced. The valuable minerals are localized in relatively small ore shoots in long vein zones. The most productive

veins are in quartz monzonite, however some ores have been mined from veins in the volcanic rocks.

Rhyolite was subsequently extruded onto eroded batholithic and older rocks. Some of the rhyolite is welded tuff which is not necessarily related to the petrogenetic province of the Boulder batholith, whereas some rhyolite in adjoining areas is clearly related to the batholith.

Two stages of glaciation have been recognized, neither of which is the youngest glacial stage recognized regionally. During one stage, a small ice cap was present in the southern part of the district. Glaciers moved out radially from the cap eastward across the Continental Divide, and northward in the Little Blackfoot Creek as far north as Elliston.

#### SEDIMENTARY ROCKS

Sedimentary rocks crop out over an area less than five square miles in the northwest part of the district. Except for the area immediately north of Elliston, exposures are very poor. The rocks have not been mapped in detail, except in the area of good exposures, and only incidental attention has been given to petrologic characters of the rocks.

#### Madison Group (Mississippian)

Limestone of the Madison group is the oldest rock known in the district. Only the Mission Canyon formation is exposed; if the lower, or Lodgepole formation is present in the region, it is covered by



glacial debris. The Mission Canyon limestone is overlain by shales of the Big Snowy group. The thickness of the formation in the district is less than 1000 feet; Sloss and Hamlin (1942) indicate a possible 1,500 feet of Madison in the adjoining region.

The Mission Canyon formation is composed of thick, massively bedded buff to bluish-white limestone which is fine grained, locally dolomitic, and which contains scattered chert nodules in a few beds. The formation is well exposed in the western quarries of the Elliston Lime Company two miles east of Elliston (NW  $\frac{1}{4}$  sec. 5, T. 9 N., R. 6 W.). In the eastern quarry, the structure is very complex. The massive limestone has been sheared and recrystallized to a friable, poorly consolidated breccia. The shearing is probably caused by the large fault which is postulated to be present in the adjacent valley.

#### Big Snowy (?) (Mississippian)

Dark gray shales of the Big Snowy group (?) are very poorly exposed. They are of undetermined thickness and uncertain correlation. Dr. O. D. Blake, who discovered a few exposures of shale, expressed the opinion that they may be correlated to the Heath formation of central Montana, which is Upper Mississippian in age. (Personal communication) The shales overlie the Madison and underlie the Amsden formation. Nowhere, however, have the contacts been observed.

West of the outcrops of Mission Canyon limestone and east of the nearest Amsden outcrops, there is an interval which, if not complicated by faulting, would represent possibly 1,000 feet of section in the

Table 1. Stratigraphic Section of Sedimentary Rocks

<u>Age</u>	<u>Formation or Group</u>	<u>Lithology</u>	<u>Thickness</u>
Pleistocene and Recent		Alluvium, glacial outwash gravels in stream valleys, morainal deposits on valley slopes, terminal moraines.	
Mid- Cretaceous (?)	Colorado group	Black and gray shales, silts and mudstones. The Fall River (?) quartzitic sandstone is 50-100 ft thick and occurs about 50 ft above the base.	200 +
Lower Cretaceous	Kootenai formation	Red and green shales, "salt and pepper" sandstones, siltstones. Gastropod limestone at top, quartzite and black chert pebble conglomerate at base.	1,500 ±
	Unconformity		
Jurassic ? to Lower Cretaceous ?	Morrison formation	Varicolored shales, mudstones, silts and sandstones.	100 +
Upper Jurassic	Swift formation	Conglomeratic sandstone at base, thin limestone and buff colored sandstones	20 - 35
	Unconformity		
Permian	Phosphoria formation	Dark chert and quartzite with 4 ft oolitic phosphate rock in middle	33
Pennsylvanian Lower Permian	Quadrant formation	Limestone, thin shale and quartzitic beds overlain by more massive, white, fine grained quartzite above with interlayered limestone	300 ±
Mississippian part Pennsylvanian ?	Amsden formation	Red and buff shales at base, overlain by gray dolomite and thin quartzitic beds, the latter more abundant at top	200 - 300
Mississippian	Big Snowy ?	(Covered) Dark shales	?
Mississippian	Madison group Mission Canyon Formation	Massive, thickly bedded, white to gray limestone, uppermost beds are somewhat dolomitic and buff colored, very minor nodular chert	1,000 ±
	Lodgepole formation	Limestone, more thinly bedded than Mission Canyon	Not exposed



position of the Big Snowy group. So great a thickness is considered excessive by Dr. Blake. Glacial debris covers the hillslopes and only in one or two of the deeper gulches are the underlying shales visible.

#### Amsden Formation (Mississippian-Pennsylvanian ?)

The lower contact of the Amsden formation is not exposed, but the upper part of the formation grades conformably into the overlying Quadrant formation. It is uppermost Mississippian (Chester) in age (Perry 1953, p. 11) and in its uppermost beds may be Pennsylvanian in age (Dr. H. E. Wheeler, personal communication). The beds assigned to the Amsden in the district are estimated to be from 200 to 300 feet thick.

The base of the formation is marked by reddish shales which rarely crop out, but which are clearly identified by the red color imparted to the derived soils. Thinly bedded limestone, with interlayered dolomite and thin quartzitic sandstone beds, comprise the remainder of the formation. Quartzitic beds become progressively more numerous and thicker in the upper part of the formation. The exposures are so poor that few details were easily obtainable about the lithologies of the formation. In the region east of the batholith, the Amsden has been described as a "heterogeneous assemblage of sandy, argillaceous, and dolomitic and clarcareous rocks." (Klepper and Freeman 1954 p. ) The description would adequately describe the rocks noted in the small number of outcrops east of the Quadrant outcrops about one mile east of Elliston.

## Quadrant Formation (Pennsylvanian-Permian)

The Quadrant formation grades conformably from the underlying Amsden below and without distinguishable unconformity into the Phosphoria above. The formation is generally considered to be of Pennsylvanian age (Perry 1953, p. 11) but in its upper part it may well be lower Permian in age (Wheeler, H. E., personal communication). The thickness is about 300 feet.

The formation consists of quartzite, limestone, and some shale. Shale and limestone are more abundant at the base, and contain thin beds of quartzitic sandstone, whereas quartzite is more abundant in the upper part of the formation. (For mapping purposes, it is customary to consider the first thick quartzite in the Amsden-Quadrant section the base of the Quadrant formation. In the Elliston district, three conspicuous quartzite layers are in the section; one at the base of the formation, another about 150 feet above the base, and a third, which is the most massive and is about 80 feet thick, at the top of the formation. The quartzites are composed of clean, white sands and are only moderately iron stained along major joints. The shaly and thinly-bedded limestones in the formation rarely crop out, but are usually marked by grassy and sparsely forested slopes. Large slabs of quartzite float are abundant in the soil over the limestone layers, and tend to give an incorrect impression of the amount of quartzite actually present in the formation.



### Phosphoria Formation (Permian)

The Phosphoria formation rests more or less conformably on the Quadrant formation and is marked by an erosional unconformity above, which may have cut out a part of the original Phosphoria section. The Elliston district is near the northern boundary of the outcrop area of the Phosphoria formation (Sloss 1950). The thickness is but 33 feet.

The base of the formation consists of 18 feet of massive dark chert and fine quartzite in which a thin phosphatic bed occurs near the top of the chert. An oolitic phosphate-rock bed four feet thick, in which a 0.1 foot shale parting occurs near the middle of the bed, overlies the basal chert. The upper part of the formation is dark chert and quartzite.

Phosphate rock from the Phosphoria formation is being mined in the Avon district, about 15 miles northwest of Elliston, and in the Garrison district about 30 miles west of Elliston. Because of its possible potential as a future economic resource, a detailed section, measured by Paine (1952) and the writer is given in Table 2.

### Swift Formation (Jurassic)

The Swift formation rests unconformably on the Phosphoria formation. The relations with the overlying Morrison formation are uncertain. The Swift formation is the uppermost member of the Ellis group of Upper Jurassic age (Perry 1943, plate 7). The formation is about 20 to 35 feet thick, and is composed of thin, buff-colored quartzitic sandstone and thin limestone. The base of the formation is

a conglomerate with scattered, well-rounded, dark chert pebbles. A bluish-gray limestone, which contains many marine fossil fragments, was noted at one locality at the top of the formation.

#### Morrison Formation (Jurassic ?-Lower Cretaceous ?)

The Morrison formation is poorly exposed in the district. It overlies the Swift formation and underlies the Kootenai formation. Elsewhere in Montana it is a much thicker formation, but in the Elliston area only about 100 feet of section is considered to represent the formation. The age of the Morrison is uncertain. According to Perry (1953) it is uppermost Jurassic or very earliest Lower Cretaceous. The formation is composed of purplish-gray to dark gray varicolored shales and mudstones.

#### Kootenai Formation (Lower Cretaceous)

The Kootenai formation is Lower Cretaceous in age. It rests unconformably on the Morrison and may grade conformably into the Colorado group above. The basal sandstone is widely distributed in western Montana, and it is an important oil producing horizon in central Montana (Perry 1953). The formation is estimated to be about 1500 feet thick in the Elliston district.

Quartzitic sandstones and red silty shales are the principal lithologic types found in the formation. The basal quartzitic sandstone is about 100 feet thick and contains a black chert pebble conglomerate at the base. Several other dark chert pebble beds are found in the



Table 2

Measured section of Phosphoria formation NE  $\frac{1}{4}$  sec. 2, T. 9 N.,  
R. 7 W. (Paine 1952, 79-81).

Ellis formation (unmeasured)

1 Sandstone and 2 + basal conglomerate  
of subangular to round quartzite and  
chert pebbles cemented in slightly cal-  
careous tan sandstone

Phosphoria formation	<u>feet</u>
Zone E	
26- Quartzite, gray, sandy, medium- grained, slightly resistant, poorly exposed generally	2.0
25- Quartzite, gray, fine-grained, re- sistant, forms ridge, weathers red- dish; lower 2 feet cherty	9.6
24- Covered, probably chert or sandy quartzite	3.9
23- Quartzite, gray, medium-to-fine-grained, cherty, resistant, forms ridge, weathers grayish red.	2.5
Zone D	
22- Phosphate rock, black to blackish, speckled gray, white, oolitic, nodular, weathers gray, nonresistant.	0.4
21- Phosphate rock, gray, speckled white, oolitic, shaly, coarse-grained, non- resistant.	
20- Phosphate rock, gray, shaly, thin- bedded, oolitic, fine-grained, non- resistant.	0.4

19- Phosphate rock, black, speckled white, oolitic, coarse-grained, nonresistant	0.2
18- Phosphate rock, gray, ferruginous, oolitic, fine-grained, weathers white to bluish phosphate bloom, non resistant	0.3
17- Shale, gray, phosphatic, thin bedded nonresistant	0.1
16- Phosphate rock, gray, siliceous, glauconitic, oolitic, coarsely crystalline, nonresistant, massive	0.2
15- Phosphate rock, gray oolitic, medium-to-coarse-grained, massive, weathers reddish, nonresistant	0.5
14- Phosphate rock, gray, oolitic, fine-to-medium-grained, thin bedded weathers reddish	0.3
13- Phosphate rock, gray like above but massive.	0.9
12- Phosphate rock, black to grayish black, shaly, ferruginous, thin-bedded, slightly oolitic.	0.4

## Zone C

11- Siltstone, tan, slightly calcareous, nonresistant.	0.3
10- Chert, black to black gray, quartzitic.	0.3
9- Siltstone, black, argillaceous, calcareous, nonresistant, weathers whitish	0.4
8- Chert, black to gray, massive, generally covered with talus from below, weathers gray.	
7- Chert, black, thin-bedded, nonresistant, weathers gray	0.4



6- Chert, black, like bed 8.	0.4
5- Phosphatic chert, black, speckled white, forms dip slope of chert ridge; appears to be gradational in next lower unit.	1.3
4- Quartzite, gray, grading into buff at base, cherty, massive, resistant, forms crest of ridge	2.1
3- Quartzite, buff, cherty, less massive resistant, ridge former	1.0
2- Chert, gray, nodular, thin-bedded in irregular layers, resistant, ridge former.	3.7
1- Chert, gray, massive, base of ridge	<u>0.3</u>
Total	32.7

Quadrant formation (unmeasured)

Upper 1 ft covered, below massive white quartzite which forms back slope of the above ridge.

upper part of the quartzite. The entire unit is a coarse "salt and pepper" quartzitic sandstone which is well exposed northeast of the railroad station at Elliston. Several additional sandstone beds are interbedded with red silty shales in the lower part of the section, but sandstones are thin and poorly developed in the upper part of the formation. The upper part is generally poorly exposed, but is indicated by deep red colors imparted to the soils. The top of the formation is marked by a thin, gray to nearly black limestone which is commonly crowded with gastropods. The "gastropod bed" so-called, is about 20 to 30 feet thick. The bed is particularly well exposed about one-half mile north of Hat Creek at the extreme west edge of the map (NW cor. sec. 2, T. 8 N., R. 7 W.), at which locality a basaltic sill splits the limestone. The beds are nearly vertical in a small area, influenced by faulting just west of the boundary of the map.

On the west side of Little Blackfoot Creek, west of Nigger Mountain, (SW  $\frac{1}{4}$  sec. 1, T. 8 N., R. 7 W.) dark gray limestone underlies volcanic rocks. While no gastropods were seen, the limestone is tentatively correlated with the "gastropod bed" at the top of the upper Kootenai formation. The limestone is sparsely mineralized at the locality mentioned above.

#### Colorado Group ? (Middle Cretaceous ?)

Colorado shale overlies the "gastropod bed" of the Kootenai formation without obvious unconformity. The shales are the youngest Cretaceous sedimentary rocks in the district and are overlain unconformably, by Upper Cretaceous volcanics. The formation is considered



lowermost Upper Cretaceous (Perry 1953). About 200 feet of sandstone and shale are found in the Elliston district.

Above the limestone bed at the top of the Kootenai formation there are about 50 feet of gray shales to the base of the Fall River sandstone which is a fine-grained, massive, quartzitic sandstone about 50 feet thick. Above the Fall River sandstone, about 100 feet of gray shales and silty shales are rarely exposed. About one-quarter mile north of Hat Creek at the west edge of the volcanics, the Colorado shales crop out, however, the contact with the overlying volcanics is not exposed.

On the west side of Little Blackfoot Creek opposite the Hopkins mine, shales which are considered to be of Colorado age have been locally altered to low-grade spotted hornfelses. The same formation, exposed on Slate Creek about one-half mile west of the Little Blackfoot Creek, is not metamorphosed, and is a dark gray, somewhat fissile shale.

#### IGNEOUS ROCKS

Three distinct groups of igneous rocks are found in the district: a pre-batholithic volcanic pile, the intrusive Boulder batholith, and post-batholithic volcanics. All of the volcanic rocks, both pre-batholithic and post-batholithic, are described before the batholithic rocks. Within each main group of rocks, the petrographic types are discussed approximately in their chronological order.

PRE-BATHOLITHIC VOLCANIC ROCKS

## Introductory Statement

The older volcanic rocks rest unconformably on Colorado and upper Kootenai formations. They include flows, flow breccias, a variety of pyroclastics and minor amounts of volcanic-derived sediments. The rocks range in composition from basalt to latite but andesite is the most abundant type.

Because the earlier volcanic rocks are poorly exposed throughout the northern and western parts of the district, they are mapped as a single unit. In most areas, the exposures are so poor that detailed sections are unobtainable. However, along the Little Blackfoot Creek in the southwestern part of the district, an incomplete but reasonably well exposed section of volcanic rocks was sampled. (See Table 4). An excellent section was measured at Cliff Mountain, southwest of the district. (See Table 5.) These two sections, coupled with the observations made throughout the district, form the basis for the generalized section of volcanic rocks given in Table 3. This is a composite section of nearly 5000 feet of volcanic rocks, but at no locality is so complete or thick a section known.

## Basic Andesitic Breccias

## Areal Distribution.

The breccia at the base of the volcanic section is well exposed in the north central and western part of the district. On Nigger



Table 3

## Composite Section Upper Cretaceous Volcanics

Emery Basalt (Uncertain pre- batholithic age)	Porphyritic basalt flows commonly amygdaloidal "oatmeal" basalt and flow breccias. Thin interlayered andesitic tuff beds	1,000 plus
Unconformity ?		
Latite and Andesite  (upper part latitic)	Latite welded tuffs and flows 0 - 1,000 (?) with interlayered andesitic tuffs and basalt flows in upper part.	600-800
(lower part andesitic)	Andesitic pyroclastics including welded tuffs, lapilli tuffs and tuff-breccias. Interlayered basalt and diorite porphyry sills and volcanic- derived sediments of andesitic composition	
Basic Andesitic Breccias	Flow and pyroclastic breccias, ? - 1,500 massive basic andesite, possibly partly intrusive. Basalt porphyry sills interlayered.	3,000-5,000

## Table 4 Little Blackfoot Creek Section

(Secs. 12, 13, 14, T. 8 N., R. 7 W.) North Side of Creek

Emery Basalt

9. Basalt flows, porphyritic basalt, pyroxene phenocrysts 500 +  
 Porphyritic basalt flow with "green" amygdaloidal zone  
 Amygdaloidal basalt

Andesite-Latite (Lower Part Only)

8. Andesitic tuff { Irregular discontinuous  
 banding in crystal  
 tuff. } 200 +
7. Diorite porphyry--intrusive sill 50-200
6. Andesitic tuff breccia 200
5. Andesitic lapilli tuff, partly welded 200
4. Porphyritic basalt, pyroxene phenocrysts  
 (intrusive sill ?) 50
3. Andesite breccia 200
2. Hornblende andesite intrusive sill 200 +

Basic Andesitic Breccia

1. Basic andesitic breccias 1500 +



Table 5 Cliff Mountain Section

(From Base of Cliff Mountain to Sugar Loaf Mountain)

<u>Emery Basalt.</u>		Thickness
15. Emery basalts, porphyritic and amygdaloidal		800 +
14. Basalt flow breccia, dark gray		300-400
<u>Latite and Andesite</u>		
13. Latitic welded tuff (one or two units ?)	{ Tuff breccia Tuff, lapilli fragments Welded tuff	500 ?
12. Porphyritic basalt flow, plagioclase phenocrysts		100
11. Latite tuff breccia		?
10. Andesitic tuff breccia		?
9. Latite tuff (probably welded)	{ Fragmental tuff, wispy bands Lower part strongly banded, 2" -6" greenish and reddish tuff	200-300
8. Basalt flow (top unit at Cliff Mountain)		108
7. Andesitic tuff breccia		30
6. Basalt flow (?)		34
5. Andesitic tuff breccia	{ Polymictic breccia, coarse pebbles; Lapilli tuff Welded lapilli tuff	31
4. Andesitic tuff, water lain. Carbonate matrix some woody material at base ?		38
3. Red andesitic mud flow deposit (?), fragments 50 percent to 1 ft diameter, "avalanche" deposit ?		190
2. Andesitic tuff (probably welded)	{ Lapilli tuff, small frag 60' Massive tuff, few fragments, miarolitic cavities 225' Strongly banded tuff, welded 20' glassy base 2-3'	310
<u>Basic Andesitic Breccias</u>		
1. Andesite breccia		150+ 2800-3500 feet

Mountain the breccia pile is possibly more than 1500 feet thick; on the hills west of Little Blackfoot Creek, the breccia, which rests on Colorado shale, is several hundred feet thick; and in the northeast part of the district, there are breccias of undetermined thickness. The breccias do not persist to the south, and distinctly thin out to the west.

Two kinds of breccias have been recognized. One is a very thick, oligomictic\* breccia which is found on Nigger Mountain. It is partly a massive basic andesite, partly a flow, partly conspicuously brecciated, and possibly partly intrusive. The brecciation is best shown on weathered slabs which show differential coloration of the fragments and the matrix. On fresh surfaces, the breccia structure is commonly indistinct. Along the west side of Nigger Mountain, the flow breccia (?) unit is massive, fine grained, and resembles basalt, but petrographically it is a basic andesite. The rock, in this area, is not conspicuously brecciated.

The other unit, which clearly lies on Colorado shale in the area west of Little Blackfoot Creek, but which does not clearly underlie the basic andesites on Nigger Mountain (although it may), can be traced from Hat Creek and the southwestern segment of Little Blackfoot Creek to

---

\*Oligomictic conglomerates (Petijohn 1949) are one rock type conglomerates. The term should serve a useful purpose in the description of volcanic breccias. Polymictic conglomerates are conglomerates with several varieties of rock types.

---



Cliff Mountain. It is probably a pyroclastic breccia in which there is a variety of types of fragments.

Figure 3 shows an outcrop of the breccia on the Hat Creek Road just east of the gate about  $\frac{1}{2}$  mile west of Little Blackfoot Creek. Although it is a polymictic breccia, all the fragments are similar in fundamental mineralogy and porphyritic textures.

#### Petrographic Types.

In the massive dark andesite at the northwest base of Nigger Mountain, plagioclase phenocrysts, 1 to 3 mm in length, comprise 30 to 50 per cent of the rock. The phenocrysts are andesine, about Ab 55 An 45. The plagioclase phenocrysts in some thin sections are quite fresh; in others they are strongly altered to clinozoisite and sericite along fractures within the crystals. The pyroxene crystals are almost completely altered to light green, slightly pleochroic uralite, with or without chlorite. Some bluish-green to olive green hornblende was noted at some localities, especially in the Big Dick area near the top of Nigger Mountain. Small accessory magnetite and secondary dusty iron oxide grains are abundant in the rock. Although a variety of textures are found in the rock, the most common is a felted texture shown in Figure 4. The groundmass is strongly altered to clinozoisite, sericite, chlorite and other very fine-grained alteration products.

The pyroclastic breccia is very similar in its fundamental petrographic features to the flow breccia. Slightly altered andesine and strongly altered pyroxene are the principal essential minerals. The





Figure 3. Outcrop of basal andesitic pyroclastic breccia on Hat Creek road, NW $\frac{1}{4}$  sec. 13, T. 8 N., R. 7 W.

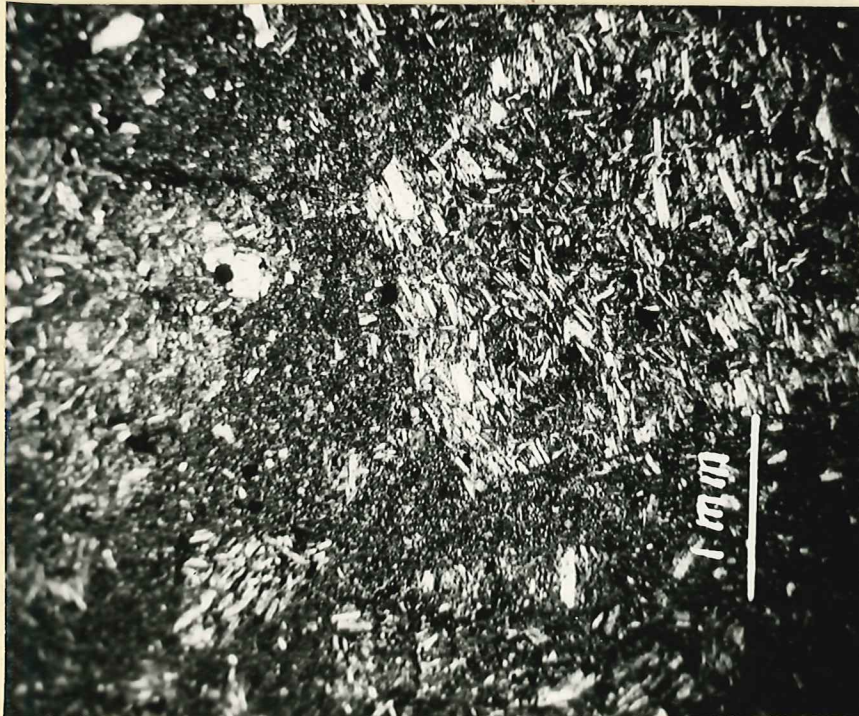


Figure 4. Photomicrograph showing microlitic to felted texture and small fragments in groundmass of basal andesitic breccia at Cliff Mountain. Crossed nicols. (28-65-1) (Thin section number in brackets)



groundmass textures are more complex than in the flow breccia. In part, the texture is felted to microlitic, as shown in Figure 4. Devitrification textures, in which strongly sutured, poikilitic, microgranitic feldspar grains occur, indicate that some of the groundmass was formerly a glass.

#### Origin of Breccias.

The breccias have been very incompletely studied. Detailed mapping would undoubtedly provide useful information which would lead to a better understanding of the origin of each of the breccia units and their interrelations. The impression is gained that the massive unit of near-basalt composition is a thick body, mostly extrusive but possibly partly intrusive, which shattered and brecciated as a consequence of minor movements within the main mass following partial solidification.

The pyroclastic breccia may be an explosive breccia of the neûée ardent type, or possibly a rubbly flow-breccia formed at the advancing front of a flow. The petrographic similarities of the fragments, despite the color differences noted on the outcrop, suggest the possibility of an auto-brecciated mechanism similar to that described by Curtis. If lateral, textural and structural continuity can be demonstrated between Little Blackfoot Creek and Cliff Mountain, the neûée ardent mechanism seems the most reasonable for the unit.

## Latitic and Andesitic Rocks

## Lithologic Types and Distribution.

Andesitic and latitic volcanics overlie the basal, dark andesitic breccias and comprise a series of welded and pyroclastic tuffs and breccias. Associated with these rocks are interlayered basaltic flows, minor amounts of volcanic-derived sediments, and variable amounts of intrusive sills.

Andesite is more abundant in the lower portion of the section, and latite is generally more abundant in the upper part. Andesitic rocks are more abundant in the western side of the district, whereas latitic rocks are more abundant in the southern part. Latite is conspicuously absent in the Little Blackfoot Creek section, which is compiled from rocks exposed on the north side of the creek, whereas about one mile south of Little Blackfoot Creek, latite is exposed. A local topographic high at the time of the latite eruptions or erosion following the eruption of the latite, may account for their absence north of Little Blackfoot Creek.

## Petrography of Units in Little Blackfoot Creek Section.

The andesitic rocks in the Little Blackfoot Creek section are found on the mountain north of the creek, north of Larabee Gulch, and on the ridge which extends west to the western border of the district. Outcrops are few, however a reasonably complete section was obtained as shown in Table 4, p. 26. The three andesitic volcanic units described



might be traceable stratigraphic units, and although the descriptions are arranged in their stratigraphic sequence, the purpose is to describe the petrographic characters with only incidental consideration of their stratigraphic importance. Only the petrographic characters of the volcanic rocks are herein described. The numbers in the following headings refer to the numbers in the table.

Andesitic Lapilli Tuff. (5) The first andesitic tuff unit above the basal breccias is separated by at least 200 ft of intrusive sills. The lapilli tuff is poorly exposed and is about 200 ft thick at the locality measured in NE  $\frac{1}{4}$  sec. 14, T. 8 N., R. 7 W. A single representative hand specimen serves only to identify the petrographic type. It is a dark purplish-brown tuff with small lithic fragments averaging  $\frac{1}{2}$  in with some individuals to 2 in in diameter.

The groundmass is incompletely devitrified glass and contains some fine-grained chalcedonic quartz in thin devitrified layers. Figure 5 is a photomicrograph of the rock in plane polarized light showing what appears to be squashed shards and a few plagioclase crystals in a groundmass which was originally glass. The phenocrysts are probably andesine altered to sericite and clay minerals. The microscopic structures indicate that the unit maybe a welded tuff. Welded tuffs of andesitic composition are discussed under a separate heading in this section on the pre-batholithic volcanic rocks.

Andesitic Tuff Breccia (6). The next overlying andesitic unit, which is about 200 feet thick, is a mottled tuff breccia with inconspicuous fragments. The rock is light gray with dark reddish flow layers. All

the fragments are a single rock type and grade from sand size to angular fragments several inches in diameter.

Plagioclase crystals of andesine composition are of two generations; the distinctly early crystals, which are rarely more than 1 mm in length, have automorphic forms, are internally fractured, and in some instances are broken and resorbed; the smaller crystals of the younger generation have imperfectly developed crystal forms and are conspicuously fresher. Although not abundant, there are a few phenocrysts of potash feldspar with 2V about 60°. Pyroxene phenocrysts were formerly abundant, and comprised as much as 20 per cent of the rock. However, they are almost completely altered to antigorite and a carbonate mineral. Figure 6 shows altered pyroxenes in plane polarized light, and concentrations of released iron oxide at the margin of the altered grains. The groundmass textures are devitrified glass in which some of the reddish layers are still glassy.

Andesitic Tuff (8). A diorite porphyry sill lies between the tuff breccia and the overlying andesitic tuff (8). The upper tuff is probably more than 100 ft thick and possibly as much as 200 ft thick. The uncertainty is due to the fact that the top of the unit is not exposed at one locality, and on the next hill to the west, the base is not indicated because of a minor fault. The lower part of the unit, which is strongly layered, grades upward into a massive rock with smeary and irregular lenses and layers. The lower part is distinctly welded. Compressed and flattened shards are conspicuous as shown in Figure 7. Chlorite and a carbonate mineral occur in many clusters which suggest that mafic minerals were formerly abundant.



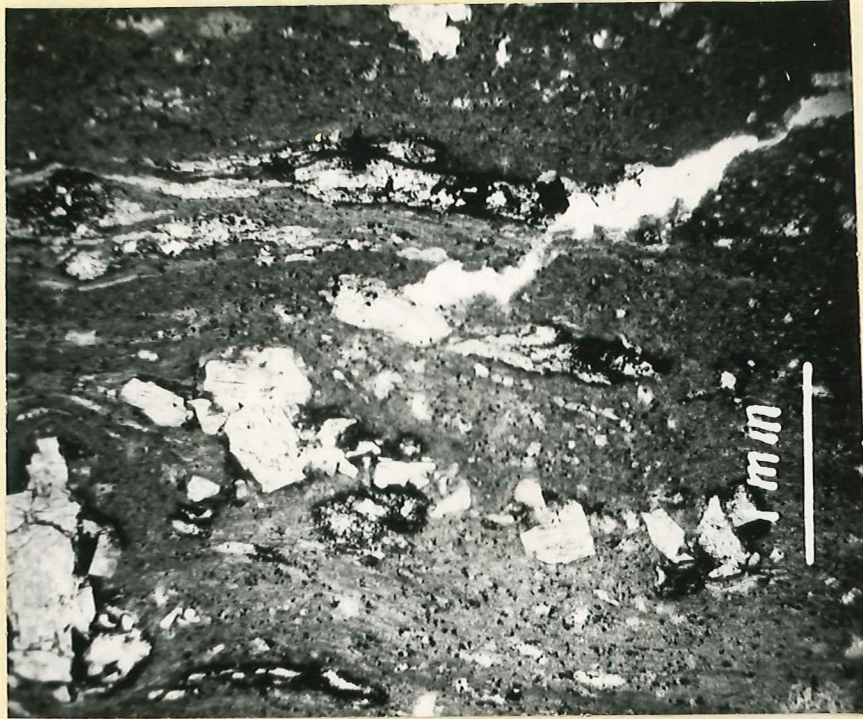


Figure 5. Photomicrograph of andesitic tuff. Note somewhat squashed layers. Plane polarized light.

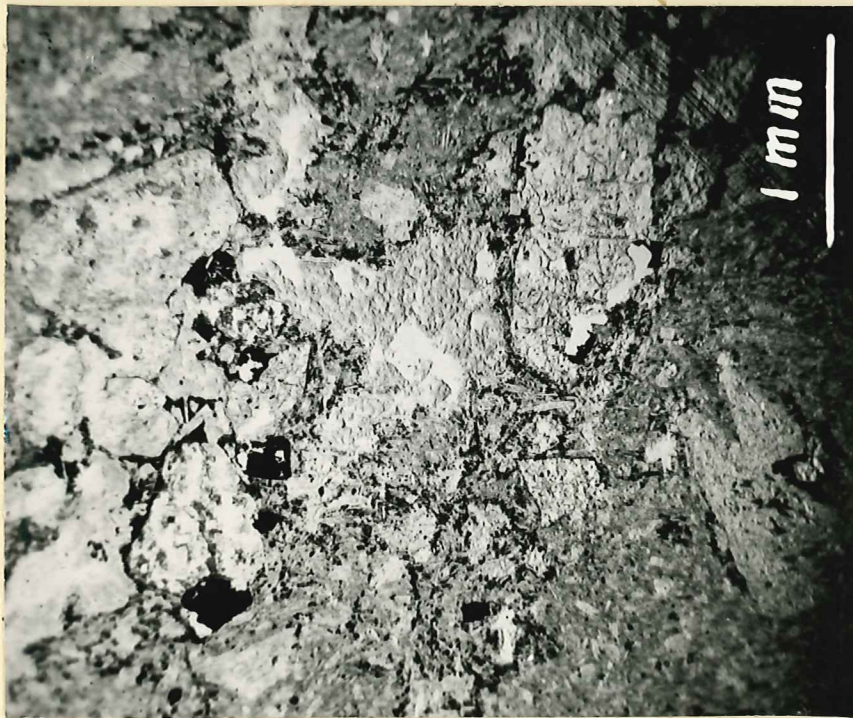


Figure 6. Photomicrograph of altered pyroxene phenocryst. Note distribution of opaque grains around the borders of the altered grains. Polarized light. (39-115-3)





Figure 7 . Photomicrograph showing partly collapsed shards in glassy matrix. Crossed nicols. (39-117-3)



Figure 8 . Photomicrograph showing crowded plagioclase phenocrysts in nearly glassy groundmass. Note resorption of larger phenocrysts and the clear, sharp margins of the smaller plagioclase phenocrysts. Also note orientation of inclusions in larger phenocrysts. Plane polarized light. (39-117-2)



The upper part, which in the field, was thought to be continuous with the lower part already described, is not noticeably welded and is, at least in part, a crystal tuff in which plagioclase crystals comprise about 30 per cent of the rock, and mafic relics about 5 per cent. The larger automorphic to hypautomorphic early plagioclase phenocrysts are about 1 mm long, are strongly fractured, and are in part resorbed by the groundmass as shown in Figure 8. Small plagioclase phenocrysts (0.5 mm and less) are quite fresh and are calcic andesine. The larger phenocrysts were probably well formed at the time of extrusion, whereas the smaller crystals appear to have crystallized in situ. The mafic minerals have been completely altered to chlorite, variety penninite.

A more detailed study of the field relations and a more complete collection of samples will be necessary before the relations of the two parts of this unit are completely understood.

Amygdaloidal basalts lie unconformably on the andesitic welded tuffs and are part of the Emery basalts.

#### Petrography of Units in Cliff Mountain Section.

At Cliff Mountain, the andesitic volcanics are very well exposed on the east face of the mountain and were measured in detail. From the top of the mountain west to Sugar Loaf Mountain, latite, with some interbedded andesitic units, are quite well exposed, but they were not sampled and measured in detail. Each of the andesitic and latitic units in this section is described to indicate the petrographic types. The numbers in the headings refer to the numbers in Table 5, p. 27.

Andesitic Tuff (2). This andesitic tuff is 310 ft thick and rests on a unit of basal breccia. At the base of the tuff, 2½ ft of glass, presumably welded, is overlain by prominently layered massive rock which appears rather homogeneous on close examination in the field. The layered part grades upward into a zone of blocky, massive tuff with abundant small miarolitic cavities and indistinct wispy lenses.

The strongly layered part of the unit contains strongly flattened shards as shown in Figure 9. Small plagioclase phenocrysts to 1 mm are abundant in the rock, many are in part resorbed by the groundmass, some are bent and broken. Both the plagioclase phenocrysts and small plagioclase crystals in the nearby glassy groundmass, are altered to sericite. A small amount of altered biotite and hornblende (?) is represented by very fine grains and patches of chlorite with included sphene and epidote. A few small lithic fragments are present.

The main body of the unit is about 165 feet thick and makes a massive, bold, outcrop, the upper 100 ft of which is a cliff. The rock breaks into large blocks suggestive of massive columns. Flat wispy layers and flattish miarolitic cavities are found throughout this part of the unit. The uppermost part of the tuff, which is above the cliff and is about 60 ft thick, is a dark red andesitic lapilli tuff in which there are sparsely distributed small rock fragments, one of which is shown in Figure 10.

The upper part of the unit is not so conspicuously welded, however, in thin section, small wisps of collapsed shards are clearly indicated. Plagioclase phenocrysts, which comprise as much as 20 percent of the



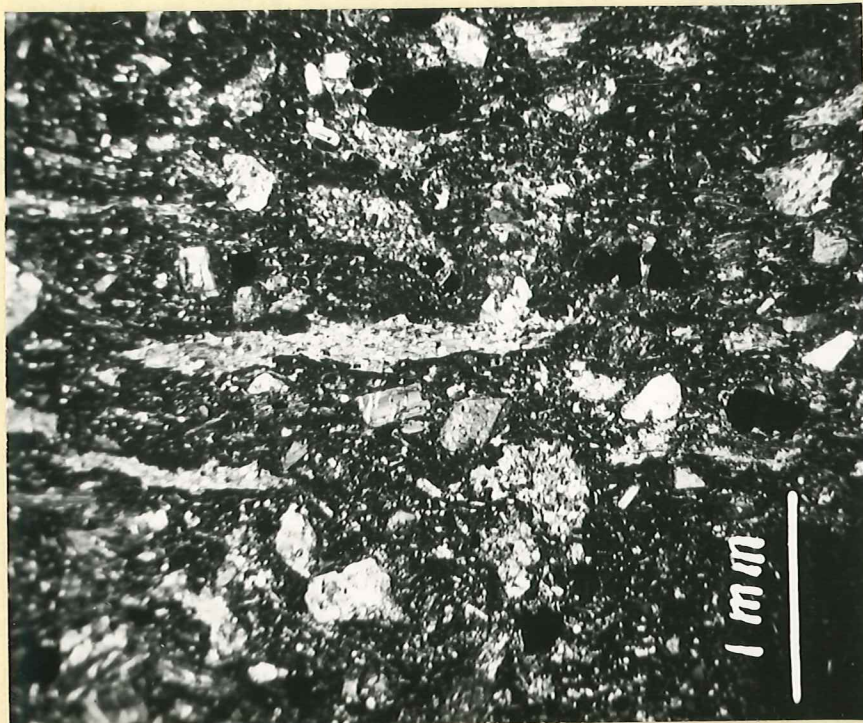


Figure 9. Photomicrograph showing collapsed and compacted shards in slightly welded part of andesitic tuff (Unit No. 2 Cliff Mountain section. Crossed nicols. (28-63-1-1)



Figure 10. Photomicrograph of lapilli tuff in massive part of Unit No. 2, Cliff Mountain. Plane polarized light. (28-63-1-3)



rock, are fairly fresh but internally fractured, and show a greater tendency to form albitic alteration at the rims along fractures than in the phenocrysts lower in the unit. Biotite is strongly altered to clusters of opaque grains, probably magnetite and sphene. Sphene is actually quite abundant in the rock and is also found in clusters with coarse epidote which may represent former mafic minerals other than biotite.

Andesitic Tuff-Breccia (7). A thin tuff-breccia, only about 30 ft thick, is fine lapilli tuff at the base and coarse polymictic breccia at the top. The lower part of the unit is conspicuously layered and contains a few small fragments. Some plagioclase crystals are fresh, others are altered to clinozoisite; the mafic minerals are completely altered to chlorite, dusty iron oxide, and some sphene.

The middle part of the unit is a lapilli tuff with numerous lithic fragments distinguishable in hand specimen and in thin section. Plagioclase phenocrysts are in part altered to clinozoisite, and in part resorbed by the groundmass. However, no marked albitic rims were observed, a feature which is characteristic in the upper part of the unit. Plagioclase probably comprises 70 percent of the rock, (exclusive of the lithic fragments); and mafic minerals, now completely altered to chlorite, (variety penninite), and sphene, comprise about 10 percent of the rock. The rock is a lithic and crystal tuff, as shown in Figure 11.

The breccia zone at the top, which is about 20 ft thick, is characterized by an abundance of fragments varying in size from fine grains to rather coarse cobbles. A wide variety of rock types are



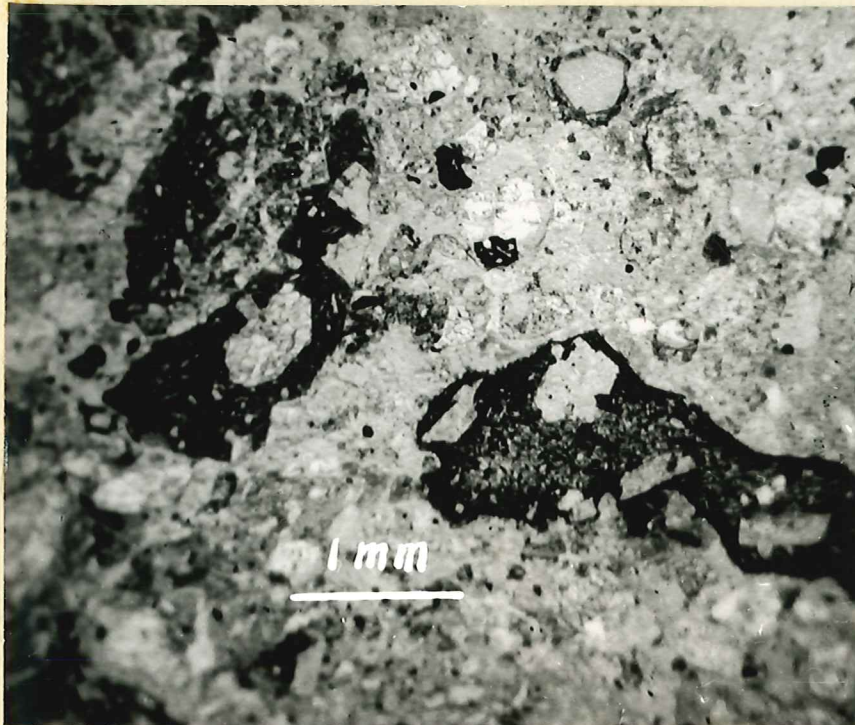


Figure 11. Photomicrograph of andesitic tuff breccia. Note plagioclase crystals and lithic fragments. Plane polarized light.



Figure 12. Photomicrograph of welded crystal tuff of latitic composition. Plane polarized light.

represented in the fragments. In thin section, the groundmass is seen to be much more highly altered than the groundmass in the middle and lower parts of the unit. The plagioclase phenocrysts are in part resorbed by the groundmass and altered to albite and clinozoisite along fractures in the crystals. Several distinctly different sizes of plagioclase crystals are present, grading from 0.1 mm downward to minute, altered microlites in the groundmass. The mafic minerals are in large part altered to chlorite, but a few relics of augite are preserved.

Latite Tuff (9). The first unit west of the ridge on Cliff Mountain is a dark red latite with grayish green layers. In the one thin section from the unit, the groundmass is dominantly glass which is partly devitrified. It is probably a welded tuff as suggested in Figure 12. There are two generations of plagioclase phenocrysts; the early crystals are in part resorbed and altered, whereas the younger plagioclase crystals are sharp and clear. A small amount of potash feldspar is also present. Many small lithic fragments are in the rock; some possibly pumice fragments are conspicuously flattened. There are several varieties of volcanic fragments.

Andesitic Breccia (10). The next overlying unit is a porphyritic andesite breccia which weathers to a chalky gray color. Plagioclase, both in phenocrysts and microlites in the groundmass, is partly fresh, and partly altered to sericite and albite, particularly along internal fractures in the phenocrysts. Biotite is strongly altered to chlorite and sphene. The groundmass is partly very fine grained, partly glass.



Latite Tuffs and Breccias (11) and (13). The upper units in the latitic part of the section are poorly exposed and incompletely represented by the specimens and field notes. The lower part of 11 (assuming that it is all one unit) is a fine-grained tuff with thick layers 2 to 6 in wide, which were described as "flow layers". In one part of a thin section, plagioclase phenocrysts are well oriented, as if by flowage, whereas in another part, the rock appears to be a tuff with numerous glass shards.

In the upper part of the unit, the rock is a breccia, the matrix of which has a brownish-colored glass. Figure 13 is a photomicrograph which shows a lapillus in a crowded crystal tuff matrix. Potash feldspar and biotite are present, the latter is partly altered to a white mica and released iron oxide.

Unit 13 is probably about 500 ft thick, if it is a single unit. The lower part is a latite welded tuff which has a dominantly glassy groundmass, in part devitrified, and contains conspicuous compaction lenses characteristic of welded tuff and shown in Figure 14. Andesine and a few potash feldspar phenocrysts are fresh and comprise possibly 10 percent of the rock. A very small amount of quartz was also identified. A few small lithic fragments were noted in thin section.

The top of the unit is a latitic lapilli tuff breccia. It may represent a pyroclastic breccia or a poorly welded, coarse, fragmental ignimbrite. A weathered boulder is illustrated in Figure 15. In thin section the tuffaceous groundmass contains altered biotite, hornblende and strongly altered feldspars. Plagioclase is altered to clinozoisite

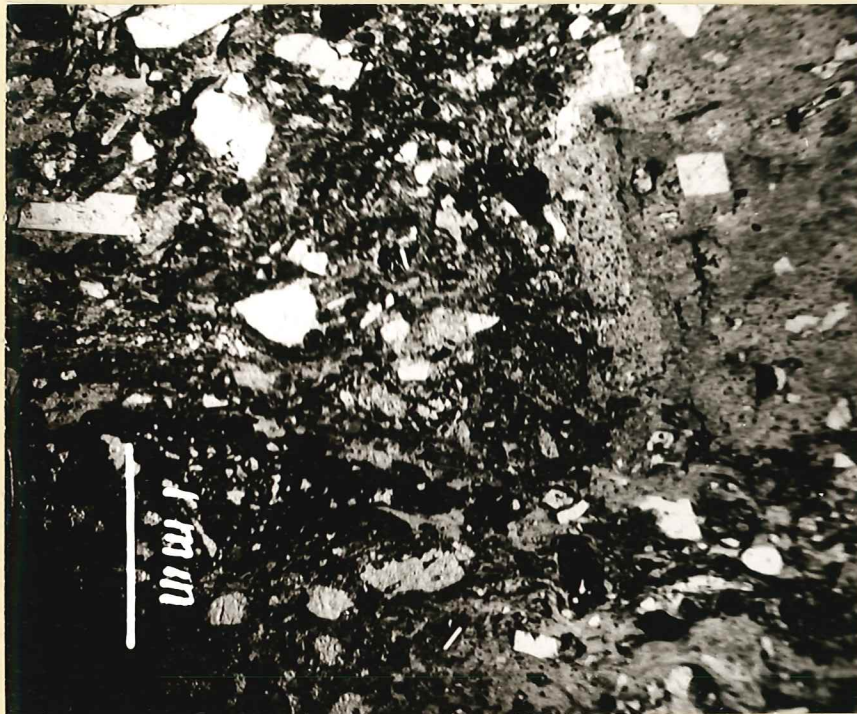


Figure 13. Photomicrograph of lapillus fragment in crowded crystal tuff matrix. Note compacted layers in lower right hand corner of photograph and irregular shards (?) in lower center. Crossed nicols. (28-67-3)





Figure 14. Outcrop of welded latitic tuff at Cliff Mountain.  
Note "wispy" lenses.

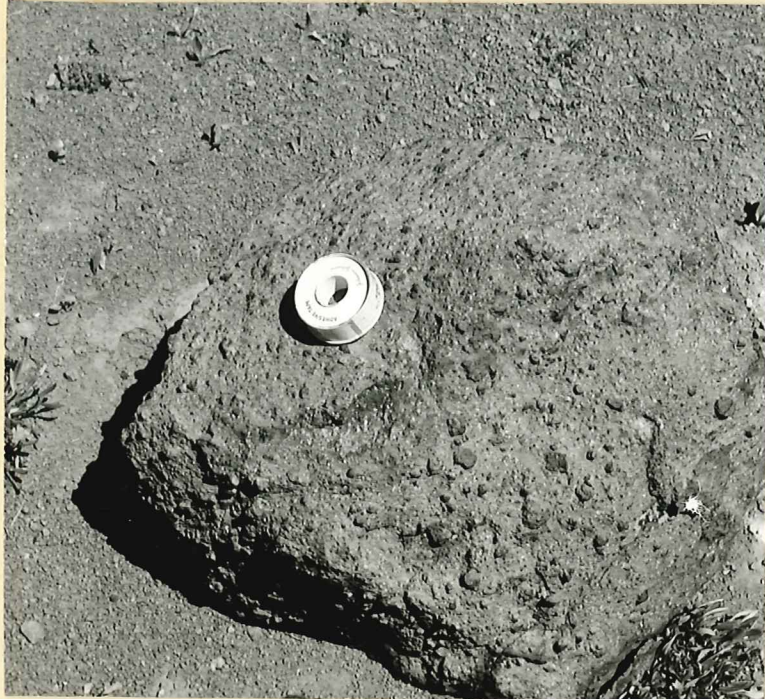


Figure 15. Lapilli tuff from upper part of latitic welded tuff,  
Cliff Mountain.



and sericite; biotite is altered to permenite and iron oxides; and hornblende is quite fresh, pleochroic from light green to olive brown.

#### Other Latite Localities

On the southwest side of Treasure Mountain, several hundred feet of latite is exposed, likewise on the south side of Little Blackfoot Creek (in sec. 13, T. 8 N., R. 7 W.) latite is thick. On the east side of Ontario Creek (in SE  $\frac{1}{4}$  sec. 13, T. 8 N., R. 7 W.), and on the southwest side of Bison Mountain, the latites are welded tuffs in which the groundmass was formerly mostly glass, but now is partly devitrified into micro-sutural, micro-poikilitic grains. The larger of two generations of plagioclase phenocrysts are partly resorbed. The younger, smaller plagioclase phenocrysts are quite fresh. The composition of all the plagioclase is andesine, and all the phenocrysts taken together comprise about 10 per cent of the rock. Potash feldspar phenocrysts, fewer in number than plagioclase, are partly altered to clay minerals. Biotite is strongly pleochroic  $x =$  light tan,  $y = z =$  reddish brown,  $2V = 0^\circ$ . The accessory minerals include magnetite, apatite and very minor amounts of zircon.

#### Welded Tuffs of Andesite Composition

Most, if not all, of the published descriptions of welded tuffs are for rocks of rhyolitic or quartz latitic composition. (See bibliography for section on rhyolite in the following part on post-batholithic volcanics.) At least two of the andesitic units (numbers 2



and 5 in the Cliff Mountain section, see Table 5, p. 27, and probably numbers 2 and 8 in Little Blackfoot Creek section, see Table 4, p. 26) are welded tuffs. The uppermost latite in the Cliff Mountain section (No. 13) and the latite exposed on the east side of Ontario Creek (SE  $\frac{1}{4}$  sec. 12, T. 8 N., R. 7 W.) are welded tuffs.

The composition of the volcanic units is based on the mineralogy only, and is not supported by chemical analyses. The andesitic volcanic units contain andesine and hornblende without recognizable potash feldspar and quartz. The latites have potash feldspar and biotite in addition to andesine and hornblende, but no essential quartz.

The welded features are suggested on many isolated outcrops by thin wispy lenses. Figure 14, p. 45 is a photograph of a reddish latitic boulder which is virtually identical in appearance to a welded tuff specimen collected in the Iron Springs district, Utah, from one of the welded tuff units mapped and described by Mackin and Nelson (1950).

The andesitic tuff unit No. 2 in the Cliff Mountain Section (Table 5, p. 27) has a succession of features characteristic of welded tuffs, starting with a thin welded glass at the base of the unit and progressing upward through a dense, welded, microlitic, wispy-layered zone, to an incompletely welded massive columnarly jointed tuff at the top.

The microscopic features of the welded andesitic and latitic tuffs are similar to illustrations published by Mansfield and Ross (1935) and Gilbert (1939) for rhyolitic welded tuffs in Idaho and eastern California. The collapsed shard structures are the clearest microscopic indication

of compaction. The lenses and layers of welded shards are not so well illustrated in the andesitic welded tuffs because of devitrification textures which are so commonly encountered in the andesitic tuffs and which tend to obliterate the diagnostic glassy structures.

One possible criterion which suggests, (but does not necessarily prove) compaction structures vs. flow structures in glass, was recognized in a latitic welded tuff. In the dense, strongly welded part, flowage structures are a conspicuous feature. If the structures were produced by laminar flow, a consistent direction-sense should be expected between any pair of vectors representative of differential movement. Consequently small drag folds, and congestion of flow bands at the corners of crystals frozen while in the process of being rotated, should have the same direction-sense.

On the other hand, flowage produced by compaction and squeezing of glass around an obstacle, such as a crystal or rock fragment, should produce different direction-senses for differential movement on each side of the obstacle. In a densely compacted, mostly glassy latite, different direction-senses were observed over the thin section, and within a single field of observation about 2 mm in diameter. Figure 16 is a photomicrograph which shows two plagioclase phenocrysts arrested in rotation in strongly-banded glass. The figure does not definitely indicate the type of flowage, it merely indicates a criterion for determination of the direction-sense of the flow layers. Both phenocrysts are being acted upon by a couple. The upper vector faces left, and the lower faces right.



Figure 17 shows two phenocrysts, the upper appears to have pushed the middle crystal into the flow layers. Note the many small drag folds and concentrations of layers on either side of the middle crystal. (The cloudy bright part of the picture is a devitrified patch. Note that the flow layers are merely interrupted by the patch indicating a static devitrification of the glass.) The orientation of the drag folds is not clear enough to positively indicate compaction. In one section, oppositely oriented direction-senses were observed across a layer of glass less than 0.1 mm wide. Such folds could not have been produced by laminar flow. Differential compaction stresses could produce these structures.

If this is not the first, then it is one of the few accounts of welded tuffs of andesitic composition. No obvious reason is seen why the mechanics of extrusion of a tuff of rhyolite composition, which becomes welded and which forms a widespread stratigraphic layer, would not be applicable, under favorable conditions, to a tuff of andesitic composition. Bibliographic references to textures and structures of welded tuffs and the mechanics of their possible emplacement are discussed under the younger volcanics.

#### Chemical Analyses of Andesite and Latite.

Chemical analyses of andesites and latites which may be representative of volcanic rocks found in the Elliston district are given in Tables 6 and 7. Table 6 lists chemical analyses of three andesites from the region of the Boulder batholith but not close to the Elliston district. Analysis No. 1 is the composition of an andesite from the upper

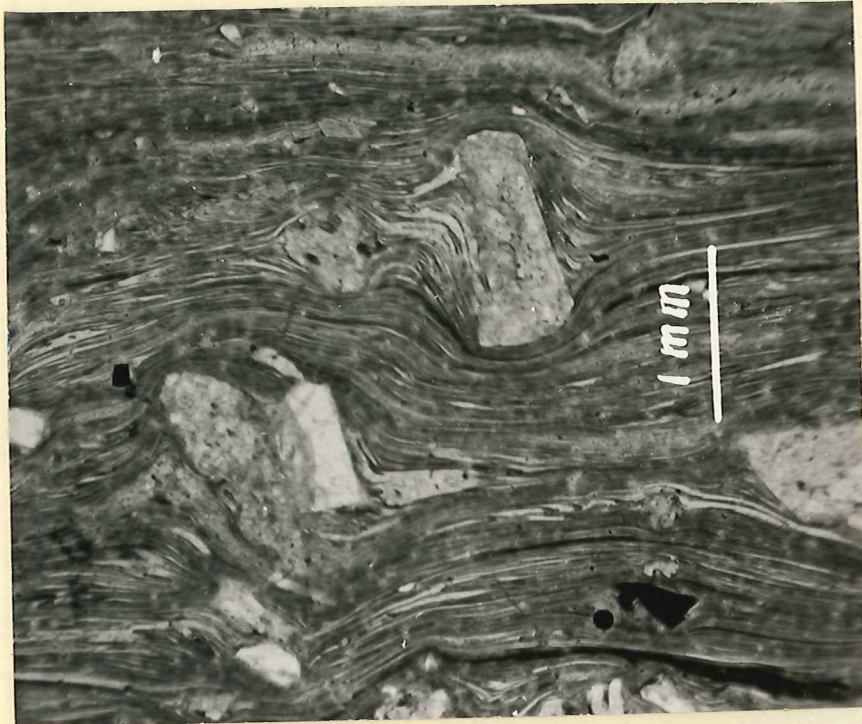


Figure 16. Photomicrograph showing compaction structures in latitic welded tuff. Note relative direction sense of movement. Plane polarized light. (31-115-2-2)



Figure 17. Photomicrograph of compaction structures in latitic welded tuff. Note small drag folds which indicate direction sense of relative movement. Plane polarized light. (31-115-2-2)



part of the Elkhorn Mountain volcanic pile. The second analysis is representative of an andesite porphyry near Butte. The third analysis is representative of an andesite from the Radersburg mining district and which has been subjected to hydrothermal alteration.

Table 7 contains a chemical analysis of latite from Thunderbolt Creek, which is a few miles south of the Elliston district. It is probably representative of the composition of one of the latitic units in the volcanic section of the district.

The chemical composition of the andesites is similar to quartz monzonite of the batholith. The normative analyses show virtually identical orthoclase; the andesites have about 6 to 8 per cent less quartz, and comparably more plagioclase. The Elkhorn Mountain andesite is in the same subrange as five quartz monzonite analyses from the Butte district. (Washington 1917)

X

#### Emery Basalt

Porphyritic and amygdaloidal basalt flows at the top of the older volcanics have been described from the Zosell (Emery) mining district. (Robertson 1951). The name Emery basalt was informally suggested by Dr. Klepper (personal communication). Although basaltic flows dominate the section, a few thin, dark reddish brown, andesitic tuffs are inter-layered with the basalt flows in the Zosell district.

In the Elliston district, basalt flows and sills which are virtually identical to basalt flows in the Emery basalts are inter-layered in the andesitic and latitic volcanic sequence and comprise possibly 1000 ft of flows above the andesite-latite units.

Table 6

## Chemical Analyses of Andesites

	1	2	3
SiO <sub>2</sub>	59.64	62.53	60.78
Al <sub>2</sub> O <sub>3</sub>	16.64	19.01	18.10
Fe <sub>2</sub> O <sub>3</sub>	2.33	1.96	3.15
FeO	4.35	1.44	.97
MgO	2.10	1.29	2.04
CaO	4.59	5.17	4.61
Na <sub>2</sub> O	3.31	3.45	2.81
K <sub>2</sub> O	4.16	3.30	2.75
H <sub>2</sub> O <sup>+</sup>	.69	.45	2.62
H <sub>2</sub> O <sup>-</sup>	.14	.21	.68
TiO <sub>2</sub>	1.08	.65	.24
P <sub>2</sub> O <sub>5</sub>	.49	.17	.62
MnO	tr.	.03	.39
S	.01		
SO <sub>3</sub>			.16
CO <sub>2</sub>	none		
BaO	.10	.13	.18
SrO	.05		.04
Li <sub>2</sub> O	tr.	tr.	
Total	99.68	99.83	100.10



## Normative Analyses

	1	2	3
Quartz	11.16	17.04	11.64
Orthoclase	25.02	19.46	16.12
Albite	27.77	29.34	26.20
Anorthite	18.07	25.02	26.97
Diopside	.92		3.56
Hypersthene	9.19	3.20	5.26
Magnetite	3.25	3.02	6.03
Ilmenite	2.13	1.06	1.37
Apatite	1.34	.34	1.01

## References

1. Andesite, Elkhorn Mountains, Elkhorn, Montana.  
Weed 1901, p. 525 and Washington 1917, p. 361.
2. Andesite porphyry, Butte, Montana.  
Weed 1900, p. 119.
3. Andesite, altered (hydrothermally) Keating Mine,  
Radersburg, Montana.  
Washington 1917, p. 837. (one of three very similar  
analyses)

Table 7

## Chemical Analyses of Latites

SiO <sub>2</sub>	64.45
Al <sub>2</sub> O <sub>3</sub>	17.69
Fe <sub>2</sub> O <sub>3</sub>	1.33
FeO	1.93
MgO	.57
CaO	3.73
Na <sub>2</sub> O	3.85
K <sub>2</sub> O	3.68
H <sub>2</sub> O <sup>+</sup>	.80
H <sub>2</sub> O <sup>-</sup>	.59
TiO <sub>2</sub>	.69
P <sub>2</sub> O <sub>5</sub>	.16
MnO	.05
S	.04
CO <sub>2</sub>	.29
BaO	.19
Total	100.04

Latite, Thunderbolt Creek, Helena Mining Region, U.S.G.S.

Bull. 527, p. 26, 1913.



The basalts are poorly exposed and are found exclusively in the western part of the district, over an area of several square miles, mostly north of Little Blackfoot Creek. Several types of basalts are recognized including porphyritic basalt with pyroxene phenocrysts, porphyritic basalt with pyroxene and plagioclase phenocrysts, and "oatmeal" basalt which contains flat tabular plagioclase phenocrysts. The flows are commonly amygdaloidal. One unit is strongly brecciated.

#### Porphyritic Basalt with Pyroxene Phenocrysts.

Several flows consist of dark greenish gray basalt with nearly black pyroxene phenocrysts of augite and diopsidic augite. Some of the flows contain minor amounts of very small olivine crystals which are in part altered to fibrous amphibole, or antigorite, but which yield very little iron oxide in the alteration process. Indistinct, small lath-shaped plagioclase phenocrysts are commonly aggregated into glomerophytic clusters in a microdiabasic groundmass. Magnetite, and small amounts of apatite, are the important accessory minerals. Analyses 1 and 5, Table 5, are representative analyses of rocks of this type. A typical model analysis is:

Labradorite (Ab 40 An 60)	31.5%
Pyroxene	7.8
Olivine	9.2
Magnetite	4.5
Groundmass	47.0

Amygdaloidal zones are commonly found at the top of these flows as well as in porphyritic basalts with both pyroxene and plagioclase phenocrysts. In the amygdaloidal parts of the flows, the groundmass is noticeably finer, and the essential minerals are strongly altered deuterically.

#### Porphyritic Basalt with Pyroxene and Plagioclase Phenocrysts

These basalts are mineralogically similar to the basalts above. The rocks weather to dark brown on the outcrop, but on the fresh surfaces they are dark greenish gray, aphanitic, and contain abundant phenocrysts which range from about 2 to 4 mm, and may comprise as much as 25 percent of the rock. Labradorite laths occur in individual crystals and in glomerophytic clusters, and are more abundant than augite and clinoenstatite. Clinoenstatite is recognized in thin sections and is conspicuous because it is so commonly altered to fibrous amphibole and chlorite. The optical properties determined are: positive optic sign,  $2V = 59^\circ \pm 2^\circ$ , birefringence relatively low, maximum extinction angle  $2\Lambda c$  about  $22^\circ$ . Magnetite and apatite are the common accessory minerals. Small amygdules composed of calcedonic quartz and a carbonate mineral were noted in some thin sections.

A typical model analysis of the rock from NW cor. sec. 9, T. 7 N., R. 8 W. is given in the following table:

Labradorite Ab 33 An 67	16.8%
Augite	3.7
Clinoenstatite	6.3
Groundmass	73.2



Analysis 3 Table 5 is representative of the basalt found in the Zosell mining district.

"Oatmeal" Basalt. A distinctive porphyritic basalt was found in the basalt sequence just west of the Elliston district. The rock has a dark gray groundmass in which well-oriented plagioclase phenocrysts, to 10 mm long, are observed. The crystals are distinctly tabular and are about the size of rolled oatmeal.

The plagioclase phenocrysts comprise 10 to 20 per cent of the rock, have the composition of labradorite, about Ab 42 An 58, and are altered along internal fractures to a more albitic composition, as shown in Figure 18. Plagioclase is likewise abundant in the felted groundmass, the individual crystals averaging about .2 mm in length. Pyroxene phenocrysts to 1 + mm are rare, and inconspicuous, and in thin section are observed to be partly altered to uralite. In the groundmass, pyroxene is strongly altered to secondary amphibole which is pleochroic from bluish green to pale greenish tan. Small opaque iron oxide grains are present but they are not abundant. Sphene, partly altered to leucoxene, is quite abundant, comprising 2 to 3 per cent of the rock. Sericite and carbonate occur in the groundmass.

The rock is lithologically very similar to a basalt flow observed in the upper part of the Upper Cretaceous volcanics found in the South Boulder Creek section southeast of Whitehall.

Amygdaloidal Basalt. Amygdaloidal zones at the top of basalt flows are abundant. Amygdules range in shape and size from small spherical shapes of sand size, to almond-shaped amygdules up to an inch in length, to

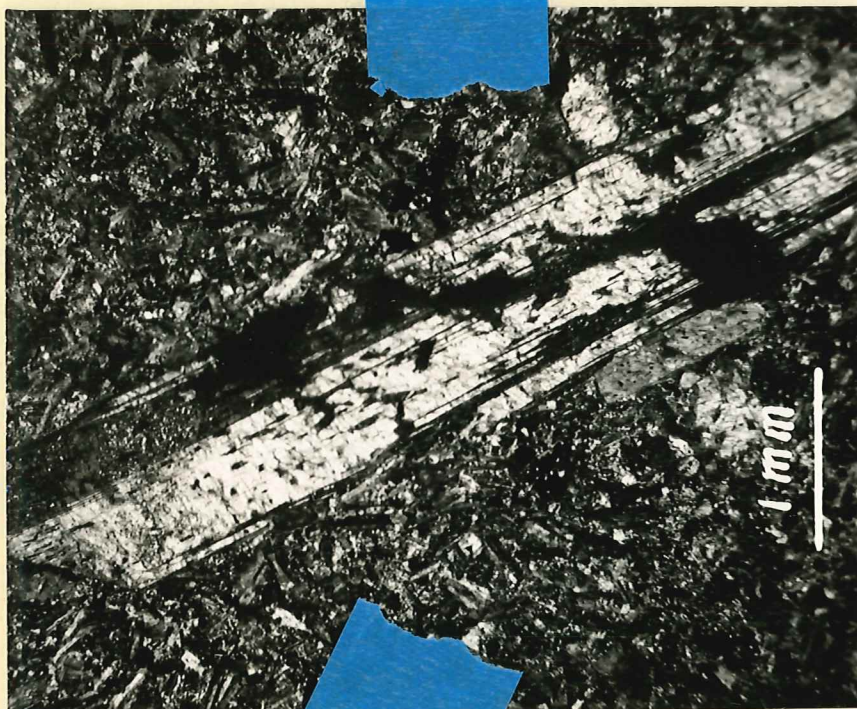


Figure 18. Photomicrograph of plagioclase phenocryst (side view) in "Oatmeal" basalt. The mottled appearance is due to alteration to more albitic composition. The albitic material is white and obscures albite twinning. Note orientation of inclusions. Crossed nicols.



irregular knobby amygdules the size of a small potato. Although rudely elliptical shapes are common, many are very irregular in form. The amount of amygdaloidal material varies from a small per cent in porphyritic rocks, to as much as one-half of the rock. Amounts of 10 to 20 per cent are common at the top of porphyritic flows.

The amygdules are vesicle fillings of chalcedonic quartz and/or a carbonate mineral, along with minor amounts of hematite, chlorite and epidote. Where chalcedonic quartz is the dominant mineral in strongly amygdaloidal rocks, the amygdaloidal horizon comprises a resistant zone which holds dip slopes.

Analysis 2 Table 5 is representative of the chemical composition of an amygdaloidal basalt for the Zosell district.

Distinctive mineral coatings in some amygdaloidal flows might serve as stratigraphic markers. In the lower half of the basalt sequence in the Elliston district, a dark bluish-green mineral coating of amygdules was noted at several localities. Similar mineralogical features have been observed in vesicles in lavas of the Columbia plateau. Distinctive blue or green coatings are characteristic of separate lava flows. (Mackin, personal communication.)

Analysis 2 Table 5 is representative of the chemical composition of an amygdaloidal basalt for the Zosell district.

#### Cliff Mountain Basalts.

At Cliff Mountain, there are two basaltic flows in the upper part of the section, the lower of the two (unit No. 6, Table 5) has a dense

aphanitic base and top, but the middle part is fine grained and contains small, gray, calcic plagioclase phenocrysts in a dark greenish-gray groundmass. Labradorite phenocrysts are up to 2 mm in length, are internally fractured, strongly resorbed, and partly altered to clinozoisite and clay minerals. Smaller plagioclase crystals up to 0.5 mm long comprise about 25 per cent of the rock. The mafic minerals are completely altered to chlorite, some of which exhibit complex spherulitic clusters.

The basalt at the top of the mountain (unit No. 7, Table 5) is a flow 108 ft thick, the crust of which solidified, but the mobile lower part continued to move as lava under the crust. The formerly fluid part has a very uneven upper surface marked by intrusive projections into the top of the unit. The rock weathers to dark reddish gray in which light gray feldspar phenocrysts and dark greenish clusters of mafic minerals occur.

The dense upper crust of the flow is indistinctly microlitic and contains small labradorite phenocrysts to 0.1 mm. The mafic minerals are altered to chlorite and opaque iron oxide. The base of the unit has an almost glassy groundmass containing plagioclase microlites and small phenocrysts of andesine and mafic minerals which are strongly altered to sericite and chlorite respectively.

Basalt Breccia. At Sugar Loaf Mountain, west of Cliff Mountain, (Cliff Mountain section No. 14, Table 5) there is a conspicuous basalt flow breccia at the base of the Emery basalt section which is not recognized in the Elliston district. The unit is quite thick, probably



300 to 400 ft, as the entire knob of Sugar Loaf Mountain is made up of the breccia. Part of the breccia is composed of a mass of angular fragments, part has merely a crackled appearance. The fine-grained brecciated matrix weathers conspicuously on the outcrop, but on fresh surfaces, the brecciated character is indistinct. The rock weathers to a dull brownish-gray color on the outcrop, but on fresh exposures is a dark bluish gray. The groundmass is very fine grained and practically devoid of visible phenocrysts.

In thin section, the rock shows small reasonably fresh crystals of labradorite, augite and pigeonite, as well as another mafic mineral which is strongly altered. A small amount of altered amphibole is also present, which has been partly changed to chlorite. Plagioclase makes up from 60 to 70 per cent of the rock, however, the largest individual crystals are less than 0.5 mm in length. Magnetite, apatite, and a trace of quartz comprise the accessory minerals.

#### Chemical Composition of Basalts.

Although there are no chemical analyses of basalts from the Elliston district, there are several analyses of basaltic rocks from the near-by Zosell district. Analyses 1, 2, 3, and 5 in Table 8 are representative of specific basaltic types found in the western part of the Elliston district. Analysis 4 is representative of an andesite "near basalt composition" according to Weed (1901, p. 529). According to the Johannsen system, it would be classified basalt.

## Volcanic-Derived Sediments

Sediments of volcanic derivation comprise only a small part of the volcanic pile in this district. On the east side of Mike Renig Gulch, about 5 mi south of MacDonald Pass, hornfelsed sedimentary tuffs of sand and fine gravel size crop out at one locality.

In the Cliff Mountain section, two sedimentary units were measured. One is a water-laid tuff of andesitic composition (Unit No. 4, Table 5). The rock is a fine-grained, greenish-gray tuff which weathers to irregular subconchoidal masses. The unit is variable in thickness over the outcrop area on the face of the cliff. Along the measured section it is 38 ft thick. Small limey concretions were observed near the base, and some dark, flinty material was noted which maybe silicified wood. In thin section, the rock is seen to be a fine-grained lithic tuff with a large amount of carbonate in the groundmass. Small andesine crystals, about 1 mm long, and strongly altered sericite, comprise 10 per cent or more of the rock.

The other unit (No. 3, Table 5) is of uncertain origin. It is a mottled, dull, reddish-purple and greenish breccia, about 190 ft thick. Large cobbles appear to be somewhat rounded and are concentrated in crude beds. In the first set of field notes, the phrase "possible sediment" appears. In company with Dr. Klepper in the field at a later date, many rounded boulders, up to 1 ft in diameter, were noted imbedded in a tuff matrix. The unit may represent debris from an avalanche.



## Age and Correlation of Older Volcanics

The breccias at the base of the volcanic pile lie on Colorado shale and older Lower Cretaceous Rocks in the western part of the Elliston district. About 10 mi west, the basalts rest directly on Upper Cretaceous sediments more than 1500 ft above the base of the Colorado.

The basal breccias, the andesites and the latites of the volcanic pile, were intruded by quartz monzonite of the Boulder batholith. At no locality have batholithic rocks been observed to be intrusive into the basalts. Because there is no obvious unconformity or structural discontinuity between the andesitic-latitic series and the basalts, and because basalt flows are interlayered in the middle part of the volcanic section, the basalts have been considered part of the pre-batholithic volcanics into which the quartz monzonite was intruded.

On the last day of work in the 1955 field season, Dr. Klepper found several granitic xenoliths in one of the basalt flows about 1 mi west of the Elliston district. Some of the boulders resemble alaskitic rocks of the batholith. Figure 19 shows the texture of one of the boulders which is very similar to the texture of some alaskite in the Elliston district. (See section on aplite-alaskite.) Another boulder found in basalt at the same locality is a breccia which is similar petrographically to a rheomorphic breccia from Cornucopia, Oregon (Goodspeed 1953). The texture is shown in Figure 20. No counterpart of this rock is known in the region.





Figure 19. Photomicrograph showing texture of leucocratic granitic inclusion in basalt. Note small fracture filled with quartz. Crossed nicols.

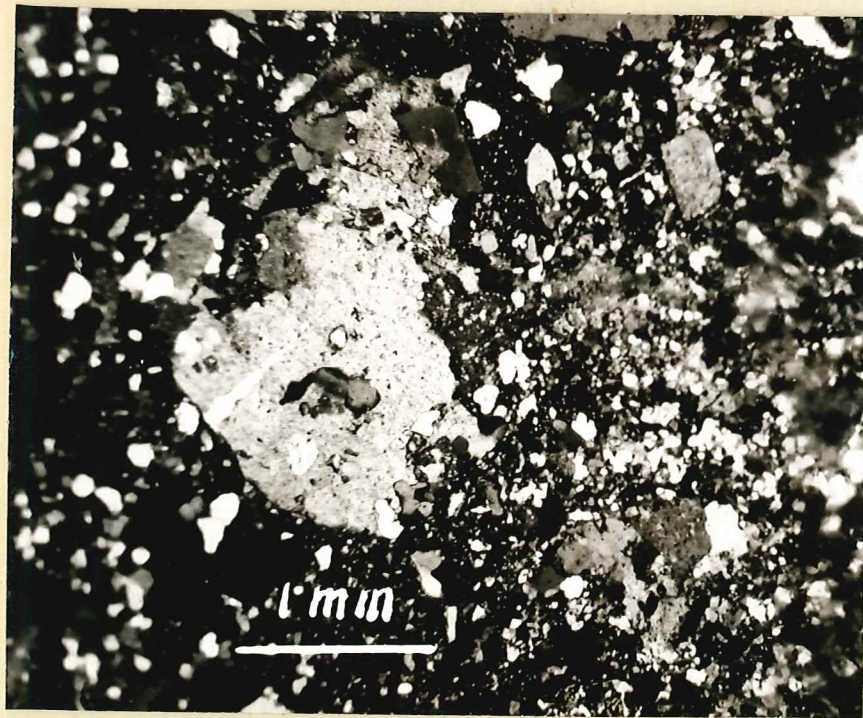


Figure 20. Photomicrograph showing cataclastic texture in inclusion in basalt. The hazy line at the right marks the boundary between a red colored breccia fragment and the dark cataclastic groundmass. Crossed nicols.



Two or three granitic boulders were removed from the basalt in an attempt to collect enough rock for a zircon age determination. An age of 50 million years was determined by the U. S. Geological Survey (Klepper, personal communication). This determination may be in error by as much as 100 per cent, because zircon is broken down in contact zones and in rocks subjected to metamorphism. The limits of error bracket the Mesozoic and early Tertiary granitic rocks of the Boulder, Philipsburg and Idaho batholiths (see Table 15). The granitic boulders are thus not derived from the pre-Beltian metamorphic rocks. The age of the metamorphic breccia xenolith is unknown. It may have been picked up in the pre-Beltian basement rocks.

If the granitic boulders were incorporated into the basalt flow as it moved across a boulder-strewn surface of batholithic rocks, a post-batholithic age for the basalt would, of course, be indicated. Although this explanation for the granitic boulders is one of the possibilities which must be considered, it does not solve the problem of the metamorphic boulder which, so far as is known, has no possible counterpart in the rocks exposed in the region.

If the granitic boulders were picked up en route to the surface, the age of the boulders is indeterminate relative to the age of the exposed batholithic rocks.

The discovery of these boulders has cast some doubt regarding the age of the basalts. However, the interlayering of similar basalts in pre-batholithic andesitic and latitic volcanics, the absence of any obvious structural discontinuity between the latitic and basaltic parts

Table 8

## Chemical Analyses of Basalts

	(1)	(2)	(3)	(4)	(5)
SiO <sub>2</sub>	47.4	50.0	53.0	54.50	54.61
Al <sub>2</sub> O <sub>3</sub>	13.8	14.1	14.7	17.97	15.23
Fe <sub>2</sub> O <sub>3</sub>	5.1	3.4	4.0	8.55	3.51
FeO	5.3	4.7	4.8	nd.	4.80
MgO	7.1	4.0	4.7	3.37	4.69
CaO	10.7	6.9	8.2	9.37	7.41
Na <sub>2</sub> O	2.8	2.6	2.3	2.34	1.46
K <sub>2</sub> O	1.4	2.3	1.4	1.62	2.70
H <sub>2</sub> O <sup>+</sup>				.96	2.47
Loss	2.8	8.9	5.3		
H <sub>2</sub> O <sup>-</sup>				.14	.32
TiO <sub>2</sub>	1.2	.58	.62	.94	.86
P <sub>2</sub> O <sub>5</sub>				.31	.35
MnO				tr.	.09
Co <sub>2</sub>					1.46
S				.63	none
CO <sub>2</sub>					
BaO				.06	.11
SrO				.05	.04
Li <sub>2</sub> O				tr.	
Totals	97.6	97.5	99.0	100.80	100.11



Table 8

Chemical Analyses of Basalts

1. Pyroxene basalt, Helena Gulch, Zosell District. Montana Bur. Mines and Geology Mem. 34, p. 7, 1953.
2. Amygdaloidal basalt. Footwall Bonanza lode, Montana Bur. Mines and Geology Mem. 34, p. 7, 1953.
3. Basalt porphyry, pyroxene and plagioclase phenocrysts, near Emery Mine, Zosell district. Montana Bur. Mines and Geology Mem. 34, p. 7, 1953.
4. Hornblende andesite porphyry altered from augite andesite approaching basalt, Elkhorn district, U. S. Geological Survey AR 22 II, p. 529, 1901.
5. Andesite porphyry, Daylight Mine, Zosell District. U. S. Geological Survey Bull. 168, p. 119, 1900.

of the volcanic section, and the petrographic similarities of the "oatmeal" basalt in the Elliston district to pre-batholithic "oatmeal" basalt in the Whitehall area, are considered evidence favoring pre-batholithic age of the basalts in the Elliston district.

The basalts may be related to the granogabbro sills which mark the initial stage in the emplacement of the batholith. On the other hand, the possibility exists that they may be, at least in part, post-batholithic in age. Detailed mapping of the basalts in the area, and particularly in adjoining areas, will be necessary before the age of the basalts can be settled.

#### Upper Cretaceous Volcanics in Adjoining Regions.

Upper Cretaceous volcanic rocks are found at a number of localities in the region of the Boulder batholith and along the front of the Rocky Mountains in western Montana. They include the Livingston, Three Forks, Whitehall, Elkhorn Mountain, Radersburg-Winston and Wolf Creek areas.

In the Livingston area, about 125 mi east of Butte, the Livingston formation was first described by Weed (1893). The volcanics and volcanic-derived sediments which comprise the formation are in the lower Montana group in the upper part of the Eagle formation (Vhay 1939). Augite andesite, hornblende andesite, dacite and basalt comprise the principal rock types found in the formation (Parsons 1942).

In the Three Forks area, andesitic volcanics which rest on the Colorado shales were mentioned by Berry (1942). I have seen and mapped volcanics in the area and have observed that they are dominantly



andesitic in composition and comprise tuffs, lapilli tuffs, breccias, agglomerates and sediments of volcanic origin.

While conducting a class in geologic mapping in the South Boulder Creek section southeast of Whitehall, more than 2,000 ft of volcanics were mapped. They rest on Colorado shales and were strongly folded and thrust faulted prior to the emplacement of the batholith. The rocks in the section are andesite tuffs, basalt flows, volcanic sediments, agglomerates, and particularly distinctive "oatmeal" basalt near the top of the section. It has the same petrographic characters as the "oatmeal" basalt in the Elliston district. Some of the basalts are amygdaloidal and have features which are very similar to the Emery basalts found in the Elliston district.

On the northeast side of the Boulder batholith in the Radersburg-Winston districts, volcanic sediments overlies upper Colorado group formations. Volcanic-derived sediments, known as the Slim Sam formation, (Klepper and Freeman 1951), are overlain by the Elkhorn volcanics (Klepper 1951). The Slim Sam formation consists of about 900 ft of sedimentary tuffs, lapilli tuffs, volcanic sandstones and mudstones. The lower part of the formation contains thinly-bedded sandstone in which quartz is dominant over feldspar. The sediments grade upward into massive sandstones in which quartz comprises less than five per cent of the rock. Thin, black, argillite interbeds are common throughout the section. The upper part of the formation grades into tuffs and volcanics of the Elkhorn Mountain volcanics, which comprise a series of

massive crystal tuffs, tuff breccias, and welded tuffs 2,000 ft or more thick.

In the southern Elkhorn Mountains, Klepper (1951) measured a section of Upper Cretaceous volcanics which is summarized in Table 9.

Table 9

## Elkhorn Mountains Volcanics

Andesite flows, dark, amygdaloidal, some red jasper in amygdules, amygdaloidal basalt interbedded	630 feet
Andesite and latite (?) breccia, tuff-breccia lapilli tuff, tuff	1255 "
Andesite and latite (?) flows with interbedded tuff breccias, lapilli tuff, and breccias. Coarse pyroclastic flows near base.	<u>1035 "</u>
	2920 feet

The overall similarity of the Elkhorn Mountain section and the Elliston section is striking. Apparently the Elliston volcanics are somewhat more basic, but in other particulars, namely the presence of coarse pyroclastic flows at the base, andesitic and latitic tuffs in the middle, and andesitic flows with some amygdaloidal basalts at the top, the two sections are similar. The Elkhorn Mountain volcanics are older than the early intrusives related to the Boulder batholith found in that area.

In the Wolf Creek area south of Great Falls, the Adel Mountains volcanics rest unconformably on the Two Medicine formation of Upper Cretaceous age at about 2700 ft above the base of the Colorado (Lyons 1944, p. 449). Of special interest is the recognition of glassy welded tuffs of possible latitic composition (Barksdale 1951, p. 440).



Conclusions. The andesitic and latitic volcanics in the Elliston district overlies lower Upper Cretaceous sediments and are older than the intrusion of the batholith. Because of their petrographic similarities to other Upper Cretaceous volcanic rocks, the Elliston prebatholithic volcanics are considered Upper Cretaceous in age. They appear to be more or less equivalent to the Upper Cretaceous Elkhorn Mountain volcanics, the Livingston volcanics, the volcanics in the Whitehall area, and to the Adel Mountain volcanics. The basalts at the top of the volcanics are likewise considered to be Upper Cretaceous because:

1. There is no obvious unconformity or obvious structural discontinuity recognized at the base of the basalts;
2. Basalt flows of about the same composition are found inter-layered in the andesitic and latitic units;
3. The uppermost basalts in the Elliston and Elkhorn districts are very similar petrographically and chemically and if correlative in age the Emery basalts must be pre-batholithic;
4. If the "oatmeal" basalt in the Elliston district is correlated with the "oatmeal" basalt in the Whitehall area, then the Emery basalts are pre-batholithic in age.

#### YOUNGER VOLCANIC ROCKS

##### Rhyolite

Rhyolite ignimbrites and possibly flows are found; (1) in the southern and southwestern part of the district over an area of about 8 sq mi; (2) in isolated remnants on the Continental Divide; (3) in

somewhat larger remnants in the western part of the area; and (4) abundantly distributed over the adjoining area to the west. Within the district, two types of rhyolite have been recognized. The older is completely devitrified and contains minor amounts of altered biotite, whereas the upper unit is distinctly a welded tuff which contains virtually no biotite or magnetite, and is found over a wide area not only within the district, but in the region south and west of the district.

#### Older Rhyolite.

The older rhyolite crops out in the valley on the west side of Bison Mountain. The areal extent is not known. The rhyolite is completely devitrified and includes small phenocrysts of sanidine, oligoclase, quartz and minor amounts of mafic minerals which have been altered to chlorite and epidote. The groundmass is devitrified to micro-spherulitic and complex micropoikilitic grains of feldspar and quartz as shown in Figure 21.

#### Younger Rhyolite.

The younger rhyolite is the more widespread unit, and it is clearly a welded tuff. Mineralogically it is a quartz latite with sanidine, quartz and oligoclase phenocrysts. Biotite and magnetite are conspicuously absent. The base of the rhyolite is commonly obsidian vitrophyre which grades sharply upward into a dense tuff which shows strong layering due to compaction of welded shards. Mirolitic cavities



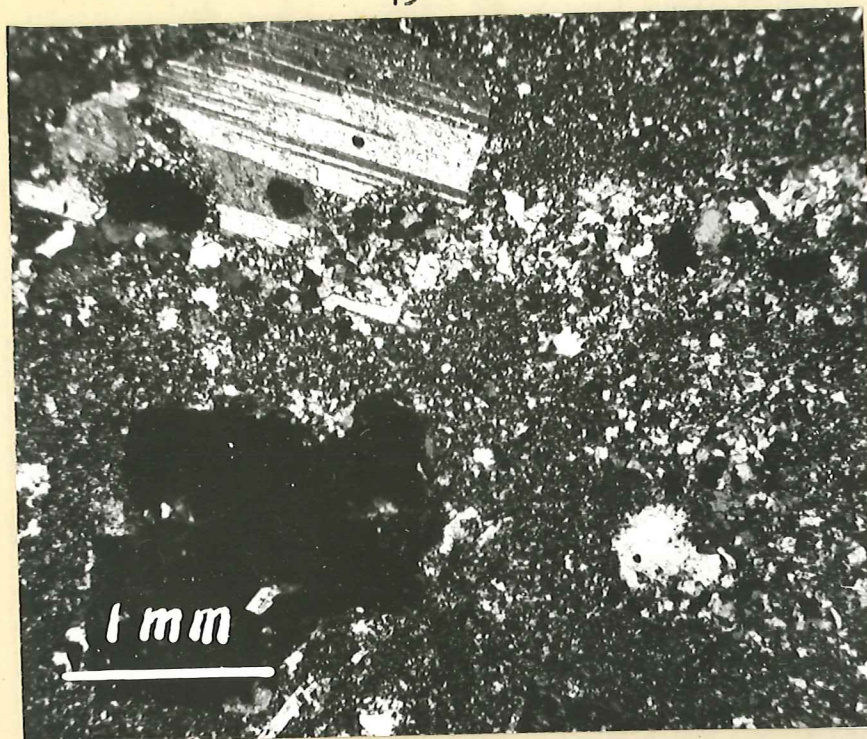


Figure 21. Photomicrograph of plagioclase phenocryst in devitrified rhyolite. The microgranular layer is a flow-like layer. Crossed nicols.

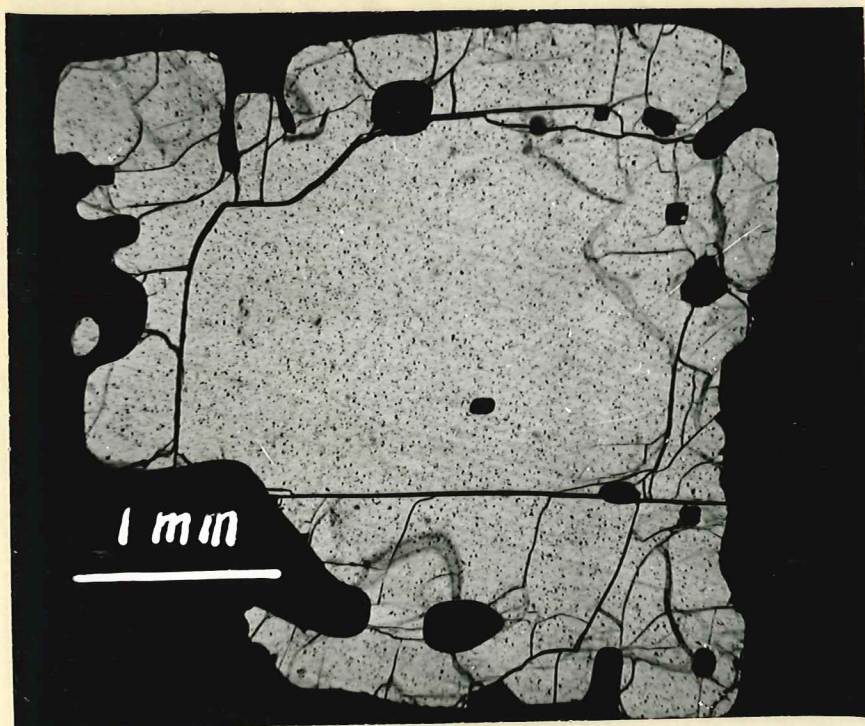


Figure 22. Photomicrograph of partly resorbed sanidine phenocryst in obsidian vitrophyre. Note cracking at margin of grain. Similar curved cracks noted also in quartz. Crossed nicols.



abound in the less conspicuously welded upper part of the rhyolite. Small, white cristobalite crystals were identified in miarolitic cavities at a locality 1 mi west of Elliston.

About 0.5 mi west of Elliston, there is obsidian vitrophyre at the base of the rhyolite. Streaks of dark reddish obsidian are in the black glass. Clear, glassy, yellowish sanidine phenocrysts are abundant but are partly corroded and all have a distinctive internal fracture pattern shown in Figure 22.

Just a few rods south of the southern boundary of the district in NW  $\frac{1}{4}$  sec. 34, T. 8 N., R. 6 W., an obsidian was found which is somewhat vitrophyric and contains scattered, resorbed sanidine phenocrysts. The glass is microperlitic, as shown in Figure 23. The indications of intense welding are more distinct in the photograph than was suspected from an examination of the thin section. Mansfield and Ross (1935) report a thin, completely welded, perlitic obsidian at the base of a welded rhyolite in Idaho. Elsewhere the basal obsidian shows signs of welding and subsequent fracturing as shown in Figure 24. The rhyolite above the glassy base is a high-grade welded tuff with densely compacted and squeezed shards, which gives way to a dense, somewhat miarolitic, but strongly banded rock above. A thin, light gray, sandy tuff was noted in NE  $\frac{1}{4}$  sec. 32. It may represent a non-welded upper part of a welded rhyolite unit, or comprise a thin pyroclastic tuff between two rhyolites.

The textures in the rhyolite are very similar to textures illustrated by Mansfield and Ross (1935) to represent welding and compaction of glass shards in a welded tuff.



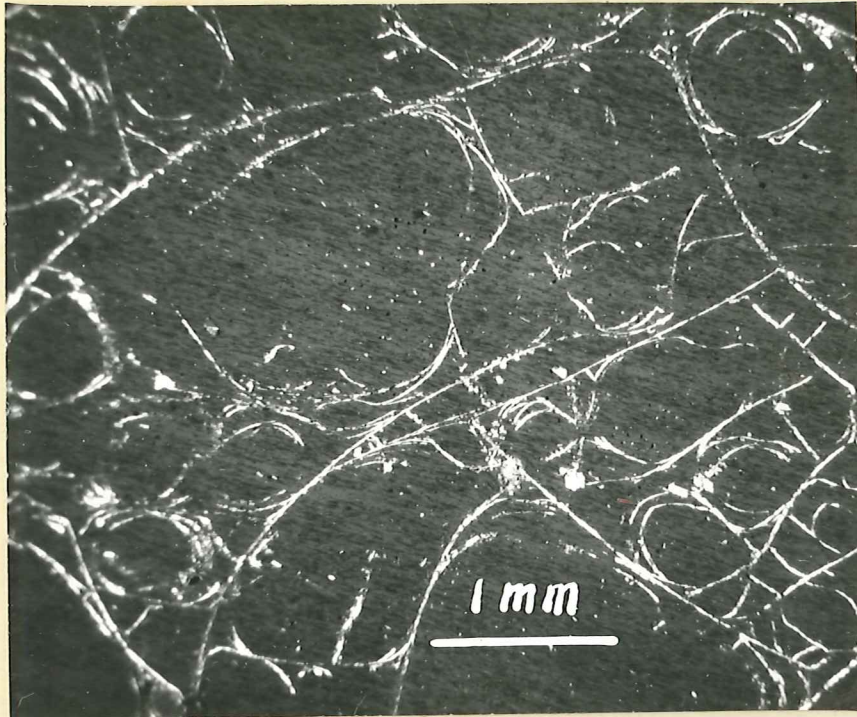


Figure 23. Photomicrograph of perlitic fractures in glass. Compacted structures inconspicuous, nevertheless clearly visible. Crossed nicols.



Figure 24. Photomicrograph of welded and microbrecciated glass at base of welded rhyolite tuff. Plane polarized light.



Previous Work on Welded Tuffs of Rhyolitic Composition.

The welded nature of some rhyolite tuffs was early recognized by Iddings (1899) and, according to Mansfield and Ross (1935), even earlier by W. Von Fritch and W. Reiss in 1868 in their descriptions of rhyolites from the Canary Islands. A great many observations have been made which have accumulated to demonstrate the probable welded character for many rhyolites. Some of the examples of acidic volcanics which are clearly welded tuffs are described by Grange (1934, 1937) and Marshall (1932, 1925) in New Zealand; Green (1915, 1919), Marr (1916), Mitchell (1929, 1934), Oliver (1954), and Hartley (1942) in England; Richards and Bryan (1934) in Queensland; Fenner (1948) and Jenkes and Goldich (1954) in Peru; and Iddings (1899), Mansfield and Ross (1935), Gilbert (1938), Williams (1942), Mackin and Nelson (1950), Moore (1934), Barksdale (1951), Staples (1950), and Wilkinson (1950) in the United States.

The recognition of the great lateral extent of some welded tuffs and their uniform thickness (where the lower contact is nearly flat) is a more recent observation. J. H. Mackin, H. W. Wheeler and Paul Williams have successfully traced individual rhyolites over broad areas in the Great Basin. Their stratigraphic significance is becoming more apparent. (Personal communications.)

Any rhyolite, latite, andesite (or even basalt ?) which exhibits stratigraphic continuity over a wide area, which is reasonably uniform in thickness, and which exhibits mineralogical and textural identities over a wide area, is suspect of an ignimbrite or welded tuff. Actually,



the first field work done by the writer in the pre-Cambrian igneous rocks in Missouri involved the tracing of rhyolites which are uniform in thickness over a considerable area. They are almost certainly welded tuffs.

The natural phenomenon which produces the welded tuffs which are spread over many hundreds and even thousands of square miles, calls for a remarkable mechanism which permits the extrusion of vast quantities of materials in a medium which has the fluidity of water, or something even more fluid. Blasts of gas-solid phase matter at high temperature, producing hot, high-density clouds, have been observed, on a relatively small scale, in the spectacular eruptions on the Island of Martinique, and described by Lacroix (1904) and Perret (1935). A similar mode of origin was postulated by Fenner (1937) for the tuffs at the Valley of 10,000 Smokes. From purely metallurgical reasoning, Reynolds (1955) applied the concept of fluidization to the mechanism of the ignimbrites. The idea has a high degree of reasonableness, however, the origin of the conditions which gives rise to the ignimbrites is unsolved.

#### Chemical Composition.

No chemical analyses are available for rhyolite in the Elliston district, however, an analysis is available from the adjacent Rimini district. Two additional analyses from the Butte district are included in Table 10.

Table 10

## Chemical Analyses of Rhyolite-Quartz Latite

	1	2	3
SiO <sub>2</sub>	75.30	74.34	67.55
Al <sub>2</sub> O <sub>3</sub>	11.95	12.97	15.68
Fe <sub>2</sub> O <sub>3</sub>	2.17	.75	.98
FeO		.54	1.02
MgO	.05	.86	1.11
CaO	.62	.85	2.51
Na <sub>2</sub> O	3.09	2.49	4.15
K <sub>2</sub> O	4.96	4.72	2.86
H <sub>2</sub> O <sup>+</sup>	.61	1.11	2.76
H <sub>2</sub> O <sup>-</sup>	.36	1.03	.38
TiO <sub>2</sub>	.17	.18	.34
P <sub>2</sub> O <sub>5</sub>	tr	.07	.12
MnO	tr	tr	tr
SO <sub>3</sub>	.44	.03	
Cl	tr		.05
BaO	tr	.07	.11
SrO	tr	tr	.03
Organic	.45		
Li <sub>2</sub> O		tr	
ZrO <sub>2</sub>	—	—	—
Total	100.17	100.06	99.63



## Table 10 cont.

## Chemical Analyses of Rhyolite and Quartz Latite

1. Rhyolite, Rimini Mining District, U. S. Geological Survey  
Bull. 168, p. 119, 1900
2. Rhyolite, Butte, Montana, U. S. Geological Survey Prof-  
essional Paper 74, p. 44, 1912.
3. Rhyolite-Dacite obsidian, Big Butte, Butte, Montana,  
U. S. Geological Survey Professional Paper 74, p. 45, 1912.

### Age of the Rhyolites.

The rhyolites which crop out in the district are clearly younger than the granitic rocks because they rest unconformably on batholithic rocks. Some rhyolite just west of the district has been involved in faulting, but for the most part, the youngest rhyolites obscure the lineament trends in the southern part of the district, and overlies a strong north-south fault zone. West of Elliston a body of rhyolite gives the impression that it flowed down a valley in relatively modern (pre-glacial) time.

In the Butte district, rhyolite is clearly intrusive into and extrusive upon the quartz monzonite at Big Butte. Weed (1912) considered that Big Butte is a remnant of a vent as there are extrusive flows south of the butte, and many dikes appear to swarm out to the south from the center of the butte. The rhyolite has been encountered in the mine workings underground and is clearly post-mineralization. Most of these so-called rhyolites are, however, dacites. (Sales 1914).

In the Rimini district, individual flows and masses of extrusive rhyolite flows are reported by Ruppel (personal communication) including the body of rhyolite at Red Mountain described briefly by Knopf (1913, p. 39).

Knopf (1913, p. 41) considered the rhyolite at Mullan Pass to be possibly upper Miocene or younger in age. Pardee and Schrader (1933, p. 21) consider the rhyolite near Avon (West of Elliston) to be pre-Oligocene on the basis of Tertiary "lake beds" which contain Oligocene vertebrate remains. (Douglas 1902.)



The youngest rhyolite in the Elliston district, which is obviously a welded tuff, is found in isolated patches over an area of nearly 100 sq mi. To this may be added the apparently identical rhyolite found just west of Elliston, and those found 10 mi farther west near Avon. Possibly the rhyolites found to the south, northwest of Butte, and the similar rhyolites observed 80 mi southwest of Butte at Pipe Organ Lodge are all related.

Several of the hand specimens of the younger rhyolite from the Elliston district were examined by Dr. Harry Wheeler, Mr. Paul Williams and Mr. Dwight Schmidt. Wheeler (personal communication) considered the younger rhyolite to be strikingly similar to the Rattlesnake rhyolites of Pliocene age found in Idaho which are described by Wilkinson (1950). Williams (personal communication) considered that they could be Pliocene welded tuff units, based on mineralogical and lithological characters of the units he has worked with in Nevada and Idaho, where the Miocene volcanics are more intensely welded and have less biotite than the Oligo-Miocene welded tuffs which contain biotite. Schmidt (personal communication) likewise identified the Elliston rhyolite as resembling upper rhyolite series, probably Miocene, in south central Idaho. The agreement of these individuals in "correlating" hand specimens collected in Montana with specimens of known age in Idaho and Nevada, seems quite remarkable to me. Individual samples collected by Williams from near Curry, Nevada, were indistinguishable from samples in the Elliston district. This was true for two distinctly differently textured rocks, neither of which was an obsidian vitrophyre. In thin

section, rhyolites from the two areas were, in individual instances, virtually identical.

Although lithologic and petrographic correlation are admittedly hypothetical, nevertheless, as welded tuffs do have broad regional stratigraphic continuity, the possibility should be considered that the uppermost rhyolite unit in the Elliston district may be Pliocene in age and may not have been derived locally from the area of the Boulder batholith.

Thus within the area of the Boulder batholith there may be more than one age of rhyolite. Some are clearly related to the batholith, at least geographically and presumably petrogenetically. (See section on chemical petrology.) Among these are the rhyolite and dacite bodies found at Butte, near Boulder, and in the Corbin-Wickes district. Possibly the older rhyolite in the Elliston district is of local derivation, and possibly the younger rhyolite as well. However, the possibility exists that the younger rhyolite is Pliocene in age.



INTRUSIVE ROCKS

## Introductory Statement

The intrusive rocks comprise a wide variety of types grading from granogabbro to quartz monzonite to aplite-alaskite in a succession of stages in the emplacement of and crystallization of the Boulder batholith. Thin sills are found in the sedimentary and volcanic rocks, and a few dikes cut all rocks in the region except rhyolite. The rocks in sills and dikes are basalt, rhyolite (just west of district), diorite and quartz monzonite-porphyr<sup>y</sup>\*. The intrusive rocks are described in chronological order.

The batholith crops out in the eastern and southern part of the district over an area of about 20 sq mi, but within this area two intrusive stages are found. A granodiorite, herein called the Bison Mountain granodiorite, and normal or main-stage quartz monzonite, with its related aplite and alaskite, are exposed. Two earlier intrusive stages of the batholith, namely granogabbro and basic granodiorite, are not found within the district, but are found in adjoining areas.

---

\*Note: "Porphyry" is used for the name of a rock, such as quartz monzonite-porphyr<sup>y</sup>, in the same sense that Johannsen used the term. He says: "Granite-porphyr<sup>y</sup>ies must not be confused with porphyritic granites. The latter are definitely granites with a porphyritic texture, but the average grain is medium-sized and not aphanitic. Granite-porphyr<sup>y</sup>ies have groundmasses which are granular under the microscope, but to the unaided eye are rather dense; and the phenocrysts stand out quite prominently, even though they may not be, in every case, very abundant. Usually, these phenocrysts are feldspar, quartz, and the dark minerals; and of these, the latter may not reappear in the second generation in the groundmass. . . The phenocrysts are usually rather automorphic, sometimes corroded. The groundmass is nearly xenomorphic-granular, usually medium-grained, but in some cases almost, but not quite, cryptocrystalline." (Johannsen 1949, p. 294).

A thick granogabbro sill is found west of the district in an area that is not scheduled to be mapped in the near future. Because it is the only granogabbro known in the northwestern region of the batholith, a short description of the rock is included.

The first part of this section on intrusive rocks deals with the descriptive petrography of the rocks grouped into their broad petrologic types. The mineralogy of quartz monzonite is discussed in detail, particularly the alkali feldspars because of their significance in late stage magmatic processes leading to mineralization. The textural types are described from typical localities. The next part deals with the petrogenesis of quartz monzonite and related aplite-alaskite in the magmatic, late magmatic or deuteritic and hydrothermal stages of magmatic crystallization.

#### Granogabbro

Although gabbro is not exposed within the district, a 400-ft sill, intrusive into Upper Cretaceous beds about 1500 ft above the base of the Colorado formation, is exposed on Hoover Creek about 12 mi west of Elliston. (Pardee 1917, p. 212). Similar rocks crop out in the northeastern part of the batholith (Pardee and Schrader 1933, p. 18) and in the Elkhorn region on the east side of the batholith (Weed 1901, p. 426)

The rock weathers to a brownish feldspar and altered mafic mineral grush (Johannsen 1916) and rarely may fresh rock be observed on outcrops. In a highway cut on U. S. 10N about 14 mi west of Elliston, fresh rock is medium grained, dark greenish gray, and contain gray plagioclase crystals and greenish black pyroxenes. The rock has an altered



appearance due to late deuteric action. Small leucocratic stringers, consisting chiefly of alkali feldspar, occur in fractures in the rock.

In thin section, the minerals identified are: sodic labradorite (Ab 44 An 56) which occurs in automorphic to hypautomorphic crystals 1 to 3 mm long, some of which are strongly fractured internally; augite, which is partly altered to uralite and very fine-grained iron oxide; olivine which is very minor in amount and almost completely altered; and accessory biotite, potash feldspar, quartz, magnetite, and apatite. The biotite is strongly pleochroic,  $x = \text{tan}$ ,  $y = z = \text{dark reddish brown}$ ,  $2V = 0$ . Potash feldspar comprises possibly 10 per cent of the rock, is optically negative, and has a small optic angle,  $2V$  about  $30^\circ$  to  $40^\circ$ . The potash feldspar crystallized in part between plagioclase crystals, and is in part formed by replacement of the early plagioclase as shown in Figure 25. The mineral is tentatively identified as anorthoclase. (See section on mineralogy under quartz monzonite.) In virtually all places where potash feldspar is in replacement contact with plagioclase, there is a narrow albitic feldspar rim between potash feldspar and plagioclase.

The texture of the rock is hypautomorphic diabasic in which plagioclase comprises about 50 per cent, pyroxene (and uralite) 25 per cent, potash feldspar 10 per cent, biotite 5 per cent, quartz 5 per cent, and magnetite 5 per cent. Figure 26 illustrates the diabasic texture. The rock is a granogabbro (237 Johannsen\*).

---

\*Note: The Johannsen classification is used for igneous rocks except that quartz monzonite is used instead of adamellite.

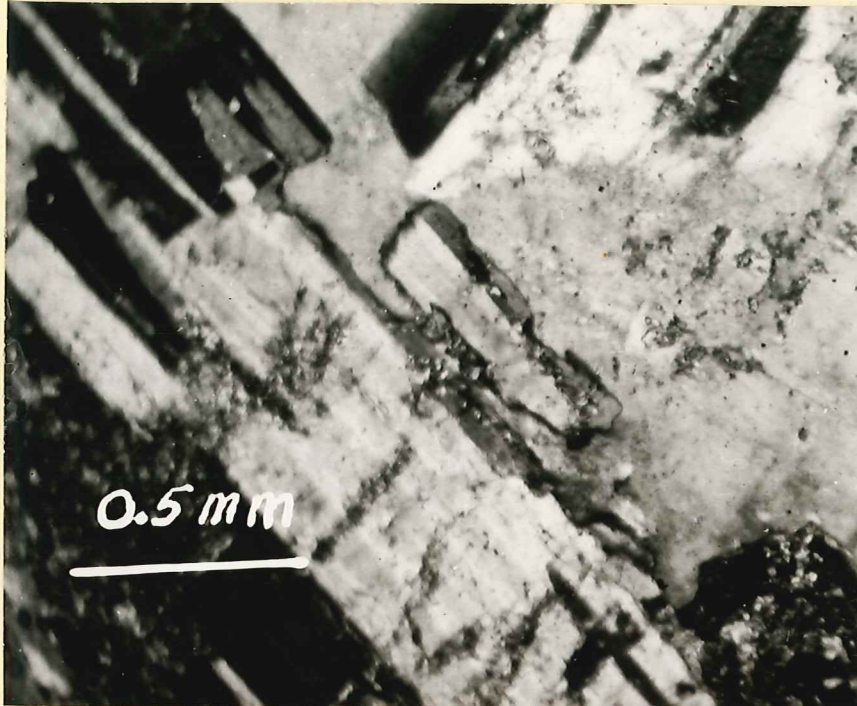


Figure 25. Photomicrograph of potash feldspar (anorthoclase ?) which has partly replaced plagioclase in granogabbro. Crossed nicols.



Figure 26. Photomicrograph of diabasic texture in granogabbro. Crossed nicols.



### Bison Mountain Granodiorite

Granodiorite crops out on Bison Mountain and adjoining areas in the southwestern part of the district, as shown on the geologic map. It is a distinctive granitic type characterized by: (1) a grayish-green color; (2) epidote coatings and iron oxide staining on joints; and (3) a greater proportion of plagioclase than found in normal quartz monzonite.

The granodiorite is not found in contact with main stage quartz monzonite, consequently its relative age to the quartz monzonite is not known. Because it is a granodiorite and because granodiorite usually precedes quartz monzonite in the emplacement of the batholith (Klepper, personal communication), the Bison Mountain granodiorite is considered to be older than quartz monzonite, however it may well be very closely related to the main stage of quartz monzonite.

There are three main textural facies within the granodiorite body. The normal granodiorite is a fine medium-grained, greenish-gray rock. In the coarser-grained facies, the rock has a distinct pinkish color, due to well-developed potash feldspar. Fine-grained diorite porphyry is formed at the north end of the intrusive where it lenses out into a sill-like body.

In its coarse-grained facies, the granodiorite is light pinkish gray, in which plagioclase and potash feldspars are both distinguishable, the latter usually in smaller, and less conspicuous, grains except where porphyroblasts formed. Relatively large potash feldspar crystals which form in situ after the rock is mostly crystallized are called porphyroblasts to distinguish from phenocrysts which crystallized from

silicate magma. The feldspar crystals range from 1 to 3 mm, rarely larger. Hornblende is the common mafic mineral, and it occurs in elongated crystals which have very ragged borders. Irregular grains of biotite and quartz are rarely distinguishable in the hand specimen.

From near the south end of Bison Mountain at a massive outcrop of the coarsest-grained granodiorite, the modal analysis of the rock is about as follows:

Plagioclase	45% (Ab 55 An 45)
Potash feldspar	30
Quartz	10
Pyroxene and uralite	10
Biotite	5
Magnetite and sphene	5

Granodiorite (227 Johannsen)

Andesine occurs in crystals up to 5 mm in length, some of which are zoned, internally fractured and altered to clinozoisite and sericite along the fractures. On the other hand, some of the plagioclase is remarkably fresh and unaltered, even where the grains are strongly fractured. Augite is in part fresh, but most commonly it is almost completely altered to uralite (actinolitic hornblende) with minor amounts of biotite and iron oxide. Potash feldspar and quartz are largely confined to the xenomorphic groundmass composed of grains ranging from very small to 0.5 mm in diameter. Many incipient porphyroblastic growths are found in the rock and a few well-developed porphyroblasts to 3 mm are scattered throughout the rock. Biotite in the rock is different from biotite of the normal quartz monzonite in that its pleochrism is  $x = \text{very light tan}$ ,  $y = z = \text{light brownish orange}$ . The optic angle,  $2V=0^\circ$ , is the same as in biotite in quartz monzonite.



In the fine medium-grained granodiorite, which is the common rock in the intrusive, plagioclase incrustals 1 to 3 mm long is the most abundant mineral, and pyroxene altered to secondary hornblende and sphene-leucoxene (as shown in Figure 27) is the next in amount. Biotite is minor and is commonly strongly altered to chlorite and partly replaced by epidote and sphene-leucoxene. Quartz and potash feldspar occur only in the groundmass. Fresh rock from the dump of the Monarch Mine shows a flood of late quartz as shown in Figure 28. The quartz content of the rock is about 20 per cent, about twice the normal amount. The quartz may come from the vein (not considered likely) or may be related to the marginal zone of the intrusive. Magnetite, apatite and very minor amounts of zircon are the accessory minerals.

Much finer-grained rock crop out in the northern part of the intrusive where it appears to lens out to a thin sill (see section BB' on geologic map). At the north end of Bison Mountain, and particularly at those exposures north of Ontario Creek, the rock is fine-grained diorite porphyry in which plagioclase and dark green mafic minerals are the distinguishable minerals. The groundmass is composed of microgranular quartz, potash feldspar and minor amounts of fine-grained plagioclase and biotite. Plagioclase comprises 40 to 50 per cent of the rock, and is calcic andesine. The large crystals found in an early generation, are 1 to 3 mm long and were probably more or less auto-morphic, but because they are in part replaced and resorbed, the crystals are irregular in outline. The crystals are likewise generally fractured internally and are altered to more albitic composition along



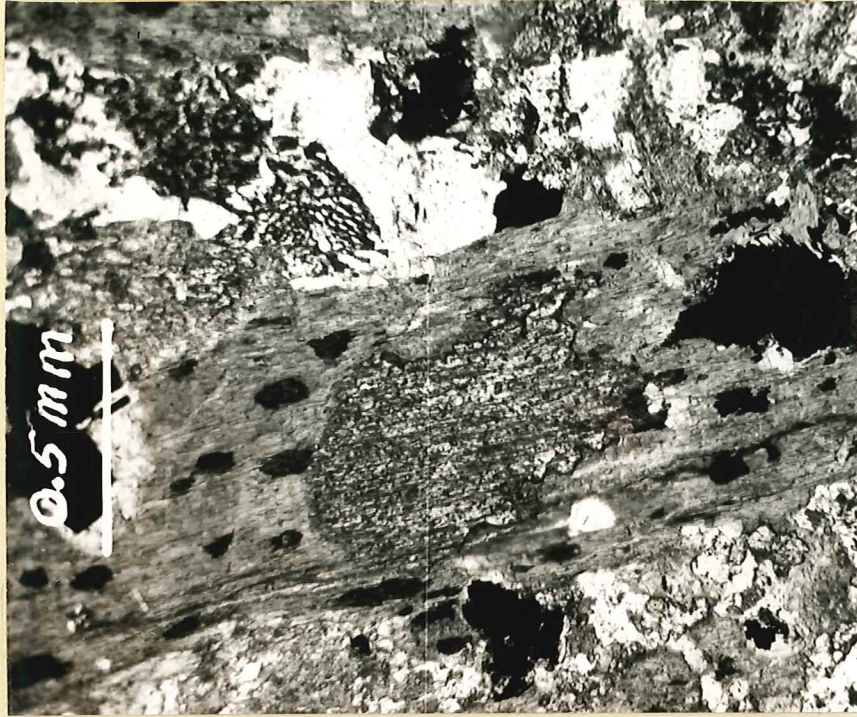


Figure 27. Photomicrograph of hornblende crystal with relic of pyroxene. Dark patches are leucoxene. Note granophyric texture in groundmass. Plane polarized light.



Figure 28. Photomicrograph of texture in Bison Mountain granodiorite showing flood of late quartz. Crossed nicols.



the fractures and at the margin of the crystals. Figure 29 shows a plagioclase phenocryst with albitic alteration along a diagonal crack, and albitic rims. Plagioclase in other parts of a section is altered to clinozoisite and sericite in the manner previously described. Biotite, where present, is pleochroic, but most commonly it is recrystallized to a fine-grained felted mass of crystals which have lower birefringence than biotite and greater refringence than ordinary chlorite. Figure 29 also shows typical fine-grained xenomorphic groundmass in which potash feldspar occurs in: (1) small xenomorphic grains with quartz; (2) very irregular poikoblastic grains, which are incipient porphyroblasts; and (3) reasonably well-developed potash feldspar porphyroblasts. (For a discussion of potash feldspar porphyroblastic growths in groundmass, see potash feldspars in the section dealing with the mineralogy of quartz monzonite).

The groundmass of the fine-grained granodiorite porphyry is composed of potash feldspar and quartz in microgranular, xenomorphic and micrographic textures. The residual magmatic solutions replaced early plagioclase phenocrysts and have converted some of the plagioclase to myrmekitic intergrowths of feldspar and quartz. In some instances, plagioclase is completely replaced by plumose, micrographic intergrowths of quartz and potash feldspar.

Near the upper contact of the granodiorite porphyry north of Ontario Creek, where the body is a relatively thin sill, the groundmass which comprises about 60 per cent of the rock, is particularly fine grained, and undoubtedly represents a chilled facies at the margin.

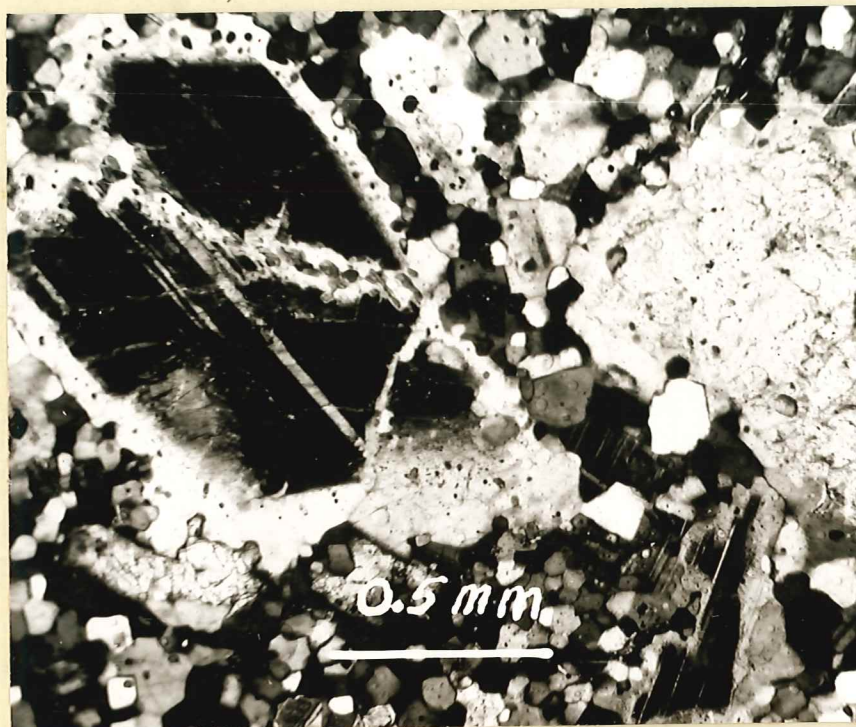


Figure 29. Photomicrograph of plagioclase phenocryst with alteration to more albitic composition at margin of the crystal and along fracture through the crystal. Note incipient stages of porphyroblastic potash feldspar growths in the groundmass and the well formed porphyroblastic grain at the right side of the photograph. The groundmass is aplitic. Crossed nicols.



Plagioclase crystals 1 to 2 mm long make up about 30 per cent of the rock, but they are strongly altered to sericite, particularly along the internal fractures. The sericite stained yellow with sodium cobaltinitrite solution indicating the presence of potash. (Gabriel and Cox 1929, p. 290). The mafic minerals, which occur in irregular clusters of crystals 0.5 to 1 mm in diameter, are altered to uralite and chlorite in which dusty opaque iron oxide has been released. The groundmass is microaplitic, xenomorphic, and is very similar to the groundmass in the chilled marginal facies of normal quartz monzonite.

#### Quartz Monzonite

Quartz monzonite is the common rock which makes up the main intrusive stage of the Boulder batholith. Quartz monzonite, with related bodies of aplite-alaskite, crops out over an area of about 16 sq mi in the eastern part of the district, as shown on the geologic map.

The quartz monzonitic rocks have been subdivided into a wide range of textural facies by the U. S. Geological Survey field parties mapping the batholith. (Becraft 1955). For purposes of this report, three broad textural facies, coarse, medium and fine grained, are described. The coarse grained are specially defined for mapping in the batholith to include those rocks which have an average grain size of 2 mm or greater, because rocks which are coarse grained in the standard sense, (4 to 5 mm average grain size) are very rare.

Although a wide variety of textures are exhibited by the quartz monzonitic rocks, and although there are minor mineralogical differences

between the various rock species, there are features which are distinctly common to all the rocks of the main stage of the batholithic intrusion. (1) The essential minerals are the same and crystallize in the same sequence; (2) the composition of the minerals is apparently the same, based on optical data; (3) the late stage crystallization features and deuteric phenomena are mineralogically the same although there are minor textural differences within the main textural facies; and (4) the accessory mineral suite is the same in all the different textural and mineralogical facies.

The mineralogy of the rocks is described in detail because a better understanding of endomorphic phenomena, particularly the succession of late crystal growths of potash feldspars is essential to a comprehension of the genesis of the rocks. The reader who is not primarily concerned with the origin of aplite-alaskite or hydrothermal veins is urged to skip over to the heading "textures" where the different types of quartz monzonite are described.

Crystallization of minerals in quartz monzonites may be considered to have taken place in three stages which merge from one into another.

Magmatic stage--crystallization of olivine (rare), pyroxene, hornblende, biotite and andesine.

Late Magmatic or Deuteric stage--crystallization of xenomorphic potash feldspar and quartz in groundmass of igneous rock (late magmatic) and porphyroblastic potash feldspar growths, potash feldspar and quartz metasomatism (deuteric).

Hydrothermal stage--the latest crystallization in parts of quartz monzonite is sericite which replaces feldspar. This feldspar-destruction stage is most conspicuously developed along the wall of the hydrothermal veins, but it is not uncommon in rocks at the top of the batholith unrelated to veins.



The petrogenesis of quartz monzonite, related aplite-alaskite and vein formation are discussed under a separate heading at the end of this section.

#### Mineralogy.

Olivine. Olivine is rarely found in quartz monzonite, although relics of olivine are not rare in chilled marginal facies. At one locality, in the Pauper area at the base of a sill-like body at the western margin of the batholith (See Structure section AA'), fresh olivine grains were identified and are illustrated in Figure 30. The distinguishing optical properties of olivine are: higher relief than augite or pigeonite,  $2V = 80^\circ +$ , optically negative. The mineral is altered along fractures to antigorite. Magnetite is not conspicuous in the fractures, but coarse grains of magnetite are intimately associated with the intensely altered olivine crystals shown in Figure 31.

Pyroxene. The common pyroxene is augite in which  $2V$  is about  $50^\circ$  to  $60^\circ$ . It occurs in small amounts in many thin sections and is usually found in the cores of hornblende crystals, although in fine-grained chilled marginal facies, augite occurs in clear, automorphic to hypautomorphic crystals. Only in the lower part of the sill-like body in the Pauper area were two pyroxenes noted. Pigeonite, in which the optic angle  $2V$  ranges from nearly zero to about  $20^\circ$ , was found in automorphic grains included in augite crystals, and inclusions in plagioclase.



Figure 30. Photomicrograph of olivine in basic quartz monzonite at base of sill-like body in Pauper area. Note alteration to antigorite along fractures. Plane polarized light.

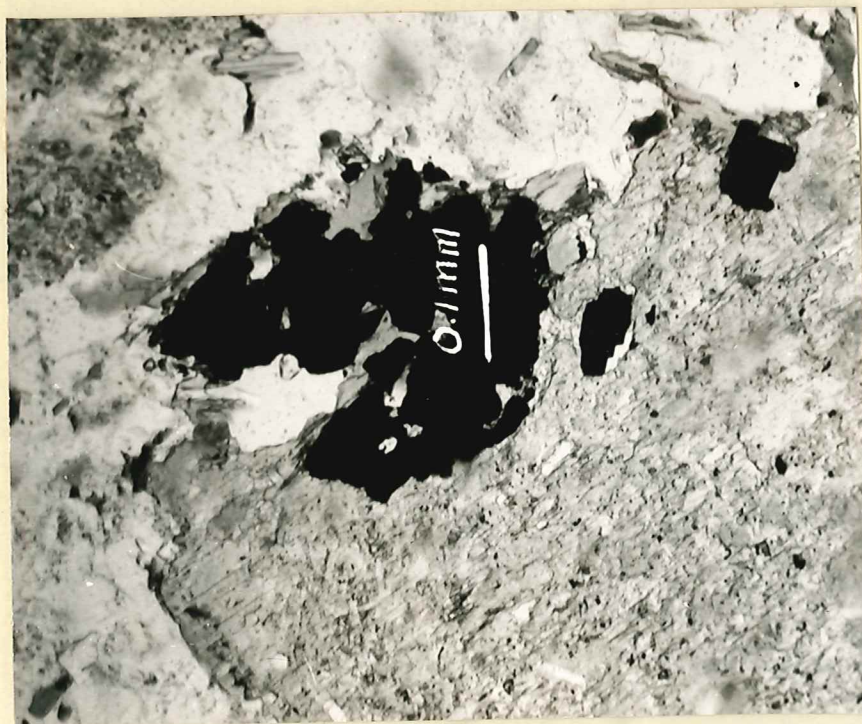


Figure 31. Photomicrograph of relic olivine in altered pyroxene. Magnetite and antigorite in round mass which has the shape and size of olivine grain. Plane polarized light.



Hornblende. The common amphibole is described in the literature (Weed 1912) as hornblende. It is colored, pleochroic, x = very pale green, y = light green, z = light bluish green. Rarely does it have darker olive green colors which are more characteristic of normal hornblende. The extinction angles measured range from  $17^{\circ}$  to  $20^{\circ}$ , and the mineral is optically negative. The mineral appears to be an actinolitic hornblende because of its light color. Crystals are commonly 1 to 5 mm in length, but often they are shredded and strongly altered to chlorite, or less commonly to a nearly white chlorite-like mineral which has moderately low birefringence ( $.008 \pm$ ), is colorless, and has a felted texture similar to that found in some chlorites. The margins of the crystals are commonly very irregular, a feature which is clearly evident when the mineral is examined with a hand lens. Figure 32 shows a typical hornblende crystal, not strongly altered, whereas Figure 33 illustrates an altered hornblende which has gone to chlorite, sphene-leucoxene, epidote and opaque iron oxide grains.

Biotite. Biotite, although occurring in a wide variety of grain sizes from very fine in the groundmass of fine-grained rocks, to crystals 2 mm across in the coarse-grained rocks, has virtually identical optical properties in all of the rocks. It occurs in hypautomorphic grains which, in hand specimen, appear bright and fresh. Figure 34 shows nearly automorphic biotite and irregular xenomorphic hornblende. In thin section it is commonly fine grained and ragged in outline where it is in intimate association with hornblende. The mineral is practically



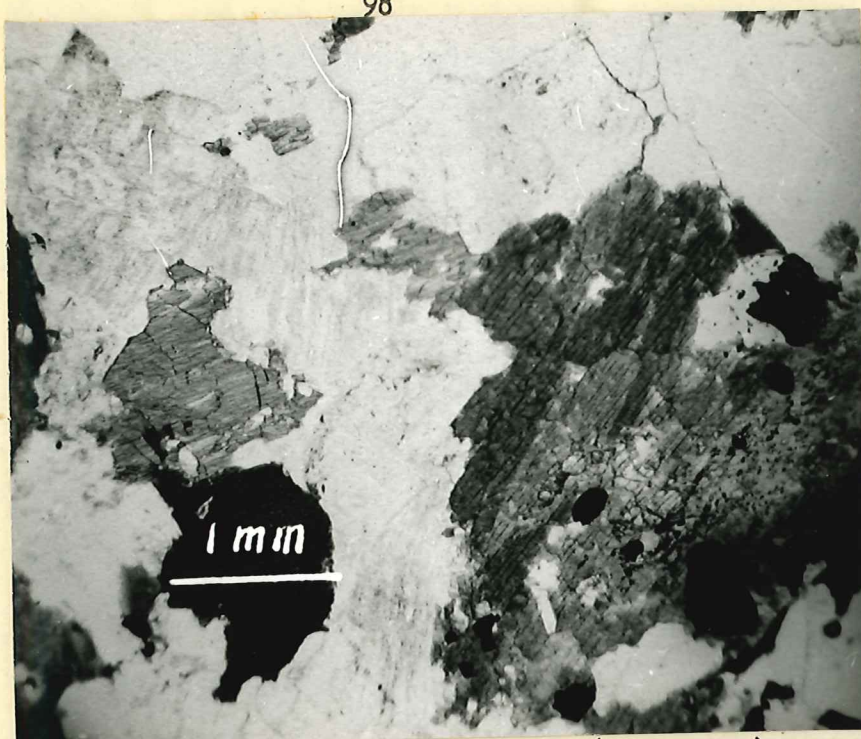


Figure 32. Photomicrograph of hornblende (dark gray) in quartz monzonite. Note irregular margins of grains. Plane polarized light.



Figure 33. Photomicrograph of hornblende altered to chlorite, opaque iron oxide, leucoxene (blotchy dark gray) and epidote (high relief). Groundmass strongly altered deuterically. Note granophyric textures. Plane polarized light.



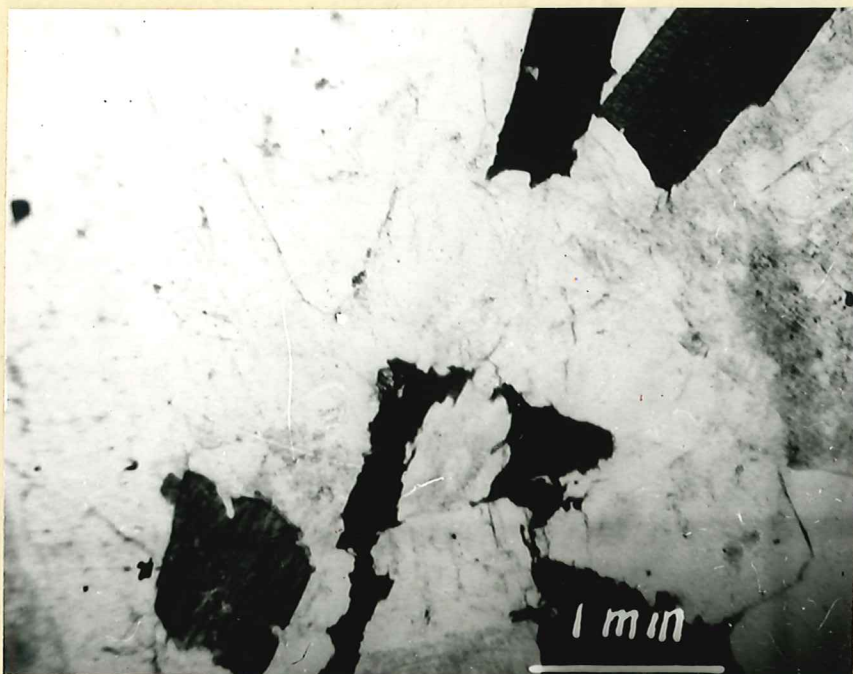


Figure 34. Photomicrograph of nearly automorphic biotite (dark gray rectangular grains) and ragged hornblende in quartz monzonite. Plane polarized light.

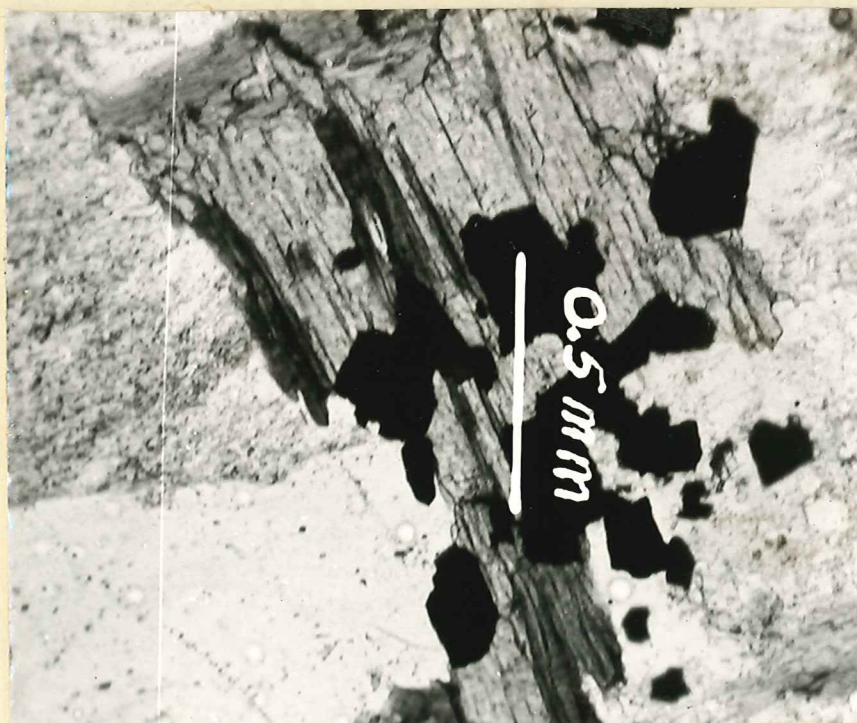


Figure 35. Photomicrograph of biotite (dark gray) altered to muscovite (gray) with released magnetite (black). Note potash feldspar (white) replacement of muscovite and biotite (left) and quartz replacement (right). Plane polarized light.

uniaxial  $2V$  is nearly zero, pleochroism is about the same throughout,  $x = \text{tan}$  or very light tan,  $y = z =$  dark brown or dark reddish brown.

Biotite and hornblende commonly occur in clusters of mafic minerals in the rock, commonly with minor amounts of coarse magnetite grains and irregular to wedge-shaped crystals of sphene. Alteration to chlorite, variety penninite, is typical. The penninite shows a somewhat lighter anomalous Berlin blue color than usual, and is commonly noted to have a slight lavender color which is quite distinctive. Alteration to coarse white mica (muscovite ?), is rare, but in those cases, magnetite is clearly liberated from the biotite as shown in Figure 35. Elsewhere, magnetite may occur in dusty specks throughout strongly deuterically altered biotite.

Plagioclase. Plagioclase occurs in automorphic to hypautomorphic crystals up to 5 mm long. They are clearly the first essential mineral to crystallize in significant amounts. The composition varies from labradorite (Ab 40 An 60) in the more calcic facies of the batholith (granogabbro) to about Ab 50 An 50 in many of the marginal facies to calcic andesine (Ab 60-50 An 40-50) in the normal quartz monzonites. It is sometimes difficult, however, to determine the composition of plagioclase accurately because it is zoned, usually with more calcic cores. In sections which exhibit Albite twinning, zoning is not conspicuous, the albite twinning is sharp and clear. Sections normal to 010 and 001 have stout rectangular to nearly square shapes which show pericline twinning and, in some cases, are clearly zoned.



Sections in which 010 is nearly parallel to the microscope stage, strong zoning is conspicuous and albite twinning is not observed. Optic axis figures are commonly obtainable from these sections. The normal calcic andesine about Ab 55 An 45 composition, have optic angles  $2V$  varies from about  $70^\circ$  to  $75^\circ$ , which is somewhat smaller than  $2V$  values given in the standard texts (Winchell 1933, Rogers and Kerr, 1942, Wahlstrom ). Krump and Katner (1953, p. 31) illustrate the range of optic angles  $2V$  in plagioclase of this compositional range in a quartz monzonite from  $72^\circ$  to  $88^\circ$ .

Albite twinning is the most common twin however, very often both Carlsbad and Pericline twins are observed. Figure 36 shows typical plagioclase crystals. In one, albite and pericline twinning is distinct; in the other the mineral is clearly zoned, and the inner part of the crystal is crowded with inclusions whereas the outer part of the crystal is fresh and practically devoid of inclusions.

At the Pauper area, near the base of the marginal sill, (which is the most basic quartz monzonite recognized), augite, pigeonite, biotite, magnetite, and apatite inclusions were identified in the core of a plagioclase phenocryst as shown in Figure 37. Normally the mafic inclusions are altered to chlorite or actinolitic hornblende. See Figures 38 and 39.

Fracturing is common in the plagioclase crystals. A distinctive internal cracking is seen in many of the large crystals, as shown in Figure 33. Some crystals are broken and healed, as if the crystals were already well developed when the magma was intruded into position.

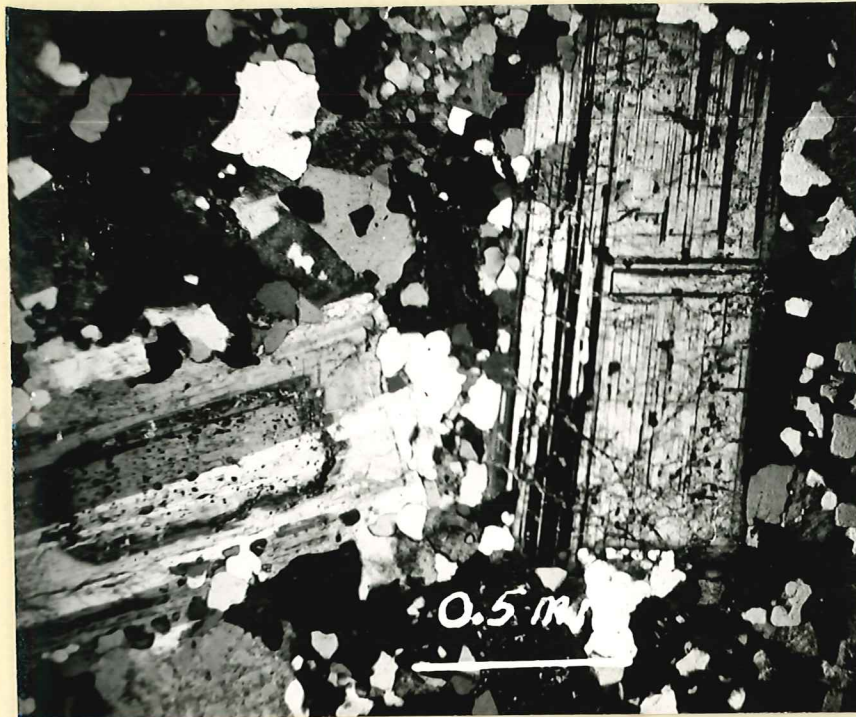


Figure 36. Photomicrograph of plagioclase phenocrysts in microgranitic groundmass in quartz monzonite. Note albite and pericline twinning in crystal at right, and zoned crystal with inclusions in core. Crossed nicols.



Figure 37. Photomicrograph of central core of plagioclase phenocryst. The inclusions are: clear rounded pigeonite and augite, clear elongated apatite, biotite (dark gray) and opaque octahedra of magnetite. Plane polarized light.



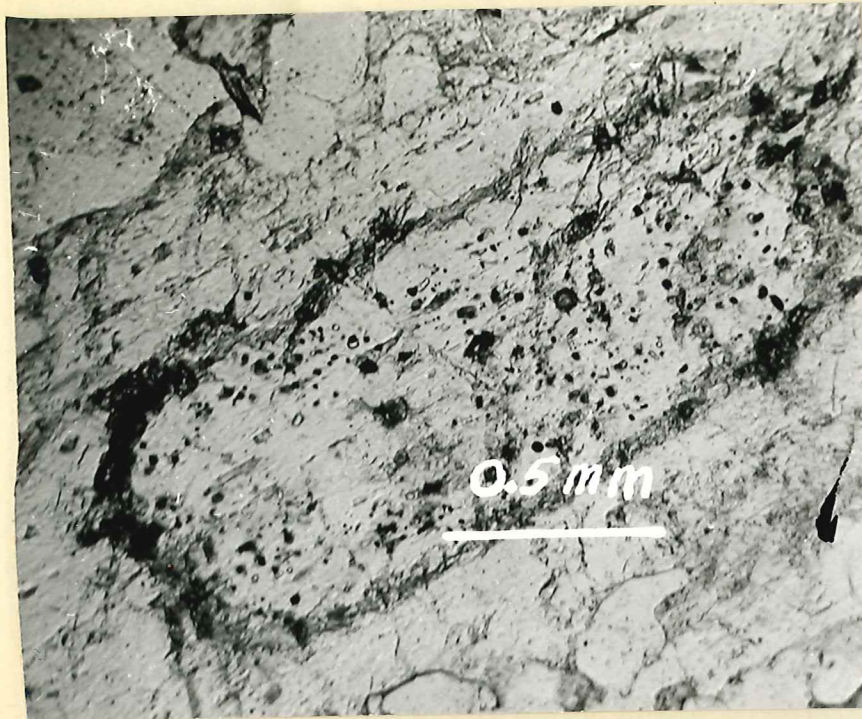


Figure 38. Photomicrograph of plagioclase phenocryst with well marked central core. The inclusions are altered (compare with figure 37). Note replacement by potash feldspar at lower margin of photograph. Crossed nicols.



Figure 39. Photomicrograph of same plagioclase phenocryst as in Figure 38. Plane polarized light.



The internal fractures in plagioclase crystals provided avenues for late magmatic solutions which, no longer in equilibrium with the early-formed plagioclase crystals, produced more albitic feldspar (albite-aligoclase) along the fractures and at the margin of the crystals as shown in Figure 40. A similar alteration zone is always present at the advancing front of replacing potash feldspar crystals as shown in Figure 41. Usually the albitic rims are not twinned, however in one instance where potash feldspar replacement of plagioclase occurs adjacent to an incipient vein-like structure, the albitic rim is conspicuously twinned. Kohler (1948) and Laves (1950) suggest that twinned albitic plagioclase is higher temperature than the untwinned albitic alteration product. If there is any thermal significance to the twinned albitic rim, then much of the deuteritic stage is normally somewhat below the high-low temperature boundary, but in a few instances may form at the higher temperature stage.

Alteration to clinozoisite is also distinctly a deuteritic phenomenon (Gilson 1939). Albitic zones, some potash feldspar patches, and fleck sericite\* are all found in plagioclase at this stage. Partial replacement by potash feldspar and quartz, and replacement to myrmekitic growths of plagioclase and quartz of graphic intergrowths of potash feldspar and quartz are common. In the hydrothermal stage, strong sericitic alteration takes place. Intense sericitization in localized "spots" in quartz monzonite are usually associated with pyrite crystals

---

\*Note: fleck sericite is distinguished from all-over sericite alteration. It is distinguished by the sparsely scattered sericite commonly noted in plagioclase at the deuteritic stage.



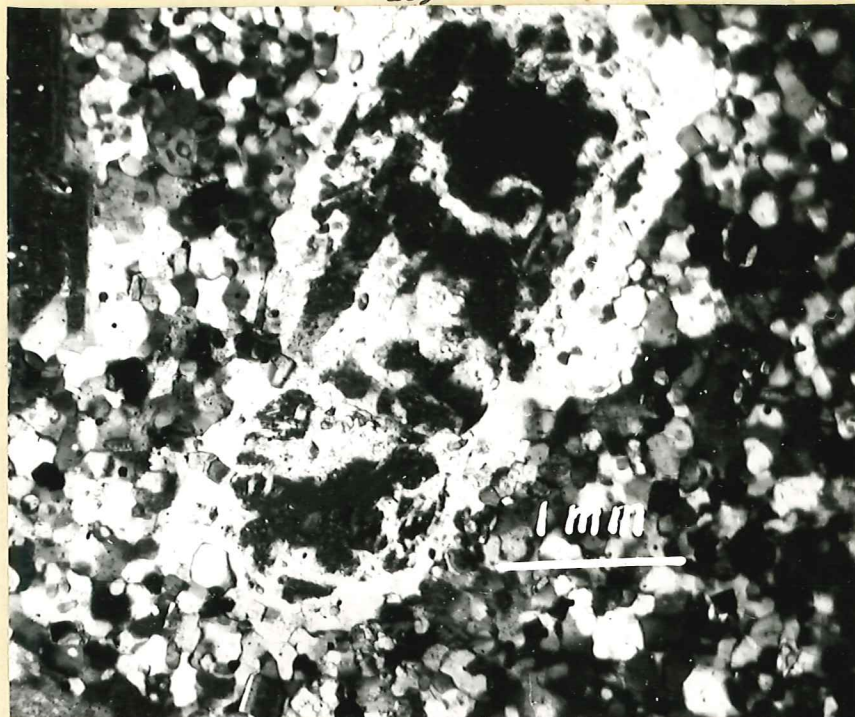


Figure 40. Photomicrograph of plagioclase phenocryst strongly altered deuterically. Note albitic rims, and patches of albite throughout phenocryst. The groundmass is microgranular with plagioclase, potash feldspar and quartz. Crossed nicols.



Figure 41. Photomicrograph of plagioclase crystal with albitic rim. Note incompletely organized potash feldspar (gray) crystal in center of photograph. Crossed nicols.



which may be sulfidized magnetite grains. Plagioclase is altered to sericite and biotite to chlorite.

Potash feldspars. The mineralogy of the potash feldspars is an especially interesting problem. Although quite a bit of time has been spent on the problem, much more time will be required to solve the complex mineralogical assemblage. The early rock-forming minerals appear to follow the normal line of crystallization found in igneous rocks generally (Bowen 1938). The potash feldspar and quartz crystallization, on the other hand, does not represent a simple sequential crystallization from magma. Actually, a variety of minerals and textures occur at this late-stage crystallization period, and these varieties of textures appear to be largely responsible for the textural variation recognized in hand specimens.

(Orthoclase). Orthoclase is the common potash feldspar. It occurs in: (1) small xenomorphic crystals in the groundmass; (2) in intergrowths with quartz in complex micropegmatitic and micrographic textures; and (3) in a wide range of granoblastic and porphyroblastic growth stages, most conspicuously developed in the coarse-grained rocks, where potash feldspars attain a cross section of 5 mm and more, as shown in Figure 42, and rarely in much larger crystals. (At a small quarry southwest of Boulder, large potash feldspar porphyroblasts 25 mm long display fine crystal forms. Rarely, however, do the large porphyroblastic crystals form in the Boulder batholith and no large crystals have been found in the rocks of the Elliston district.)





Figure 42. Photomicrograph of large potash feldspar porphyroblast. Note albitic rim on plagioclase against advancing front of potash feldspar. Note also granophyric replacement of plagioclase. Crossed nicols.

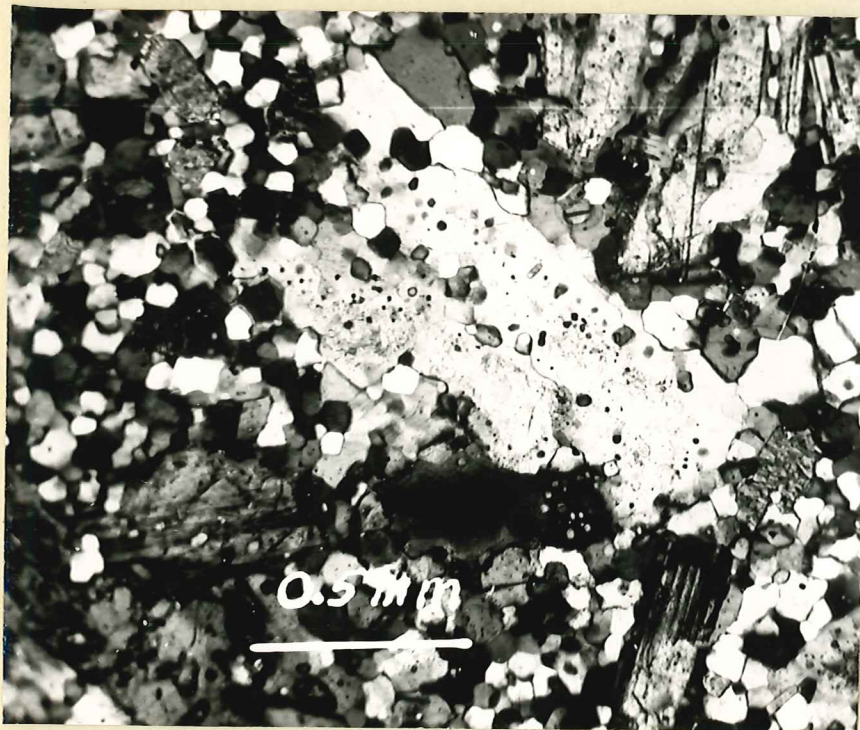


Figure 43. Photomicrograph of small porphyroblastic potash feldspar crystal. Note the abundance of inclusions. Crossed nicols.

Orthoclase is commonly perthitic, usually in irregular patches and lenses which suggest a replacement origin, however, the writer hesitates to call them replacement perthites because of the complex history of perthitic orthoclase. (See separate heading "microperthite.") Although only a few measurements have been made with the Universal Stage, (however a great many interference figures have been examined by routing<sup>Z</sup> petrographic methods)  $2V$  for the common orthoclase appears to be about  $60^\circ$ . Orthoclase stains yellow when treated with a sodium cobaltinitrite solution. (Gabriel, et. al. 1929; Chayes 1952; Rosenblum 1955.)

In many of the quartz monzonites, potash feldspar crystallizes in part in interstices between early-formed crystals, but under favorable conditions, tends to aggregate and form large xenomorphic to hypautomorphic grains of orthoclase. The succession of steps in the formation of the large crystals is similar, if not identical, to potash feldspar porphyroblastic growth in some metamorphic rocks.

The first step in the formation of a porphyroblast is the development of amoeba-like projections on a xenomorphic grain of orthoclase in the groundmass, as shown in groundmass in Figure 29. After this, optical continuity is established over an area of several xenomorphic orthoclase grains which poikiloblastically<sup>SP</sup> enclose quartz grains, as shown in Figure 43. A strongly poikiloblastic feldspar<sup>SP</sup> then appears after which some of the included quartz in the center of the new crystal is displaced. This is spoken of in metamorphic terms as "clearing" of the crystal, two stages of which are shown in Figure 44. At the margin





Figure 44. Photomicrograph of potash feldspar porphyroblast. Note that the center of the crystal is well cleared but the rim is strongly poikioblastic. Crossed nicols.

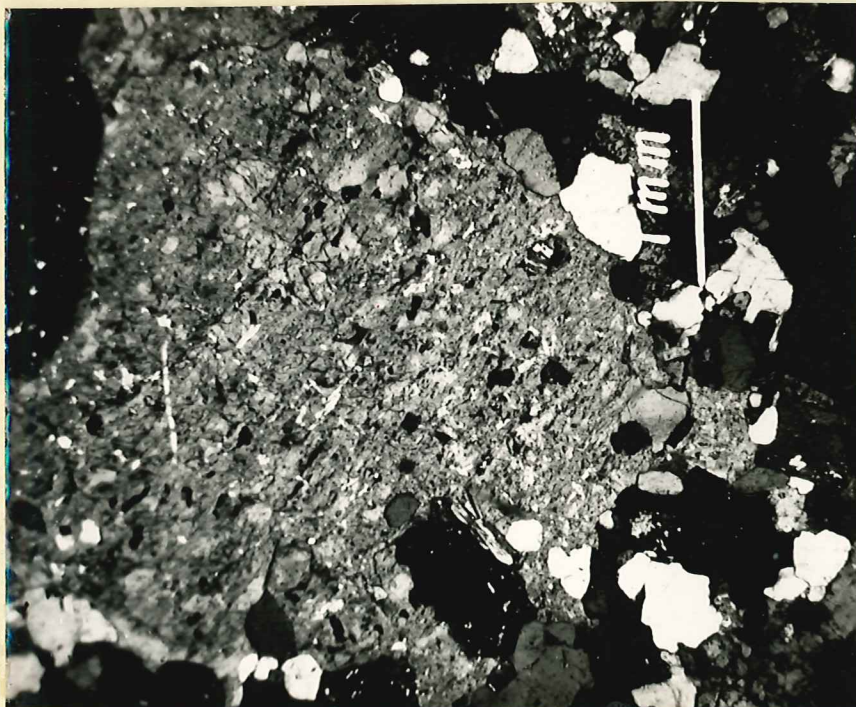


Figure 45. Photomicrograph of poikioblastic texture in anorthoclase-like potash feldspar porphyroblast. Crossed nicols.



of some of the new larger feldspar crystals, there is a concentration of quartz. Plagioclase and biotite may be poikiloblastically enclosed, however these too may be eliminated by replacement. Figure 45 is a strongly poikiloblastic porphyroblast. Finally a porphyroblast forms which usually has quite irregular boundaries. Although the core of the porphyroblast may be clear, commonly the margins are poikiloblastic.

(Microcline). Microcline occurs in minor amounts in the coarser-grained rocks and shows characteristic grid twinning. Few of the medium- and coarse-grained rocks do not have potash feldspars which exhibit microcline-like twinning. In some instances, grains of potash feldspar with pronounced grid twinning have large optic angles and are definitely microcline. The optical angle was measured on the Universal Stage,  $2V = 80^\circ \pm 2^\circ$  for one such crystal.

Grid twinning, somewhat less distinct than "typical" microcline, is a very common phenomenon in some of the potash feldspars. It is probable that most of the minerals which exhibit grid twinning are not actually microcline at all. The great majority of the grains which exhibit multiple grid twinning, particularly those cases where the twinning is rather indistinctly developed in discontinuous patches, do not have large optic angles. The mineral is not completely uniform and not necessarily one single mineral, but constitutes a variety of potash-bearing minerals herein called anorthoclase.

(Anorthoclase). Anorthoclase is chemically a potash-rich albite. The minerals which have been tentatively assigned to the anorthoclase



group comprise a series of replacement minerals in plagioclase which have a small to medium optic angle and an index of refraction greater than orthoclase and less than balsam. Indistinct grid twinning, commonly in irregular patches, is a characteristic feature.

There is considerable confusion in the literature about the optical properties of anorthoclase and doubt has been cast on the validity of the species. (Laves 1952, p. 568). Because of the lack of agreement in the literature and because of the importance of the mineral or minerals herein called anorthoclase in the mineralogy of the late-stage crystallization of quartz monzonite, the subject of anorthoclase is considered in detail.

The term anorthoclase was first proposed by Rosenbush (1855). "Anorthoklas" whose definition, translated by Laves, is as follows: "The most important property of those soda-plagioclase feldspars that will be called the 'anorthoclase series' is the cleavage angle P:M that 'appears' not to deviate from a right angle, yet which must do so. This series differs from the plagioclase which has obvious oblique cleavage." (Laves 1952, p. 368, from Rosenbush 1885). The composition range was placed between Or 33 Ab 67 and Or 18 Ab 82, with a small but variable anorthite content. Perthitic structures, so common in microcline are not observed in anorthoclase. (Rosenbush 1904, p. 326). Laves reported that Brogger (1890, pp. 539, 563, 1897, pp. 12-14) objected to the term anorthoclase and preferred soda-microcline (Nathron-mikroclin) because he assumed a close relationship to microcline, a relation which has been found to be incorrect

according to the researches of Laves (1952, p. 589). A third usage, that of Oftedahl (1948, pp. 56-58), considers the inhomogeneity of a cryptoperthitic nature of anorthoclase as an important characteristic. Certain symmetry relations were, however, ignored by Oftedahl (Laves 1952, p. 569), as well as by Alling (1926, p. 600) who considers that anorthoclase may or may not be homogeneous, and may or may not be triclinic. The range of composition of anorthoclase is Or <sup>70</sup>50 Ab <sup>30</sup>50-- Or 20 Ab 80 according to Alling (1936, p. 600).

The lack of general agreement about anorthoclase in the modern texts is discussed at some length by Laves who states: "No other mineral of importance appears to play such an ambiguous role in mineralogical and petrographic literature as 'anorthoclase'. Numerous attempts have been made to define and characterize this mineral, and virtually every publication and textbook considers it in a different way. . . Most recent textbooks describe anorthoclase as a sodium-rich variety of microcline, (for example, see Kraus, Hunt, and Ramsdell, 1951). Winchell, with this point of view, proposed a microcline-anorthoclase-analbite series. In his original paper on the subject (1925) he used the anorthoclases of Forstner (1884) to complete the optical data for his series. He neglected, however, many of Forstner's other observations on the same mineral. For example, Forstner determined, among other things, values of 'a' on analyzed specimens, three of the values being  $91^{\circ} 01'$ ,  $91^{\circ} 42'$ , and  $92^{\circ} 3'$ . He described the triclinic-microclitic inversion and thus showed clearly the variability of the crystallographic data. In Winchell's book (1951) however, anorthoclase



is described on page 311 as having a  $\alpha = 91^{\circ} 19'$ , giving the impression of a well-defined value for a fixed mineral." (Laves 1952, pp. 567-8).

The recent researches of Laves on the alkali feldspars indicate that the analbite end of the sanadine-barbierite series is stable at high temperature, but the triclinic form (analbite) may contain a considerably greater amount of Or at lower temperatures. Instead of using the term anorthoclase, the mineral is designated K-analbite. Figure 49 is particularly revealing because it shows that on cooling below about  $700^{\circ} \text{C}$ , the stable high temperature minerals become perthitic around  $600^{\circ} \text{C}$ , and at lower temperatures for the anperthites. The K-analbites are considered to be disordered, relative to Al/Si, whereas the perthites and albite are ordered forms. (Laves 1952, pp. 556, 561).

All of Laves' observations are based on experimental work with melts and investigations of various mineralogical changes that take place when a natural-occurring mineral is heated from low to high temperatures. In the investigation of the quartz monzonites and related alaskites, the anorthoclase is not considered the product of crystallization of a complex disordered mineral from a hot silicate solution, rather it was formed, in large part, by replacement of plagioclase feldspar by the introduction of potash and the elimination of lime, possibly at relatively low temperatures. (See Winchell's phase diagram in Figure 46.) Since the optical properties which have been determined, namely  $2V$ , the beta index, and the properties which have been observed, namely the fine microcline twinning, fit Winchell's data for anorthoclase, (Winchell 1933, p. 367), the term anorthoclase

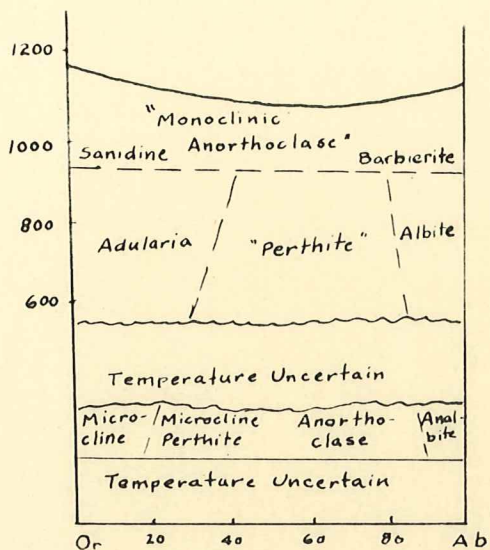


Figure 46 . Phase diagram after Winchell (1925,1951)

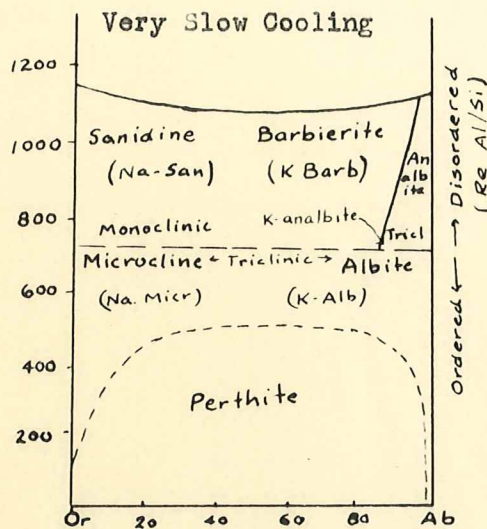
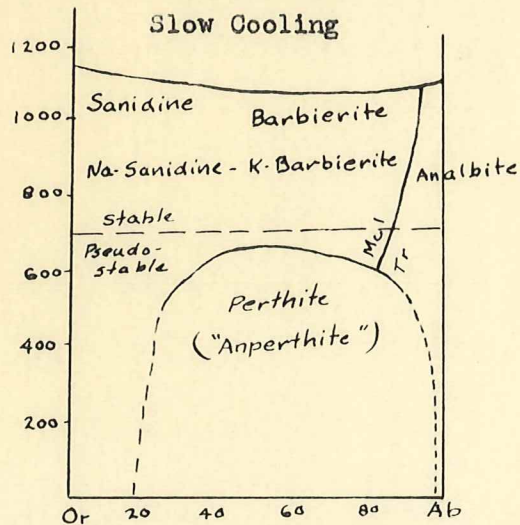
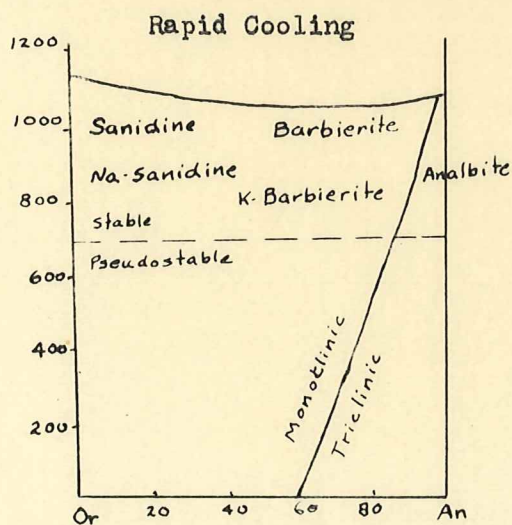


Figure 47 . The probable phase diagram of the alkali feldspars that have cooled very slowly (Laves 1952 p. 561)

Figure 48 . Phase relations of material cooled rapidly enough to suppress exsolution and ordering of Al and Si ions. (Laves 1952 p. 556)

Figure 49 . Phase relations of alkali feldspars that have cooled slowly enough to unmix with respect to the alkali ions but not slowly enough to order the Al and Si ions. (Laves 1952 p. 557)





is employed. The term K-analbite of Laves possibly should be restricted to those minerals which crystallize from hot sillicate melts or solutions. In this regard, the original diagram of Winchell (1925) Figure 46, is of especial interest. At high temperature, the sanidine-barbierite series comprises a complete series of solid solution minerals designated as "monoclinic anorthoclase". On cooling, adularia and albite are end members, with perthites in between. At some uncertain but definitely indicated lower temperature, microcline, microcline perthite, anorthoclase and analbite form. Thus Winchell's anorthoclase, formed at relatively low temperature, (and probably ordered), more nearly fits the crystallization history of the so-called anorthoclase found in the Boulder batholith than Laves' K-analbite which is considered to be formed from a melt at relatively high temperature and in a disordered state.

Anorthoclase is a common mineral in alaskite and deuteritic stage crystallization of quartz monzonite, although always much less abundant than orthoclase. It is most conspicuous in the medium-and coarse-grained rocks where deuteritic phenomena are most pronounced. Wherever potash feldspars are found replacing plagioclase, anorthoclase may be expected, and it in turn is replaced by orthoclase. Normally the amount of potash feldspar which crystallizes in quartz monzonite is somewhat less than, to about equal to, plagioclase. Where the replacement processes are operative which yield potash feldspars in greater than normal amount by the elimination of plagioclase, rocks of granitic and alaskitic composition are formed. In the marginal (or incompletely replaced) zones of alaskitic bodies, anorthoclase is, in some cases, abundant.

The common sequence in the replacement of plagioclase to orthoclase appears to take place in the following steps:

- (1) Plagioclase is attacked and altered to albitic composition in a narrow zone between the plagioclase and the invading potash feldspar. (See Figures 40 and 41).
- (2) The invading newly-forming potash-bearing mineral may develop a conspicuous grid twinning, (shown in Figure 50), which aids in distinguishing anorthoclase from normal orthoclase, or where grid twinning is absent, some other intermediate potash-bearing feldspar may form, as shown in Figure 51. The birefringence of the anorthoclase-like potash-bearing minerals is somewhat less than orthoclase. Figure 52 shows anorthoclase replacing plagioclase. In some instances, there are several different potash-bearing feldspars formed, each exhibiting optical continuity and displaying complex replacement textures, as shown in Figure 53.
- (3) The end product is normally orthoclase which is commonly perthitic. The early anorthoclase may replace the plagioclase crystals incompletely as shown in Figure 54, or, in some rather striking instances, may take on the form of the plagioclase phenocryst but have the optical properties of anorthoclase, shown in Figure 55.

The optic angle  $2V$  was determined on the Universal Stage to be  $54^\circ \pm 2^\circ$ , (from 5 thin sections). The mineral is optically negative,  $n_b = 1.529 \pm .001$  (one hand-picked grain). The mineral responds variably in sodium cobaltinitrite solution, some grains do not stain, others are partly stained. The composition of the anorthoclase is estimated to be about Or 25 Ab 75 (Winchell 1951) (See Figure 46).





Figure 50. Photomicrograph of anorthoclase (dark gray with indistinct microcline-like twinning) partly replaced by orthoclase-microperthite. Note relics of anorthoclase in orthoclase. (Scale should be 0.5 mm) Crossed nicols.



Figure 51. Photomicrograph of plagioclase crystal partly replaced by anorthoclase. Albitic rim is white in upper grain, black in lower grain. Note quartz replacement of anorthoclase. Crossed nicols.





Figure 52. Photomicrograph of anorthoclase with relics of plagioclase. The larger relics have optical continuity of albite twins which indicates that the replacement was completely static. (Scale should be 0.5 mm) Crossed nicols.



Figure 53. Photomicrograph of anorthoclase-like mineral (speckled) partly replaced by other potash feldspar. Note relic inclusion of plagioclase in potash feldspar at left. (Scale should be 0.5 mm) Crossed nicols.



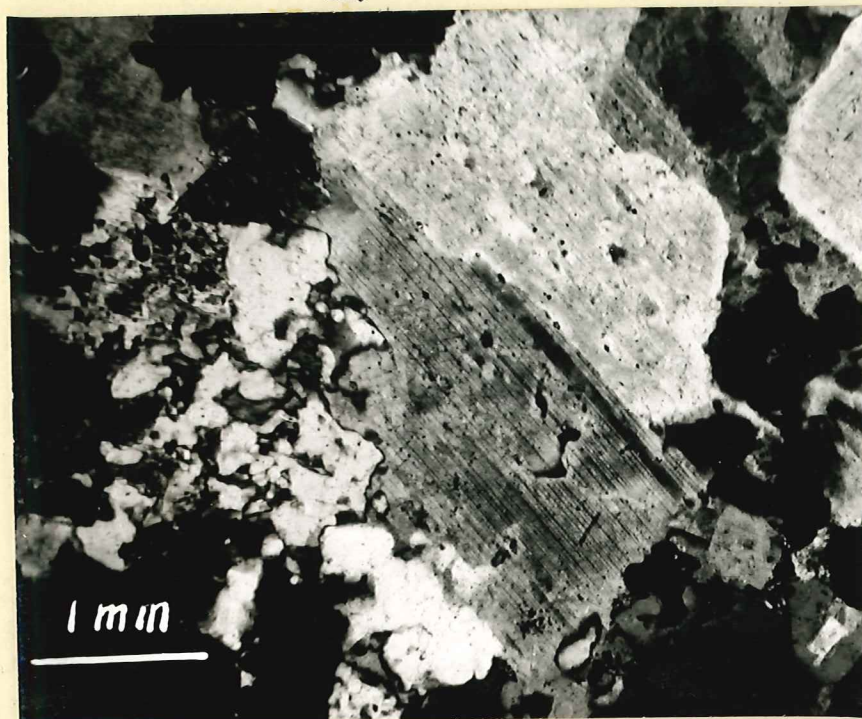


Figure 54. Photomicrograph of plagioclase phenocryst (gray, with albite twins) partly replaced by potash feldspar (lighter gray). Note quartz replacement of plagioclase. Crossed nicols.



Figure 55. Photomicrograph of anorthoclase which replaced plagioclase crystal and shape of original crystal retained. Quartz occupies position similar to the zone noted in plagioclase (see Figure 38) (Scale should be 0.5mm) Crossed nicols.



(Microperthite). Orthoclase microperthite occurs in many of the quartz monzonites, but it is conspicuously most abundant in the alaskite bodies in which the potash feldspar content is large. In most cases, fine-grained orthoclase in the groundmass of quartz monzonites is not perthitic and in a great many instances, orthoclase in larger, porphyroblastic crystals, is only slightly perthitic. Orthoclase which forms by replacement of plagioclase (commonly with anorthoclase as an intermediate step) is commonly perthitic. The blebs of albitic plagioclase composition occur as small irregular lenses, strings and ragged patches in the potash feldspars. The non-uniform distribution of the blebs and the irregular shapes of the blebs make an exsolution origin improbable. (Anderson 1928)(Alling 1921).

Replacement textures, described by Anderson (1938)(Alling 1921), Faersler and Tremblay (1946) appear to be similar to the perthites in the granitic rocks, but the lack of evidence of soda metasomatism in the late magmatic or deuteritic stage leads me to question a simple replacement origin whereby late albitic solutions replace earlier-formed orthoclase.

Observations have been made which seem to suggest that a somewhat different mode of origin for the perthites from either of the classic concepts, vis. exsolution and replacement is required. Commonly, orthoclase, which forms by replacement of plagioclase, and especially in those cases where anorthoclase is an intermediate step, is perthitic. Gates (1953, p. 65) noted the association of anorthoclase and perthite, and reports the following: "no unquestioned orthoclase has been reported in modern literature. The existence of anorthoclase and its significance has been in doubt since it was first described. It is concluded



here that anorthoclase may occur as such as single crystals but rarely does. In this study it has been seen only as remnants and usually small remnants, in crystals which are largely perthitic." Gates (1953, p. 65) implies that the anorthoclase was formed by alteration of the albitic blebs after the perthite had formed. In this investigation, some evidence points to anorthoclase as an intermediate stage in the origin of orthoclase microperthite, and thus patches of anorthoclase in perthite are regarded as relics, and so the textures of the patches indicate.

Some, if not all, of the microperthite appears to have formed in the following manner: (1) In the replacement of plagioclase feldspar, anorthoclase is commonly the first step in the formation of the potash feldspars. Lime is expelled and potash introduced to produce a mineral which is a potash-bearing albite. (2) As additional potash is added, potash feldspars are formed (orthoclase or rarely microcline) which are strongly perthitic. The soda in the anorthoclase was very incompletely eliminated and instead is thought to be organized into blebs in convenient positions in the crystal lattice. In some of the minerals, particularly in the incompletely formed alaskites, relics of anorthoclase are commonly preserved in strongly perthitic potash feldspars. Figure 56 shows coarse microperthite in alaskite.

The abundance of reorganization perthite in the alaskites fits the genesis outlined because alaskite is the product of potash metasomatism of early-formed, possibly incompletely crystallized, quartz monzonite. The recrystallization of a considerable amount of plagioclase would yield a large amount of soda which is, at least partly, accounted for in the perthitic feldspars. The lime is apparently lost





Figure 56. Photomicrograph of coarse microperthite and quartz in alaskite at Ontario mine. (Scale should be 0.5 mm) Crossed nicols.

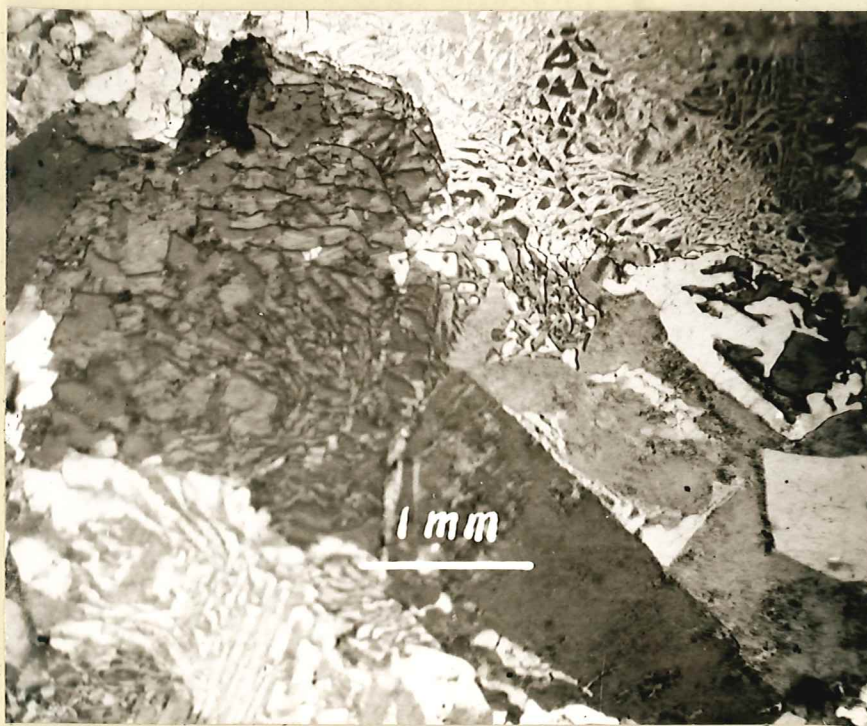


Figure 57. Photomicrograph of replacement granophyre in potash feldspar. (Scale should be 0.5 mm) Crossed nicols.



to the system except for the small amount accounted for in tourmaline. The potash feldspars in the normal quartz monzonite, on the other hand, involve a reorganization of potash feldspar in the groundmass with only incidental replacement of plagioclase, consequently the potash feldspars in the quartz monzonites are distinctly less perthitic.

Just how such a reorganization could take place is not known. Recently Chayes (1952) and earlier Phemister (1926) pointed out the stress environment and the formation of perthite. Possibly slight stresses in the well-crystallized quartz monzonite may have favored the formation of reorganization perthites. Makinen (1913, p. 35) notes a close relation between microcline grid twinning and perthite and considered that the control was one of stress. In view of the findings in this investigation, I would like to know if the microcline grid twinning observed by Makinen is actually in microcline. Insufficient optical data was presented to answer the question. If the mineral is anorthoclase, with microcline-like grid twinning, the origin of the perthite may be similar to that herein described.

This concept of the formation of perthite is in contrast to published accounts and is at the opposite pole of opinion from Tuttle (1952) who recently attempted to show that late plagioclase in granite is the ultimate of unmixing from disordered potash feldspars which, at elevated temperatures contained appreciable amounts of soda. Unmixing is supposed to have taken place with first the formation of albitic blebs; then very coarse blebs; and ultimately potash feldspar, without appreciable amounts of soda (and without unmixed albite) and

plagioclase, (albite-oligoclase). If the formation of the reorganization perthites is reasonably accurate, then the potash feldspars may have formed at such a low temperature (below 500° C) that very little soda could be accommodated in the lattice and no exsolution could have taken place. Some unpublished analyses of potash feldspars from the Boulder batholith support the hypothesis that the soda content is not as great as might have been expected if the perthites were exsolution perthites. (Informal discussion with Klepper, analyses not available.)

Quartz. Quartz occurs in xenomorphic grains in the groundmass of all quartz monzonites. It is commonly intergrown with potash feldspar, and less commonly with plagioclase, in granophyric intergrowth. Figure 57 shows some of the intricate granophyric intergrowths which are found in replacement aplites and in deuterically altered groundmass of the Bison Mountain granodiorite. Figures 42, 51 and 54 show more typical granophyric patches in quartz monzonite.

Some very interesting work by DeVore (1956, 1955) shows that the tendency in all of the acidic rocks is to yield an end product of near eutectic composition of feldspar and quartz, whether by crystallization of a magma, or by metasomatic reorganization to granitic composition by granitization. If the minerals are not influenced by external forces, the boundary between the minerals becomes extremely intricate, such as those found in the granophyric textures.

It has been the observation in this investigation that static, late potash metasomatism favors the growth of granophyric textures, where any evidence of movement or deformation tends to destroy the



intergrowths. In replacement aplites, micrographic intergrowths are the rule, whereas in the through-going aplites, granophyric textures are not found. In one very small replacement aplite, granophyric textures are found at the end of the thin section, and are absent in the vein-like part of the aplite not  $\frac{1}{2}$  in<sup>ch</sup> away. In the groundmass of the Bison Mountain granodiorite, an abundance of granophyric textures are found, likewise they are common at the boundaries of small replacement pegmatites, and at the advancing front of the potash feldspar segregation pods which form in the deuteritic stage.

Quartz is commonly the youngest essential mineral to crystallize. At the margins of the batholith, quartz is often present in greater than normal amounts, and is represented by a flood of late quartz which replaces all earlier-formed minerals, particularly in the groundmass.

Accessory Minerals. Spene is a common accessory in the quartz monzonites, and is usually found in association with mafic minerals. The mineral commonly occurs in fairly coarse grains and clusters of grains and in some cases has the distinctive wedge shape commonly attributed to spene. It is often formed as a by-product of biotite and pyroxene destruction.

Apatite is found in clear crystals, commonly in close association with magnetite and mafic minerals. The crystals are fairly stout and rarely do they attain a length as great as .1 mm.

Magnetite is very common throughout the rocks except where pyrite is found, in which case magnetite may be sulfidized to pyrite. In the alteration of biotite, magnetite is commonly released. In some

cases, it occurs as "dusty" iron oxide grains which remain in the altered mineral, in other cases some migration and organization into irregular grains outside the original host mineral is demonstrated.

Zircon is a rare mineral in the quartz monzonite. It occurs in biotite, characteristically with a dark brown pleochroic halo. In many sections, a search of the biotite revealed but one or two grains of zircon over an entire thin section. In some alaskites, however, zircon is relatively abundant. This suite of accessory minerals is common to quartz monzonites found at other localities. (Dapples 1940.)

Tourmaline is not as uniformly distributed in the quartz monzonites as the other accessory minerals. (See mineralogy of deuteric stage for details on tourmaline.) Some allanite (?) is indicated. It should be stressed that no systematic attempt has been undertaken in this investigation to exhaust the accessory mineral suite.

#### Textures.

Quartz monzonite is separated into coarse-\*, medium-, and fine-grained textures, the fine-grained facies are normally quartz monzonite-porphry, the coarse-grained rocks are commonly somewhat porphyroblastic. In this section some detailed petrographic data are repeated, because some readers may have skipped the detailed systematic mineralogy and they may want the detailed petrography of the different rocks described.

---

\*Note: Coarse-grained facies of the Boulder batholith are distinguished by 2 mm + average grain size instead of the customary 4 mm. This usage follows the practice of the U. S. Geological Survey field party which is mapping much of the Boulder batholith.



Coarse-Grained Quartz Monzonite. The coarse-grained quartz monzonites have an average grain size 2 mm or somewhat greater. Two sub-types are recognized on the basis of the presence or absence of a conspicuously fine-grained groundmass. In one, large crystals comprise 75 per cent or more of the rock, the remainder is a fine-grained groundmass. These rocks are characteristically gray to dark gray, mottled with pinkish gray and speckled with patches of black mafic grains. The fine-grained groundmass is dark grayish and contributes to make the rocks darker in color than the second type. Figure 58 shows coarse feldspar crystals with much finer-grained groundmass.

The other sub-texture is no more coarse grained, but the groundmass is less distinct, and medium grained. These rocks are commonly lighter in color, distinctly mottled pinkish and gray, representing the feldspars and quartz, and speckled with clusters of black mafic minerals. Biotite is commonly in larger crystals than in the other sub-texture and is more conspicuous in the hand specimen. Potash feldspar porphyroblasts are commonly larger and pinker in these rocks. Typical coarse-grained texture is shown in Figure 59.

The coarsest-grained rocks in the district are in the Jerrico Mountain region, especially on the south side (sec. 2 T. 8 N., R. 6 W.) and in the area east of the Julia Mine (SW $\frac{1}{4}$  T. 8 N., R. 6 W.). Likewise east of the Sure Thing Mine quartz monzonite is quite coarse grained (SW  $\frac{1}{4}$  sec. 14 T. 8 N., R. 6 W.). All these localities are over a mile from any exposed contact of the batholith, however, they all may have been within 1000 ft or so of the roof. (See geologic cross sections on geologic map.)



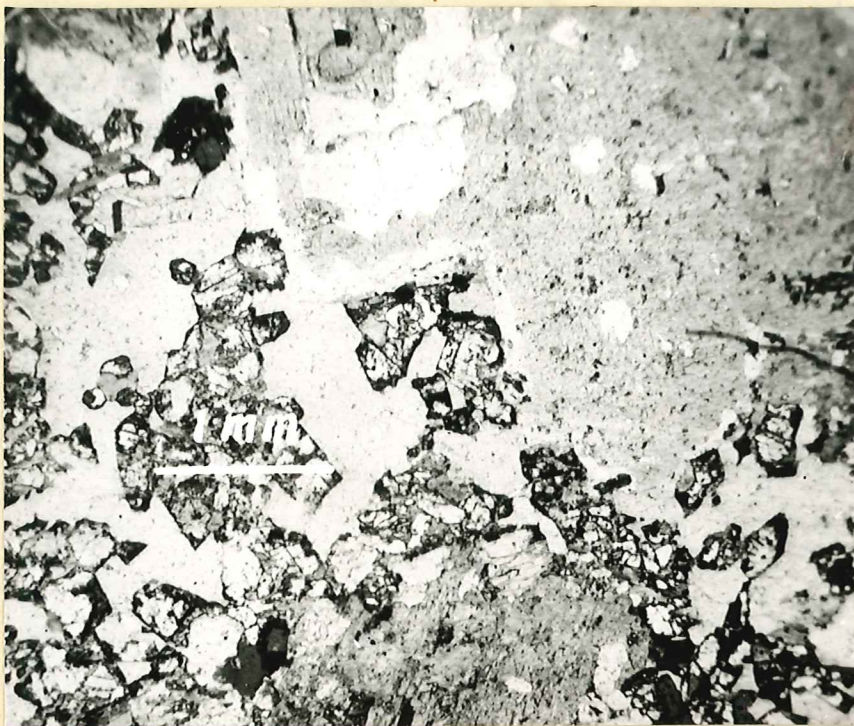


Figure 58. Photomicrograph of coarse-grained quartz monzonite with minor amount of fine-grained groundmass. Gray grains in groundmass stained with sodium cobaltinitrite. (Scale should be 0.5 mm) Plane polarized light.

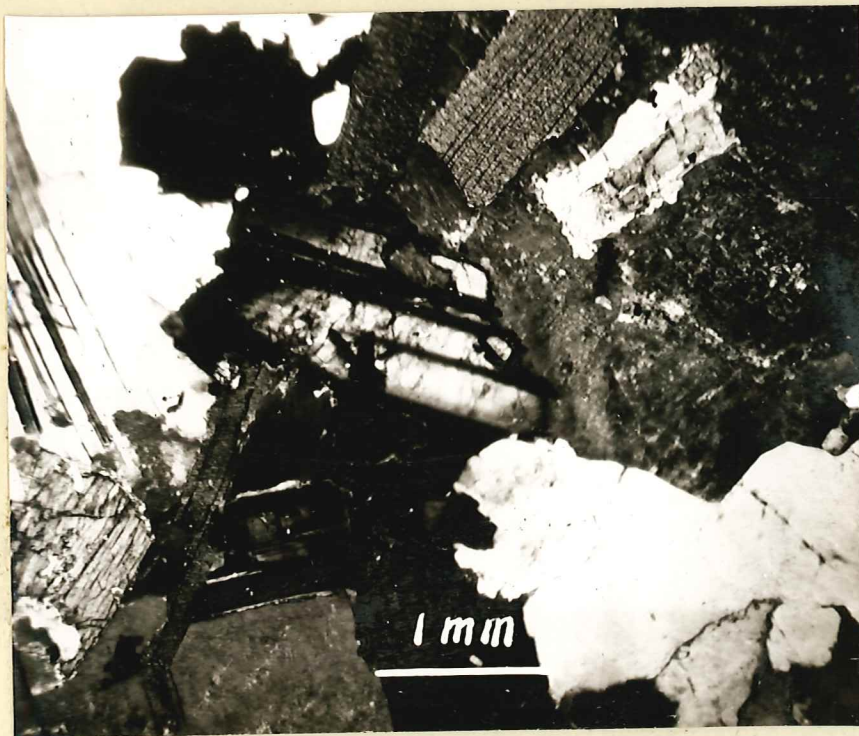


Figure 59. Photomicrograph of coarse-grained quartz monzonite with no fine-grained groundmass. (Scale should be 0.5 mm) Crossed nicols.



South of Jericho Mountain, the coarse-grained rocks have a conspicuous groundmass. The rocks are dark, mottled gray and brownish gray with scattered mafic grains. Porphyroblastic potash feldspars are buff colored and up to 10 mm long, however most are 3 to 5 mm. Gray plagioclase crystals are not conspicuously automorphic. Hornblende, in ragged clusters 3 to 4 mm across, and bright biotite flakes, 0.5 to 2 mm in diameter, comprise the mafic minerals. A trace of pyrite is locally noted.

In thin section, plagioclase crystals are large and formerly were automorphic, however many are partly replaced by potash feldspars and now have irregular borders. Potash feldspar occurs in irregular but well-organized porphyroblastic grains, some of large size, and it is sparsely perthitic. Quartz is found interstitially between feldspars, some grains to 0.5 mm, many poikilitically enclose groundmass grains of finer-grained feldspar. Hornblende, which is mostly actinolitic and contains a few relics of pyroxene, occurs in grains 1 to 2 mm but the margins of the grains are very ragged, in part replaced by biotite and younger feldspar.

The groundmass consists of xenomorphic potash feldspar and quartz and small amounts of hypautomorphic crystals of plagioclase and biotite. Figure 58 shows the texture of the rock.

The approximate modal analysis is as follows:

Coarse Crystals	Large Crystals	Groundmass	Total
Andesine	35	5	40
Potash feldspar	10	20	30
Quartz	10	5	15
Hornblende	10		10
Biotite Magnetite		5	5
	65	35	100

At a locality closer to the Continental Divide, and presumably nearer the top of the batholith, a late flooding of quartz is conspicuous in the groundmass. Hornblende in the rock is strongly altered to chlorite. The groundmass consists of clear xenomorphic grains of quartz and potash feldspar. Porphyroblastic potash feldspar crystals are conspicuous and replace some plagioclase. Near the Sally Placer (center sec. 2 T. 8 N., R. 6 W.) the rock is coarse grained, with quite large potash feldspar porphyroblast (for the Elliston district) 5 mm and larger. Aplite stringers and xenoliths are more abundant in this area than in most areas within the confines of the Elliston district.

About  $1\frac{1}{2}$  mi east of the Julia Mine (SE $\frac{1}{4}$  sec. 4, T. 8 N., R. 6 W.) coarse-grained quartz monzonite has a medium-grained groundmass. The color of the rock, although gray, has a distinctly pinkish cast and contains clusters of black mafic minerals. Potash feldspar occurs in grains 1 to 3 mm with larger porphyroblastic crystals 6 mm long. Some plagioclase crystals are as much as 6 mm, and clearly zoned. The groundmass is medium grained and not conspicuous.



In thin section, orthoclase microperthite and anorthoclase comprise the potash feldspars. Andesine crystals are hypautomorphic, modified from good crystal outlines by replacement of potash feldspars. Biotite occurs in hypautomorphic crystals to 2 mm and hornblende in larger crystals, rarely 5 mm, which are very irregular in outline. Relics of pyroxene are found in some hornblende crystals. About 20 per cent of the rock is medium grained groundmass, illustrated in Figure 59. An approximate modal analysis is:

Andesine	45 per cent
Potash Feldspar	30
Quartz	15
Hornblende	5
Biotite and magnetite	5

East of the Sure Thing Mine (SW $\frac{1}{4}$  sec. 14, T. 8 N., R. 6 W.) the granitic rock is leucocratic and resembles granite. It is light pinkish gray with sparsely scattered clusters of mafic minerals. The average grain size of the feldspars is 2 to 5 mm with some groundmass crystals 0.5 to 1 mm. Biotite occurs in bright flakes 1 to 3 mm whereas hornblende, although present in large crystals, is very ragged in outline. An approximate modal analysis shows:

Andesine	40 per cent
Potash feldspar	25
Quartz	15
Hornblende-biotite- magnetite	10

Medium-Grained Quartz Monzonite. Medium-grained granitic rocks are the most abundant in the Elliston district. The relative dark or light color of the rock, which forms the basis for the subdivision of the medium-grained textures made by the U. S. Geological Survey field party (Becraft 1955) is not so much due to differences in mafic mineral content as the fineness of grain in the groundmass.

The groundmass textures vary in mineralogy and in shapes of the constituent minerals. In some rocks, the groundmass minerals show a moderately wide range in size and are interlocked in typical igneous textures as shown in Figures 41 and 43. Plagioclase and biotite occur with potash feldspar and quartz. In other rocks, the groundmass is composed primarily of potash feldspar and quartz with minor amounts of biotite and plagioclase, in more or less equal granular xenomorphic grains which resemble the texture of aplite. This type of groundmass is described in the text of this report as aplitic groundmass and is illustrated in Figure 60. Actually Figure 60 is from a fine-grained rock but, as will presently be shown, the distinction between medium and fine grain is based primarily on the amount of fine-grained groundmass relative to the coarser-grained minerals. Likewise, the finer the grains in the groundmass, the darker the rock. Most of the rocks are petrographically quartz monzonite porphyry in which the groundmass varies from 25 to 65 percent of the rock.

In addition to the texture, there are mineral differences involving the relative proportions of biotite and hornblende. Hornblende is never absent, but in some rocks it shares prominence with fresh,



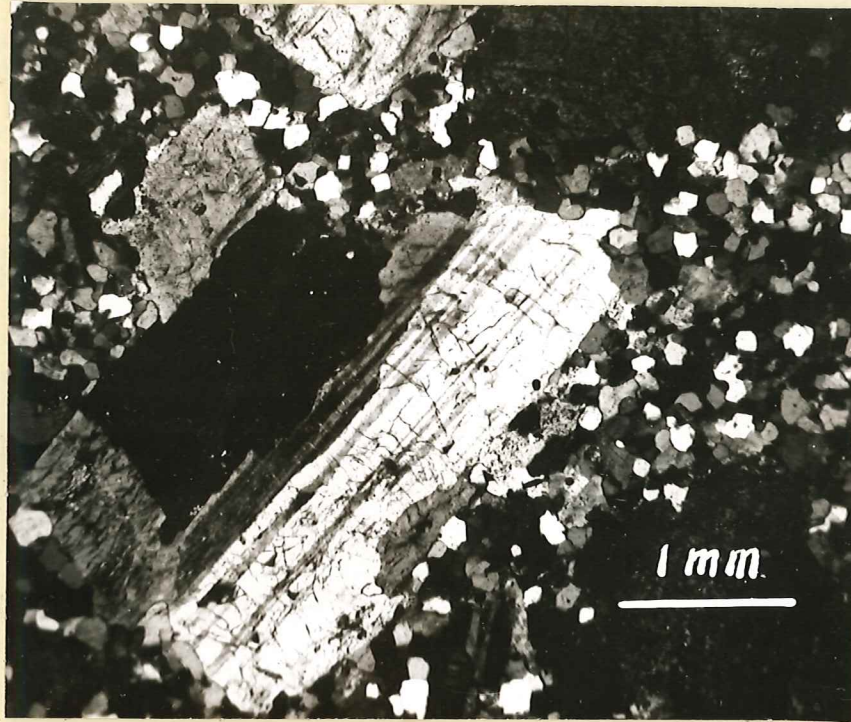


Figure 60. Photomicrograph of plagioclase phenocryst in aplitic groundmass of quartz monzonite porphyry at the Pauper adit. (Scale should be 0.5 mm) Crossed nicols.

lusterous flakes of biotite. In other rocks, biotite is not recognized in hand specimen but is found in small hypautomorphic grains scattered throughout the fine-grained groundmass.

In hand specimen, the medium-grained rocks are gray to greenish gray with patches of black mafic minerals. Some have a mottled appearance and indicate that potash feldspar is present in grains of distinguishable size. Many of the rocks were classified in the field as diorite or diorite-porphphy because no potash feldspars or quartz was recognized in hand specimen.

Probably typical of medium-grained quartz monzonite, is rock found along Telegraph Creek Road in center sec. 12 T. 8 N., R. 6 W. Plagioclase is slightly zoned and of calcic andesine composition. Automorphic grains to 3 mm, normally 1 to 2 mm, have been modified by attack of the late-forming minerals, principally potash-bearing feldspars and quartz. Anorthoclase and orthoclase are both present. In this rock, there is a greater percentage of hornblende than biotite, which in turn is more abundant than augite. The amphibole is light greenish, pleochroic, actinolitic hornblende. Hypautomorphic biotite is strongly pleochroic, dark red brown to tan.

Northeast of the Pauper workings, the medium-grained quartz monzonite has inequigranular groundmass of about 20 per cent. Andesine phenocrysts are zoned and range from 1 to 3 mm. Orthoclase occurs in scattered porphyroblastic grains, rarely to 2 mm. Quartz is confined to the groundmass. Hornblende is actinolitic and ragged in outline, whereas biotite is hypautomorphic but found only in small grains in the groundmass. The approximate modal analysis is:



	Phenocrysts	Groundmass	total
Andesine	30	7	37
Orthoclase	20	8	28
Quartz	10	5	15
Hornblende (with augite)	15		15
Biotite & Magnetite		5	5

Some of the coarser medium-grained rocks are found north of Jericho Mountain along the Continental Divide. Plagioclase crystals to 2 mm and greater comprise about 35 per cent of the rock. Potash feldspars show a very wide range in grain sizes. Some granoblastic grains are rather erratically distributed throughout the rock. Some of the porphyroblasts show optical continuity around several adjoining grains in the groundmass as shown in Figure 43. Biotite is pleochroic, dark reddish brown to tan. Magnetite is found with biotite and actinolitic hornblende, which is pale green in thin section. The groundmass is fine grained 0.1 to 1 mm comprising 10 to 20 per cent of the rock.

On the east side of Jericho Mountain the quartz monzonite is slightly porphyroblastic with large pink potash feldspar crystals to 5 mm in the groundmass which has an average grain size of about 1 mm. Some of the potash feldspar porphyroblasts have granophyric rims, and granophyre makes up possibly 20 per cent of the rock. The additional groundmass is composed of fine xenomorphic grains of potash feldspar and quartz which comprises another 20 percent of the rock. A number of dark xenoliths are present in the rock. The rock weathers to massive, rounded, joint-controlled boulders as shown in Figure 61.



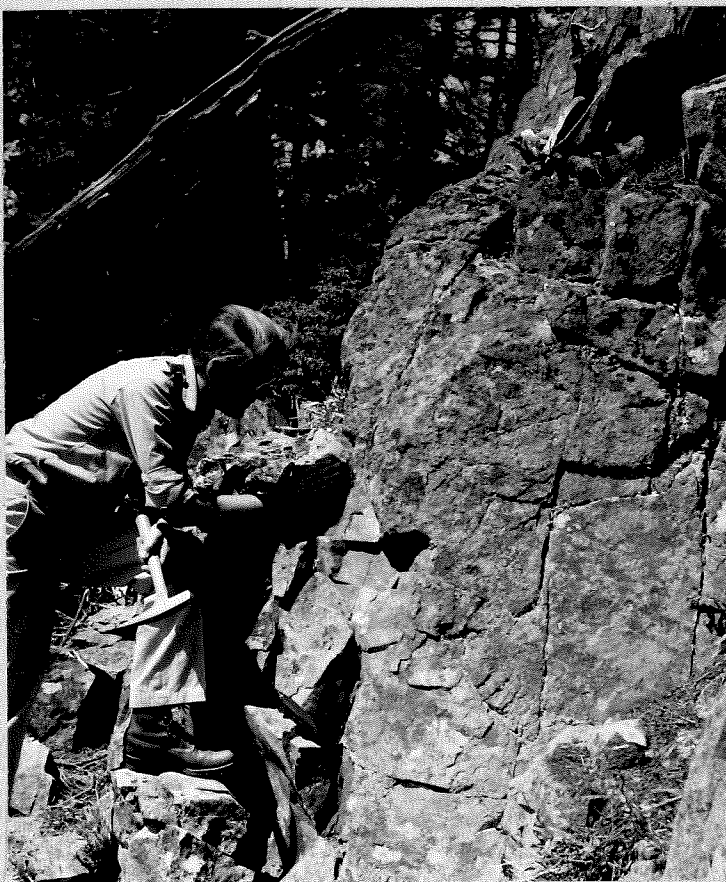
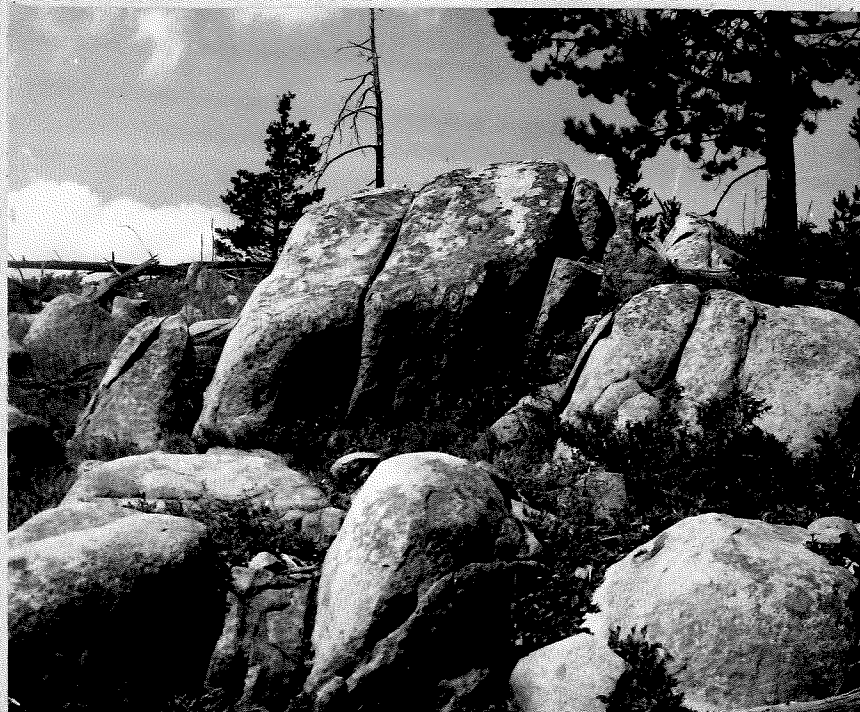


Figure 61. Outcrop of medium grained quartz monzonite on southeast slope of Jericho Mountain. Note the spacing and attitude of the joints.

Figure 62. Outcrop of quartz monzonite porphyry within 10 ft of the top of the intrusive north-east of the Pauper adit. Dr. R. W. Chapman is examining an intrusive aplite dike.



Northeast of Nigger Mountain along the contact with the volcanics, the percentage of groundmass in the medium-grained quartz monzonite is great. It is more than 50 per cent in some cases. Plagioclase crystals are 1 to 2 mm and in the groundmass some porphyroblastic potash feldspar crystals obtain a diameter of about 0.5 mm. A late flood of quartz is conspicuous--a common feature immediately below some contacts.

Downhill from the Ontario Mine in the south central part sec. 22 T. 8 N., R. 6 W., medium-grained quartz monzonite is strongly altered deuterically. Plagioclase has gone to clinozoisite and sericite, and mafic minerals altered to chlorite, variety penninite, with released dusty iron oxide grains.

East of the Lilly Mine on the Continental Divide, early plagioclase crystals are particularly abundant, potash feldspar is represented by reorganized grains in the groundmass and in late granophyric intergrowths. The plagioclase composition is about Ab 55 An 45. Biotite is present as hypautomorphic grains, and a small amount of automorphic pyroxene occurs in the rocks. Some of the pyroxene is quite fresh, some is altered to uralite. Relics of olivine are noted. Orthoclase microperthite and granophyre are conspicuous constituents. Anorthoclase is abundant, but does not enter into the granophyric intergrowths. The late quartz is slightly biaxial,  $2V$  is a few degrees.

In the extreme northeast corner of the district, on the Continental Divide, the quartz monzonite contains about 30 per cent phenocrysts of calcic andesine up to 3 mm in length. The groundmass comprises 30 per cent or more of the rock of which a small percentage is very fine-grained granophyre. The plagioclase grains in the groundmass are

strongly sericitized, whereas the large phenocrysts are not affected or very slightly altered. Hornblende is more abundant than biotite. Some of the larger orthoclase crystals are not obviously porphyroblastic, however, the very erratic size range of the crystals and the presence of only a few large crystals suggest that they are porphyroblasts and not phenocrysts.

Along the Continental Divide, at a locality which is probably close to the contact, the quartz monzonite shows a flood of late quartz which replaced minerals in the groundmass and to some extent the early-formed plagioclase phenocrysts. The rock contains about 20 per cent quartz. Biotite is minor in amount; hornblende is conspicuous in elongated hypautomorphic grains. No zircon was noted in the rock whatsoever.

Fine-grained quartz monzonite. (Quartz monzonite porphyry.) The fine-grained rocks are virtually always found at or very near the margin or roof of the batholith, however some medium-grained rocks are found at the contacts. Figure 62 shows an outcrop of fine-grained quartz monzonite porphyry less than 10 ft below the roof of the batholith. The fine-grained rocks are characteristically composed of 50 per cent or more fine-grained groundmass in which automorphic plagioclase phenocrysts, and minor amounts of mafic phenocrysts (commonly ragged hornblende, and in some cases, automorphic pyroxene) are found. Biotite, potash feldspar and quartz are normally confined to the xenomorphic groundmass. Some porphyroblastic potash feldspar crystals are found, but rarely are they large enough to be easily identified in hand specimen.



In the deuterically altered fine-grained quartz monzonite porphyries, plagioclase phenocrysts are differentially sericitized, one grain may be relatively strongly attacked, whereas another crystal only a millimeter away may be completely fresh or only slightly altered. Alteration to clinozoisite is common, along with albitic alteration, in patches and along fractures. In these rocks, the mafic minerals are variably altered to chlorite; normally released opaque iron oxide remains as small grains in the area of the original mineral. Apparently the energy of the deuteritic solutions was not competent to release the iron from its host minerals and organize it into larger grains. Epidote and leucoxene are usually found among the alteration products of the mafic minerals.

In the Pauper-Julia contact area, the fine-grained rocks show a considerable amount of xenomorphic aplitic groundmass and a large amount of micrographic intergrowths. Figure 60 shows a plagioclase phenocryst in aplitic groundmass. Potash feldspar porphyroblastic growths are rare, plagioclase occurs in phenocrysts 1 to 3 mm. In some rocks, fresh augite is present, and in some, relics of olivine are found in clusters of altered mafic grains.

In the fine-grained rocks north of Jericho Mountain, most of the essential minerals crystallized in the groundmass including plagioclase and biotite, as well as quartz and potash feldspar. The groundmass comprises as much as 75 per cent of the rocks in which case they are dark greenish gray in color. Plagioclase phenocrysts, 1 to 2 mm, and ragged hornblende phenocrysts are found in a groundmass composed of grains from 0.1 to 0.5 mm. Only rarely are potash feldspar grains

organized into small porphyroblastic growths, however at one locality, optical continuity of potash feldspar in the groundmass was recognized in poikiloblastic grains in areas 1 to 2 mm across. In some of the rocks, the biotite is sprinkled in small, well-formed crystal grains throughout the groundmass. Some alteration to chlorite is common.

#### Aplite and Alaskite

Aplite is found in minor amounts in the district, and alaskite in two bodies, one in the Ontario Mine area in the southeast part of the district, and the other in the Telegraph Mine area 3 mi northeast of the Ontario Mine. The combined outcrop area is less than 1 sq mi.

#### Definition.

Aplite is derived from the Greek word for "simple" and according to Johannsen (1949, p. 91) was first used by Retz. Von Leonard is reported to have used the term for fine-grained granites which are without mica, and composed only of feldspar and quartz. (Johannsen 1949, p. 91.) Rosenbusch (1907, p. 579) used the term in a somewhat broader sense. Brögger (1898, p. 212) considered them diaschistites (dikes or border facies) with very little dark constituents (Johannsen 1949, p. 92). Vogt (1931, p. 51) noted that the most siliceous dikes, where the silica content is 75.3 per cent or better, show distinct saccharoidal textures, indicating simultaneous crystallization of quartz and feldspar, but dikes with less silica, 73 to 74 per cent, exhibit a slight tendency for formation of phenocrysts in saccharoidal groundmass.



Alaskite was named by Spurr (1900) for "holocrystalline granular plutonic rocks characterized by essential alkali-feldspars and quartz, and little or no dark component. . . In the description of the original alaskite from Alaska, the feldspar was given as "mostly orthoclase, but some albite . . . verging towards anorthoclase." (Spurr 1900, p. 229-30) (Johannsen 1949, p. 106).

#### Structural Control of Aplite-Alaskite.

In the Boulder batholith, Weed (1912) and Billingsley (1916) consider that aplite-alaskite is generally found near the roof of the batholith, and Billingsley and Grimes (1918) specifically point to what they consider to be a close genetic relationship between hydrothermal veins and aplite.

Detailed mapping in the northern part of the Boulder batholith by Klepper and his colleagues on the U. S. Geological Survey failed to demonstrate a close spacial relationship between aplite and the roof, or aplite and ores, although locally, aplite and alaskite are found in relatively large bodies under flat parts of the roof of the batholith, and some aplite dikes have the same structural trend as the veins and may occur along one or the other walls of a hydrothermal vein. (Klepper and Robertson 1956.)

Most aplite and alaskite occurs in a broad north-northeast trending zone that is parallel to the east margin of the batholith, but which is several miles removed from the east border. Most of the mineralized veins are found west of the main belt of aplite-alaskite.

Klepper (Klepper and Robertson manuscript draft pp. 3 and 4, 1956) states that "most of the aplites are sheets, dikes, and irregular-shaped bodies of devious and variable trend, probably occupying fractures that formed as a result of local stresses during crystallization. In contrast, most of the veins occupy strong, steeply dipping east-west fractures, some of which are several miles long. These fractures belong to a regional set that is unrelated to the cooling history of the batholith and was impressed on batholithic and wall rocks alike after the exposed part of the batholith, including the bulk of the aplite, had consolidated. . . Nevertheless, a number of observations strongly suggest that the late magmatic stage graded into the hydrothermal stage almost always as imperceptibly as the early magmatic stage graded into the late magmatic stage."

Only locally in parts of the batholith is there a distinct tendency for aplite-alaskite to be concentrated at the roof, particularly the flat top of the batholith. At two localities in the Elliston district, alaskite is found at the roof of the batholith. Similar relations were observed by Weed (1901) in the Elkhorn district on the east side of the batholith and by Klepper (personal communication) in the northeast part of the batholith.

The occurrence of aplite at the margin of granitic intrusives, especially under flat roofs, has been noted by numerous investigators. Daly (1933) compiled information on the subject. He specifically cited aplite apophyses of the monzoni granite (Huber 1901) and illustrates concentrations of aplite at the roof at Nickel Plate Mountain in British Columbia (Camsell 1901, p. 101). Richey (1937, p. 675)



describes a considerable mass of aplite at the contact under flat contacts in the Mourne granite in northern Ireland, and Walker and Mathias (1946) report that the aplite appears to have intruded along the contact of the intrusive granite near Cape Town. In his discussion of the aplites in the Front Range in Colorado, Lovering states: "The aplitic rocks occur chiefly in dikes, but near the west border of the batholith they comprise moderately extensive irregular bodies. The aplite grades to fine-grained granite and alaskite." (Lovering 1935, p. 11). This relationship noted by Lovering is considered to be similar to the situation of alaskite found near the Ontario Mine in the Elliston district. In addition, Wright (1915) and Nockolds (1940) describe concentrations of aplite at the margin of the intrusive granitic bodies in Alaska and northern England, respectively.

The relations of aplites and alaskite bodies to other primary structures has not been demonstrated in the Elliston district, or in the batholith as a whole. The aplite dikes which have the same trend as the hydrothermal veins are considered to be tectonically controlled.

Types of Aplite and Alaskite. In the Elliston district, magmatic aplite bodies are few; however, alaskite and rocks of aplite composition (altered volcanic rocks) occur at the top of the batholith and are locally abundant. Several types of aplite and alaskite are recognized.

1. Intrusive dikes in quartz monzonite, rarely in volcanic rocks. Most are 1 to 2 in, but they range from  $\frac{1}{4}$  in to several feet in width.

2. Small, irregular dikelets. The aplitic material is derived in part from the walls, and in part by replacement of the wall rocks of small fractures.
3. Small pods of aplite in quartz monzonite  $\frac{1}{2}$  in to several feet across. Aplite formed by segregations of late magmatic aplitic solutions and/or by metasomatism.
4. Alaskite and aplite replacement bodies of quartz monzonite at the roof of the batholith. These bodies may be 50 to 100 ft thick and occupy an area of  $\frac{1}{2}$  sq mi or more.
5. Contact metamorphism in volcanic rocks to microaplite composition. This type is described under contact metamorphism.
1. Intrusive aplite dikes are common in the batholith, but rarely are they large enough to be shown on geologic maps. They are commonly 1 in to a few in in width, rarely are they as much as 1 ft across.

Figure 63 shows a typical intrusive dike.

In the Elliston district, only a few intrusive dikes are recognized. Very fine-grained aplite dikes occur along one wall of the Sure Thing and Moonlight veins. Several aplite dikes are found in the Lilly-Sure Thing area generally. At the Pauper, several intrusive dikes were found, likewise dikes in quartz monzonite are found in outcrops along the Continental Divide south of Bullion Park.

The dikes normally have sharp borders and show a wide range in textures, even across dikes only 1 in or so wide. Pegmatitic pods and miarolitic cavities lined with coarser-grained quartz and potash feldspars, with or without pyrite and tourmaline, are common features. The textures are complex grading from microgranitic to microstural to





Figure 63. Intrusive aplite dike in quartz monzonite. Dike exposed in highway cut south of Basin, Montana and not actually in the Elliston district.

granophyric. Figure 64 is the texture exhibited in a narrow intrusive dikelet. Figure 65 shows a saccharoidal texture in aplite.

2. Small irregular dikelets are not common but are found in the district. In individual cases, very small stringers a few inches long and not over  $\frac{1}{4}$  in wide demonstrate the local derivation of the aplitic materials. The quartz monzonite porphyry contains plagioclase phenocrysts with abundant aplitic groundmass, whereas the dikelet contains only a few plagioclase crystals in aplite. There is no distinct margin distinguishable microscopically. As the fracture formed, the aplitic groundmass migrated into the opening, probably before the groundmass of the wall rocks had more than started to crystallize, and rafted early-formed plagioclase crystals into the dikelet.

In another small stringer only  $\frac{1}{8}$  in wide from the top of the batholith south of the Julia Mine, one end of the thin section shows replacement granophyre in the dikelet, and at the other end, the texture is microgranular, saccharoidal.

Figure 66 is a photograph of an irregular aplitic stringer which shows local replacement textures and evidence of local derivation of aplitic materials from the walls. Just such a zone of fractures, formed at the most favorable time in the crystallizing quartz monzonite, might well serve as collecting channels for intrusive aplite dikes.

3. Small pods of aplite, many of which are irregular, range from  $\frac{1}{2}$  in to several feet in diameter. They are most abundant near the top of the batholith, although aplitic patches are found throughout the quartz monzonites, principally in the medium- and coarse-grained facies. Some small pods consist of intricate granophyric patches, others





Figure 64. Photomicrograph of texture of aplite in intrusive aplite dike near Julia mine. (Scale should be 0.1 mm.) Crossed nicols.



Figure 65. Photomicrograph of saccharoidal texture of intrusive aplite. Crossed nicols.





Figure 66. Outcrop of quartz monzonite showing thin aplitic stringer. Note that the stringer frays out and is lost at the left end, and is a sharp, through-going aplite diklet at the right side of the photo. Outcrop in highway cut south of Basin, Montana.



Figure 67. Pod of replacement aplite in quartz monzonite. This outcrop is in the old townsite at the Julia mine.



contain quartz and potash feldspars in intricate replacement textures. In thin section the original mafic minerals in the rock are eliminated in the replacement to aplite. Relics of mafic minerals are indicated by patches of white mica or chlorite, with traces of leucoxene. Figure 67 shows one of the large replacement pods of aplite in quartz monzonite. Small "horses" of unreplaced quartz monzonite are found throughout the body. Some aplite pods have rather sharp borders.

4. Replacement aplite and alaskite bodies occur in the quartz monzonite at the top of the batholith at two localities within the Elliston district as shown on the geologic map. At the Ontario Mine, aplite is associated with large bodies of alaskite, both of which show relics of the minerals found in the original quartz monzonite, and at one locality, strongly deuterically altered quartz monzonite is found in relic dikes within aplitized volcanic rocks.

The replacement textures are distinctive, particularly in alteration of plagioclase. In some instances, relics of the original plagioclase are preserved in the cores of potash feldspar as shown in Figures 50 and 52; in other instances the plagioclase is altered to a more sodic plagioclase; in other instances the plagioclase is replaced by anorthoclase which has a distinctive mottled microclitic grid twinning; and in still other instances plagioclase is altered to patches of sericite. In some cases, the mafic minerals are eliminated, with only patches of sericite or muscovite preserved.

Tourmaline is present in aplite and alaskite in small radiating clusters in small vugs and intergrown with feldspar and quartz.

Mineralogy. The mineralogy of aplite and alaskite is quite simple. Quartz and potash feldspars and subordinate plagioclase are the essential minerals. Tourmaline is abundant in some aplite; in lesser amounts are sericite, epidote, sphene, apatite, zircon, chlorite and actinolitic amphiboles. Magnetite, pyrite and traces of molybdenite are the ore minerals found in aplite.

(Potash feldspars). Although potash feldspars are the most abundant minerals, their mineralogy is not simple. There is strong indication that much of the microclitic-appearing minerals may be anorthoclase. Many of the larger potash feldspar crystals which are porphyroblasts are strongly perthitic. (See mineralogy section under Quartz Monzonite.)

(Tourmaline). Tourmaline is abundant in some aplites and alaskites, both in replacement and intrusive types. In the rocks of the Boulder batholith, tourmaline is black in hand specimen, pleochroic from olive green to dark blue green in thin section. Tourmaline is a common mineral in many aplites and alaskites found throughout the world. The literature has not been exhausted, yet from an incomplete survey, 13 localities have tourmaline in aplite.

Polgeiter (1950) described the occurrence of tourmaline in an aplite in South Africa. He states that it "concentrates in layers in the fine aplite to give the rock a banded appearance". The quotation adequately describes some aplite found near the Ontario Mine in the Elliston district. "Rosettes of tourmaline 3 mm in diameter are present on fracture surfaces, but none are intergrown with the minerals of the rock." (Loughlin and Koshmann 1942) describes relations of some of the



tourmaline occurrences in Colorado, but it is equally applicable to the occurrences of some aplite and alaskite in the Elliston district. Brammall (1942) reports three per cent tourmaline in aplogranite in the Dartmoor series; and Walker (1946) reports two per cent tourmaline in aplite and granite near Cape Town. Many of the aplites of the Boulder batholith appear to have comparable quantities of tourmaline. Worth (1920) reports tourmaline along joint faces in aplite of the Dartmoor similar to numerous occurrences of joint coatings found in the northern part of the Boulder batholith. Willborn (1933) reports tourmaline in spherulitic patches to 20 mm in non-porphyrific aplite in the Federated Malay States which are probably very similar to patches found in the quartz monzonite generally. Nolan (1935), Butler (1913), Parkinson (1907), Osman (1937), Gardner and Reynolds (1931), Scrivenor (1914) and Williamson (1935) have noted the presence of tourmaline in aplite. Undoubtedly there are many other localities where tourmaline is found in aplite, however, the search for such information is rendered difficult because aplite, although mentioned in geological reports, is rarely mentioned in the indexes.

(Sulfides). Sulfides in aplites are not commonly noted. In the northern part of the Boulder batholith, pyrite occurs in aplite and alaskite, and is relatively common. Some molybdenite is also present.

Park (1943) reports pyrite and epidote clusters in aplite in the Metaline district in Washington. Chalcopyrite and molybdenite in aplite are reported from the San Francisco mining district of Utah. "Both sulphides appear to be original constituents of the rock", according to Butler (1913, p. 62).

### Alaskite Occurrences.

Alaskite is found in two localities in the Elliston district, one at the roof of the batholith at the Ontario Mine (N  $\frac{1}{2}$  sec. 22, and part of NW  $\frac{1}{4}$  sec. 23, T. 8 N., R. 6 W.); and in a similar position (in NE  $\frac{1}{4}$  sec. 11, T. 8 N., R. 6 W.) just southeast of the Telegraph Mine. In the wall rock of a vein north of Jericho Mountain on the Continental Divide just south of the center of the section in sec. 36, T. 9 N., R. 6 W. a leucocratic granodiorite was observed.

The two main bodies of alaskite in the Elliston district are petrographically kalialaskite (115 Johanssen) or Alaskite (116 Johanssen) based on the mineralogical composition.

Ontario Mine Area. At the Ontario Mine, the largest body of alaskite is more than 1 mi long, and as much as  $\frac{1}{2}$  mi wide and possibly as much as 100 ft thick. It is complexly grained leucocratic rock which has a wide range in textures from aplite to pegmatite. Tourmaline is commonly found: (1) in pods and small clusters, (2) in disseminated grains, and (3) in quartz-tourmaline vein-like structures in the alaskite. Some of the best exposures of alaskite are in the open cuts of the Ontario vein and in outcrops just south of the vein. In general, the exposures of alaskite are poor. As no alaskite is found on the dump of the Ontario tunnel, situated about 200 ft below the vein outcrop, and as no quartz monzonite crops out above the portal of the tunnel, the alaskite appears to be a lens or layer of leucocratic rock above the quartz monzonite and below the roof.



In thin section, the rock is seen to be composed of xenomorphic grains of potash feldspars and quartz, with minor amounts of albite-oligoclase and relics of mafic minerals. Orthoclase microperthite has an optic angle  $2V = 60^\circ \pm 2^\circ$ ,  $\alpha X = 9^\circ$  in  $010$ . Conspicuous irregular albite blebs are oriented parallel to  $100$  cleavage. The perthites are thought to be reorganization perthites. (See mineralogy under Quartz Monzonite.)

Another potash feldspar contains conspicuous grid twinning of microcline and has a large optic angle. In the wall rocks of the Ontario Mine, the microcline-like mineral has a positive optic sign,  $2V$  approx.  $85^\circ$ . The mineral may be isomicrocline (Winchell 1951, p. 364, Luczizki 1905, p. 347).

Anorthoclase contains indistinct patches of fine grid twinning, has a mottled appearance, and is found characteristically replacing plagioclase feldspar. The optic angle is distinctly smaller than the orthoclase perthite, about  $50^\circ$  or less. Both orthoclase microperthite and microcline replace anorthoclase, and these are likewise replaced by late quartz. In most instances, anorthoclase is an intermediate step in the replacement of plagioclase by orthoclase, however, in most of the alaskite very little plagioclase remains, and virtually all anorthoclase occurs in minor amounts in relics in orthoclase. In a few instances, surprisingly large grains, which have the size and shape of plagioclase phenocrysts found in quartz monzonites generally, have the composition of anorthoclase.

In other plagioclase relics, quartz-feldspar granophyric intergrowths replace and take the shape of large plagioclase grains. Some myrmekite is present, usually associated with late quartz replacement of plagioclase. The different ways in which plagioclase is replaced by potash feldspars is described and illustrated in the section on potash feldspars under mineralogy of quartz monzonite. An additional complex series of potash feldspars, four species, are shown in Figure 68.

At the eastern end of the alaskite body on the south slopes of a small knob on the Continental Divide (cen. NW  $\frac{1}{4}$  sec. 23, T. 8 N., R. 6 W.), altered dikes of quartz monzonite are intrusive into volcanic rocks. The granitic rocks in the dikes have been partly changed to alaskite. In thin section, the texture of the quartz monzonite is preserved, but the mineralogy approaches an alaskite and is considered to represent an intermediate step in the development of alaskite by replacement of the quartz monzonite. Quartz, potash feldspar and minor amounts of biotite comprise the typical groundmass assemblage, however, micrographic intergrowths abound. Large plagioclase crystals are in part replaced by anorthoclase, a particularly striking example is shown in Figures 51, 54 and 55. Note that different parts of the relic grains have optical continuity indicating that it was all part of a single plagioclase crystal in Figure 52.

Telegraph Mountain Area. Southeast of the Telegraph Mine, alaskite crops out in the walls of a vein and in a localized area below the roof rocks of the batholith. The outcrops in the area are sparse so that the areal extent of the body is not known in detail. Orthoclase





Figure 68. Photomicrograph of plagioclase (left, very dark) partly replaced by anorthoclase (central, medium gray). Note untwinned, lighter rim of albitic composition at margin of plagioclase and sericite alteration in plagioclase at lower left. Three other potash feldspar crystals are shown, each with a somewhat different texture. Grain at upper right is perthitic, probably orthoclase. (Scale should be 0.5 mm) Crossed nicols.

microperthite, anorthoclase and potash feldspar and quartz micro-graphic intergrowths comprise 90 per cent of the rock. Minor amounts of albite (Ab 90 An 10), relics of biotite altered to shreds of chlorite with included dusty opaque grains, sphene-leucoxene and zircon comprised the accessory minerals.

North of Jericho Mountain, a leucocratic granitic rock occurs along the south wall of an east-west vein. The rock is a leucogranodiorite (127) in which plagioclase is the dominant feldspar and mafic minerals comprise less than 5 per cent of the rock. Plagioclase, which has the shape and texture of normal quartz monzonite plagioclase, has a distinctive broken-up appearance, and it is strongly altered to oligoclase. Orthoclase microperthite, a non-perthitic potash feldspar and anorthoclase are all found in the rock. Quartz is a conspicuous and late mineral, filling between grains, and replacing some feldspar. Biotite in well-formed hypautomorphic grains is clear, pleochroic in red-brown to tan, and only slightly altered to chlorite and minor amounts of magnetite. Magnetite and biotite do not comprise more than a few per cent of the rock. The nearest outcrops are normal quartz monzonite, whereas the south wall of the vein is leucogranodiorite. The position of the vein may be fortuitous with respect to the rock type, or it may represent deuteric alteration along a zone which was later fractured to produce the vein zone.

#### Magmatic Phenomena Leading to Metalization

In each of the main rock types described, there is a sameness in the crystallization histories. In granogabbro, granodiorite and



quartz monzonites, there is an orderly sequence of crystallization of the early magmatic minerals. The sequence of formation and mode of origin of the late magmatic minerals, mainly potash feldspars and quartz, exhibit a wide range of features, however, in a given structural environment, the sequence of formation and the modes of origin are the same.

In granogabbro, late magmatic or deuteritic potash feldspar growths and leucocratic alkali feldspar stringers are conspicuous, although minor, features. Basic granodiorite (not described in this report because it is not found in the district) shows even more conspicuous late magmatic features including some aplitic and pegmatitic stringers and leucocratic vein-like bodies. In the Bison Mountain granodiorite, strong alteration of the rock by deuteritic processes has been mentioned. In quartz monzonite, however, late magmatic phenomena are abundant, leading to the formation of large bodies of aplite and alaskite. Still further, late magmatic or hydrothermal phenomena are responsible for the ore minerals in the veins and the wall-rock alteration along the veins.

In this section, the crystallization sequence leading to the formation of aplite-alaskite and to the formation of the hydrothermal veins are discussed relative to the late crystallization history of quartz monzonite. Magmatic crystallization and differentiation is discussed, followed by the deuteritic stage and deuteritic differentiation of aplite-alaskite. The steps leading to vein formation, which necessarily adds another dimension to the mineral formation sequence, namely fracturing, are traced from the deuteritic into the hydrothermal

stage. The detailed discussion of the mineralogy and structure of the mineral deposits are described under the heading, "Mineral Deposits."

#### Magmatic Differentiation

Marked chemical and mineralogical differentiation is not conspicuous in quartz monzonite, with the exception of the formation of aplite-alaskite in the late magmatic or deuteric crystallization stage. The different types of quartz monzonites in the main part of the Boulder batholith are, in large part, due to differences in textural facies, and secondarily to minor differences in the proportions of the essential minerals.

Only at one locality within the Elliston district, and so far as is known, in the batholith at large, are marked variations in the mineralogical assemblage observed in a vertical section within the granitic rocks. This is in the vicinity of the Pauper prospect (cen. S  $\frac{1}{4}$  sec. 8, T. 8 N., R. 6 W.) in a thin sliver of granitic rocks which I think represents a sill-like lens of the margin of the batholith. The mineralogical variations do not appear to be related to separate intrusives, but formed as a consequence of fractional crystallization within the 200 to 250 ft thick body of quartz monzonite magma. The rock types vary from olivine-bearing granogabbro porphyry to fine medium-grained quartz monzonite. The overlying volcanic rocks were altered to rocks of aplitic composition by the intrusive, a distinctive type of contact metamorphism in the region. The mineralogical and textural variations in the rocks at this locality are described in some detail.



At the Pauper area, four specimens were collected in a more or less vertical section of about 200 ft. The lowest is from the dump of a small prospect, the poral of which is in a dark, hornfelsic volcanic rock, however dark gray, olivine-bearing granogabbro porphyry (basic quartz monzonite) was encountered in the tunnel. Large, obviously early, labradorite crystals (Ab 40 An 60) 1 to 3 mm long and pyroxene crystals, with some olivine, in aggregates 2 mm in diameter comprise the essential early minerals. The plagioclase crystals exhibit a distinctive internal cracking which is common of the plagioclases of the rocks of the Boulder batholith generally, as well as in gabbro and early volcanic rocks. The plagioclase crystals commonly contain conspicuous "cores" which contain a variety of fine mineral grains including augite, pigeonite, biotite, apatite and magnetite. The outer part of the crystals are clear and free of inclusions, as shown in Figure 36. Augite has a moderately large optic angle, whereas pigeonite has a small angle,  $2V$  about  $20^{\circ}$  to  $30^{\circ}$ .

Clusters of pyroxene grains in which augite is the most abundant mineral, but smaller, rounded grains of pigeonite are present with a few crystals of olivine, are shown in Figure 69. Augite is quite fresh and contains numerous small inclusions. The mineral alters to secondary hornblende which is pleochroic, bluish olive green to light bluish green. Pigeonite is found in smaller, equidimensional crystals, usually included in, or distinctly older than, augite, although less in amount than augite. Small opaque laths, possibly hematite, are oriented in pigeonite.

Olivine in small, equidimensional crystals, does not show distinct cleavage, but does have conspicuous curved fractures. Some of the olivine is strongly altered, other grains are remarkably fresh with alteration to antigorite only along fractures. No opaque mineral alteration was noted.

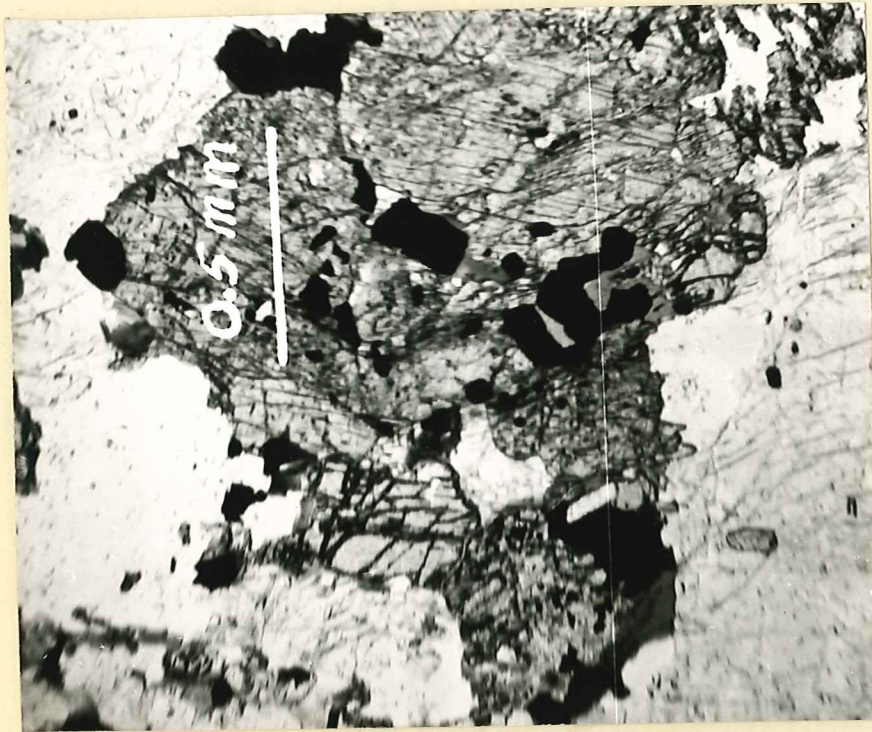


Figure 69. Photomicrograph of olivine and magnetite in cluster of pyroxene grains in basal part of quartz monzonite porphyry sill at the Pauper area. (32-98-2)  
Plane polarized light.



The relief of the olivine is somewhat greater than pyroxene and is optically negative,  $2V$  about  $80^\circ$ . The total amount of olivine is about 1 to 2 per cent of the rock.

Biotite is strongly pleochroic,  $X = \text{tan}$ ,  $Y = Z = \text{dark reddish brown}$ . It is a late mafic mineral, associated with pyroxene and with clusters of large grains of magnetite and apatite. Biotite also occurs in the groundmass with quartz and potash feldspar. Only minor amounts of chlorite are recognized. Sphene is present in small amounts. No zircon was observed in the thin sections of this rock.

The groundmass is made up of xenomorphic potash feldspar and quartz grains, and minor amounts of biotite. The grains are clear, and have sharp, well-defined boundaries. Some of the potash feldspar grains show crystalloblastic growths in the groundmass to grains several times the size of the average groundmass grains, but no large porphyroblastic grains were noted.

About 50 ft vertically above the base of the sill-like body, a dump in front of a second caved portal of an adit yielded fresh granitic rock. The prospect is well below a prominent cliff which is uphill to the northwest of the prospect. The rock is a fine medium-grained dark greenish gray quartz monzonite porphyry in which the plagioclase is similar to the normal plagioclases of quartz monzonite and has a composition about Ab 50 An 50. Inclusions are found in the large plagioclase crystals as in the rock at the lower prospect, but for the most part, the former pyroxene inclusions are altered to secondary minerals. The essential pyroxene in the rock is part fresh but is mostly altered to hornblende which is pleochroic, light bluish green to bluish olive green. Relics of olivine (?) are thought to be present as shown in Figure 31. The relic grains



are nearly round patches which have the shape and size of olivine in the lower rock and are composed of a yellowish green alteration product and opaque iron oxide. Biotite is pleochroic, X = tan, Y = Z = reddish brown. In the groundmass, potash feldspar and quartz grains are in part clear, xenomorphic and equigranular, but many orthoclase grains are micropoikiloblastic with quartz chadacrysts. The margins of some of the groundmass grains are sharp, but the poikiloblastic grains are strongly sutured. Some granophyric intergrowths of feldspar and quartz are in the groundmass.

The next highest specimen is a dark greenish gray quartz monzonite porphyry collected from the base of a 50 ft cliff northeast of the Pauper adit. The texture of the rock is somewhat finer grained than the rocks already described. Plagioclase occurs in automorphic crystals to 2 mm and which are internally fractured and have the composition of andesine, about Ab 52 An 48. Pyroxene relics occur in hornblende, and biotite is present. The groundmass is composed of strongly sutured xenomorphic grains of potash feldspar and quartz, in part micropoikiloblastic and somewhat perthitic.

At the adit of the Pauper claim, the quartz monzonite is a typical medium-grained mottled pink and greenish-gray rock which contains abundant aplite in small dikes and irregular replacement bodies. The top of the intrusive is not over 50 ft above the portal of the prospect which is near the base of a talus pile made up dominantly of aplitized volcanic rocks. The plagioclase was probably calcic andesine but is now mostly sericite. Pyroxene occurs in relics in hornblende, however a few grains are remarkably fresh, similar to pyroxene in the lowest rock described. Biotite is present and a single relic of olivine in hornblende was noted in thin section. The groundmass is much finer grained than the lower



rocks and is aplitic.

The mineralogical changes from bottom to top of this sill-like body are summarized as follows: plagioclase ranges in composition from labradorite to calcic andesine to sericitized andesine; olivine is present only in the lowest part of the intrusive body and is represented by a few relics higher up; two pyroxenes, augite and pigeonite, occur at the base but give place to hornblende with included relics of augite higher up; orthoclase occurs in small xenomorphic crystals in the groundmass at the base of the intrusive, but in the upper part, larger porphyroblastic grains are abundant.

At the base of the intrusive, the plagioclase crystals are clear, with included grains of clear mafic minerals, however in the higher rocks, although there are cores in plagioclase crystals which contain inclusions, the inclusions are altered to hornblende and chlorite. At the top, the plagioclase is strongly altered to sericite, which is probably a hydrothermal alteration produced by the latest magmatic solutions.

The overlying volcanic rocks are replaced to sericite aplite for possibly about 50 ft above the intrusive. Although the contact metamorphism is discussed under another heading, it may be desirable to point out that the groundmass materials in the quartz monzonite, of aplitic composition, were in part released from the intrusive to form aplite dikes. Some of the late "aplitic solutions" may have been responsible for the alteration in the volcanics.

The mineralogical differences which produce the differences in rock composition are partly due to early magmatic processes or magmatic differentiation, but are in large part due to late magmatic or deuteric differentiation.



## Deuteric Crystallization Stage

Introductory Statement. The late magmatic crystallization features have been of special interest to me and so far as they relate to the genesis of mineral deposits, they are important. The essential mineralogy of the deuteric stage is discussed under the heading "Mineralogy" under "Quartz Monzonite." The mineralogy of the subsequent hydrothermal stage is discussed under the next main heading. All three stages, early magmatic (including crystallization of olivine, pyroxene, amphibole, some mica and plagioclase feldspar), deuteric (including a complex assemblage of potash feldspars and quartz) and hydrothermal (including feldspar destruction to sericite) comprise an orderly succession of events in the complete crystallization of the magma. No breaks are recognized between the stages, rather, one merges into another.

Fracturing during late crystallization tends to promote the localization of some of the minerals in the different stages. In the deuteric stage, aplite and pegmatite may be localized into dikes, and in the hydrothermal stage, the fractures localize the sites of deposition for the common hydrothermal minerals. In this section, the deuteric phenomena are discussed without reference to fracturing, insofar as possible. In so doing, the deuteric-hydrothermal boundary is crossed and such hydrothermal phenomena which are noted at a few localities without structural control, are discussed. In the succeeding section on the hydrothermal stage, the discussion starts with the late deuteric stage where deuteric minerals are localized or controlled by fracturing. The dominant control in the hydrothermal crystallization stage is, of course, related to fractures.



Definition. There is probably no great difference of opinion among petrologists about the general position of the deuteritic stage in the crystallization of an igneous magma. Gillson (1939) and Colony (1923) extended Sederholm's original definition to provide a very usable and generally agreed upon crystallization stage. (Shand 1944) However, the specific limits of the deuteritic stage are not well defined. There appears to be, from the literature, a particularly fuzzy line between the deuteritic and hydrothermal stages. The petrologists seem to have been content to carry the crystallization of the magma through the normal magmatic and into the deuteritic stage. The economic geologists have started with "hydrothermal solutions" recognizing some link with pegmatites and possibly aplites as "pre-hydrothermal" but, with the exception of Ross (1935), rarely distinguished the continuity which may be evident in favorable localities. The Boulder batholith appears to exhibit a complete succession of events which grade from one stage into another from early magmatic crystallization through the deuteritic into the hydrothermal stage.

As originally defined, the term was proposed by Sederholm (1926) to include late primary albitization and myrmekitization of "solutions, gasses, and liquids which belong to the magma itself." (Gillson 1929, p. 101.)

"I think it would be advisable to discriminate between metasomatic changes which belong to a later period of metamorphism, i.e., are secondary in the strictest sense of the word, and those that have taken place in direct continuation of the consolidation of the magma of the rock itself. I propose to call the latter deuteritic, as distinct from secondary changes." (Sederholm 1926, pp. 141-142.)

Gillson strongly objected to the use of hydrothermal or pneumatolitic to describe these late magmatic features (Gillson 1929, p. 101) Colony proposed to extend the use of the term "deuteritic" to cover all magmatic end-stage emanation phenomena which frequently cause large-scale changes and very profound effects, especially in the way of mineralization. (Colony



1923, pp. 170-171) It was in the sense of profound endomorphic effects that Robertson (1954) applied the term to end stage feldspar metasomatism leading to differentiation of a gabbro body in the Mt. Sheridan diabase, Wichita Mountains, Oklahoma. Gillson summarized the use of the term deuteritic as follows:

"I describe as deuteritic effects the formation of titanite, allanite, and epidote, and the replacement of quartz and plagioclase by microcline and the chloritization of the biotite in the granodiorite of the Pend Orielle district of Idaho (Gillson 1927) and published sketches showing deuteritic magnetite, titanite, and allanite from a number of rocks . . . In a later paper (1937b) I described albitization and other effects in the granite at Conway, New Hampshire, as deuteritic. . . I thus conclude that the term deuteritic is established as covering those metasomatic changes in igneous rocks caused by the reaction between the minerals already formed and the emanations, gaseous or liquid, percolating through the solid, or almost solid rock, and which have been given off from the same magma from which the rock itself crystallized." (Gillson 1929, pp. 101-102.)

As used by Sederholm, Gillson and Colony, deuteritic is used specifically for end-stage feldspar and quartz-feldspar intergrowth crystallization.

As used in this investigation, deuteritic is broadened somewhat to include the entire potash feldspar crystallization stage. The groundmass crystals of potash feldspar and quartz in quartz monzonite porphyry are clearly late magmatic and are not normally considered to be "deuteritic." The crystallization of granophyric intergrowths in the groundmass is clearly deuteritic, as already defined. The early crystalloblastic growths of potash feldspars in the groundmass and porphyroblastic growths are later than "late magmatic" groundmass crystallization, and as such, involve the recrystallization and reorganization of the groundmass at the deuteritic stage. As there does not appear to be any clear distinction



between "late magmatic" potash feldspar in the groundmass, and "deuteric" granophyre, all of the potash feldspar crystallization is included in the deuteric stage.

The lower, or cooler, limit of the deuteric stage, that is the deuteric-hydrothermal boundary, is defined by the change from stable potash feldspar assemblage to feldspar destruction by sericitization and argillization. (Klepper and Robertson 1956.) This mineralogical boundary is an easily recognized basis for the distinction of the hydrothermal stage which is in no way defined by fracturing or by "hydrothermal solutions" of unknown origin.

Deuteric Stage Minerals and Textures. In quartz monzonite, the porphyroblastic growths of potash feldspar, the replacement of plagioclase by anorthoclase and orthoclase, the formation of reorganization perthites, and the granophyric textures are all deuteric stage minerals. These minerals and textures have already been described in detail.

The alteration of plagioclase to essentially anhydrous minerals, including clinozoisite, albite-oligoclase, calcite and flecks of sericite, are characteristic of the deuteric stage.

The alteration of the mafic minerals to hydrous minerals appears to start in the magmatic stage as shown by the alteration of pyroxenes to secondary hornblende and the crystallization of biotite. The formation of chlorite commonly takes place in the deuteric or even hydrothermal stage.

Origin of Aplite-Alaskite. In the Boulder batholith, the groundmass of much of the quartz monzonite has the composition of aplite. The crystallization of aplitic groundmass takes place after about 60 to

70 per cent of the essential minerals of the rock are crystallized. The late residual solutions are not only rich in potash feldspar and quartz, but also contain water, boron, sulfur and probably other hyperfusibles. These hyperfusibles probably play a significant role in the recrystallization of potash feldspar, in the groundmass, to porphyroblasts where there is no avenue for escape. Where fractures occur before the groundmass has completely crystallized, the late solutions are driven out by the volatiles into the joints and fractures to form intrusive aplite bodies. Thus there are two ways that late residual solutions may produce aplites and alaskites.

1. The late solutions may segregate, or actually migrate and replace earlier minerals to form bodies of aplite and alaskite. If there is a general release of pressure near the roof of the batholith, a streaming of the late solutions toward the roof would provide the materials for the wholesale replacement of early-formed minerals to alaskite or aplite or both. Likewise the volcanic rocks which form the roof of the batholith may be altered to aplite composition.
2. Fractures which form before the groundmass become crystallized may provide avenues for collection of and migration of the aplitic residues and form dikes of aplites or alaskite. The volatiles may escape, but some may be trapped and contribute to the formation of small pegmatitic patches found in many aplites.

The wide variety of potash feldspars and the many relics of plagioclase in potash feldspars in the rocks at the border of the



Ontario body of alaskite are thought to be due to incomplete replacement of the original plagioclase of the quartz monzonite.

Similarly complex mineral assemblages have been noted in alaskites by other investigators. Spurr recognized quartz, microcline, orthoclase, anorthoclase, albite and oligoclase-albite (Spurr 1906, p. 103). He states: "Two distinct periods of generations of crystals are always represented." (Spurr 1906, p. 105). Plagioclase is the principal mineral in the first stage, and microcline, anorthoclase, orthoclase, and quartz in the second. Some tourmaline is present as well. The alaskite grades from a biotite granite through alaskite, to quartz-rich vein-like bodies. He illustrates his conception of the genesis of quartz veins in alaskite in a plate which is reproduced in Figure 70. These mineralogical and structural relations to quartz veins are distinctly similar to the general sequence of observations in the Elliston district, and strongly suggest a deuteric metasomatic history, although Spurr considered the alaskites magmatic intrusions (Spurr 1900).

Loughlin (1912, p. 124) points out the "secondary character of the microcline twinning structure is clearly shown by gradation into orthoclase. No good opportunity was offered for measuring the extinction on 010 sections, and its sodic character can only be inferred by the small amount of albite present in comparison to the  $\text{Na}_2\text{O}$  in the chemical analysis." (3.56 per cent  $\text{Na}_2\text{O}$  5.33 per cent  $\text{K}_2\text{O}$ .) He further states, "The albite is closely associated with microcline."

Tyrrell (1916, p. 438) describes the orthoclase in an alaskite in southern Georgia, Falkland Islands in these words: "The thin section shows that the rock consists of a granular mixture of quartz



THIN SECTION SHOWING GENESIS OF MAGMATIC QUARTZ VEIN IN  
ALASKITE.

Figure 70. Quartz segregation veinlet in alaskite. (Spurr 1900)



and orthoclase. Some of the latter shows a mottling between crossed nicols, due to an admixture of the albite molecule, (like anorthoclase? personal note) but most of the orthoclase is perfectly uniform and may be regarded as comparatively pure. Occasionally there is a granophyric intergrowth between orthoclase and quartz." Some oligoclase and a few flakes of "decomposed biotite" are found in the rock.

Washington (1898, p. 790) states that the alkali feldspars of the Rockport "granite" (a ribeckite alaskite) "are varied and deserve more extended study. . . They are chiefly orthoclase and albite, which are almost always intergrown so as to form highly typical microperthites and cryptoperthites. In general they are closely similar to those described by Brögger (1890) and Ussing (1898) from Norway and Greenland syenitic rocks, and their figures would answer for forms seen here. Cryptoperthitic development is relatively scarce, the majority being perthitic on a fairly large scale. Microcline (possibly anorthoclase? personal note) also is not rare, but less abundant than the preceding. This and the orthoclase are apparently rich in soda. (my underlining) The feldspars are apt to be rather cloudy and dusty with decomposition products, and they often show a brecciated border and undulatory extinction. . . a very little albite-oligoclase is seen in a few sections, (relics?) showing well-developed twinning lamellae."

It appears abundantly clear that the feldspar composition of alaskites is not so simple as that found in many normal granites. The varieties and variations from the "norm", attributed to the high soda content by several investigators, (Spurr 1906, Tyrrell 1916, Washington 1898) are clearly brought out in the alaskites described above.

The complex mineral assemblage noted by other investigators in other bodies of alaskite, and the mineralogical identities of several bodies of described alaskite, cause me to wonder if alaskites may generally be formed by metasomatism of early-formed magmatic minerals.

The moderately large amount of sericite in some alaskite, and in the aplitized volcanics in the roof as well, suggests that the alaskite formation persisted late into the deuteritic stages and crystallization extended into the early hydrothermal stage at which time the feldspars were, in part, altered to sericite.

Further petrographic work on alaskites, coupled with chemical analyses, not only of the rocks but of the constituent minerals, is needed to finalize the deuteritic metasomatic origin of the alaskites in the Elliston district.

#### Hydrothermal Stage.

"It is common practice to say of any ore deposit, not distinctly sedimentary, that it has come from below, and to rest content with this statement, which, even if not susceptible of direct proof, has the merit that in one sense it cannot be disproved." (Emmons 1885, p. 572). This quotation from S. F. Emmons, which appeared over 70 years ago, is distinctly apropos today. The assignment of the vein-forming materials to unknown sources well below the depth of our observation does little, in my opinion, to effect a solution of the problem of the derivation of the ore minerals in veins in igneous rocks, particularly where there is some direct evidence of local derivation for some of the vein minerals as in the case in the Boulder batholith. These findings



are discussed in cooperation with Klepper and Pinkney. (Robertson, Klepper and Pinkney 1956).

In the preceding sections dealing with magmatic and deuteritic phenomena, evidence is presented for continuous crystallization for some minerals from the magmatic stage into the deuteritic stage. Several of the minerals which begin their crystallization in the deuteritic stage, and in some cases even before, appear to carry through into the early hydrothermal stage. If, however, as soon as a vein structure is recognized we are obligated to rotate our thinking  $90^{\circ}$ , or vertically, we might overlook or ignore relations which may have a direct bearing on the source and accumulation of some of the vein minerals.

In the crystallization history of the magma, the sequence of mineral formation events from the clearly magmatic stage, such as pyroxenes and plagioclase, are succeeded by the crystallization of potash feldspars and quartz in the late magmatic or deuteritic stage. Although very minor in amount, tourmaline and pyrite crystallizes at an even later stage, commonly accompanied by moderately strong alteration of plagioclase to sericite, and biotite to chlorite, immediately adjacent to pyrite grains in quartz monzonite. These are the same minerals which are found in the veins which are clearly of hydrothermal origin and which, in the veins, have the same alteration products adjacent to the minerals of the vein.

It is the overlap of magmatic-deuteritic-hydrothermal crystallization in the rocks, without relation to structural features, with the deuteritic-hydrothermal stage mineral formation in vein-like structures and veins, which give the most convincing evidence of a virtually uninterrupted succession of events from the first crystallization of the magma to

the quartz-tourmaline-pyrite vein structures, and presumably to the base metals found in the vein structures.

Definition. As herein defined, the hydrothermal stage is marked by the destruction of feldspar to hydrous minerals, principally sericite and clay minerals, as distinct from the feldspar-stable, or deuteric stage, in which feldspar alteration, where present, is to anhydrous minerals such as epidote, albite and potash feldspars. The hydrothermal stage is clearly marked by the crystallization of sulfide minerals and a variety of non-metallic gangue minerals.

Structures in pre-hydrothermal and hydrothermal stage. In the magmatic stage, there are no obvious localized structures which control crystallization, although in a large sense, the finer-grained quartz monzonites are distinctly related to large structural features such as the margins of the batholith. In the deuteric stage, the aplite-alaskite in the Boulder batholith is not obviously controlled by large or small structure, however a few alaskite bodies are distinctly related to the roof of the batholith in the Elliston district. Many small aplite and alaskite bodies and their related pegmatites are controlled by fractures which localized them into vein-like bodies or dikes, commonly a few inches in width, such as the small dike shown in Figure 63.

The hydrothermal stage is normally intimately related to vein formation, although some hydrothermal stage crystallization takes place without obvious structural controls, such as cases of strong sericitization related to isolated segregations of sulfide minerals in quartz monzonite.



Aplite bodies, which contain pegmatitic segregations, are relatively common. In many cases, aplite with internal patches of pegmatite are localized into vein-like structures. In these, the tourmaline is younger than quartz, and pyrite younger than tourmaline, although they distinctly overlap in the crystallization of all three minerals. Quartz-tourmaline intergrowths and tourmaline pyrite intergrowths are distinctive features.

Segregations of feldspar-tourmaline + pyrite, without significant amounts of quartz, are, in some instances, clearly post aplite in age. These minerals occur in segregated pods or clusters in quartz monzonite, rarely in bodies as much as five inches in diameter. See Figure 71. Other segregations of similar mineralogy are elongated in habit and clearly controlled by early, very inconspicuous, fractures. In some instances, a number of pods have been observed to form along a zone which undoubtedly was a minor fracture zone. In these, tourmaline and pyrite are concentrated in the core, surrounded by a leucocratic zone or selvage of potash feldspar and potash-bearing feldspar, including anorthoclase. (See Figure 72.) In this zone, the plagioclase crystals which were present, and which are clearly indicated in Figure 72, have been replaced to anorthoclase (white) and other potash feldspars, the mafic minerals have been eliminated with only patches of white mica to indicate their former presence. At the boundary of the leucocratic zone with quartz monzonite, biotite is strongly altered to chlorite. Granophyric replacements of plagioclase are commonly distinctive at this interface.



Figure 71. Hand specimen showing tourmaline-pyrite segregation in quartz monzonite. X2 natural size.



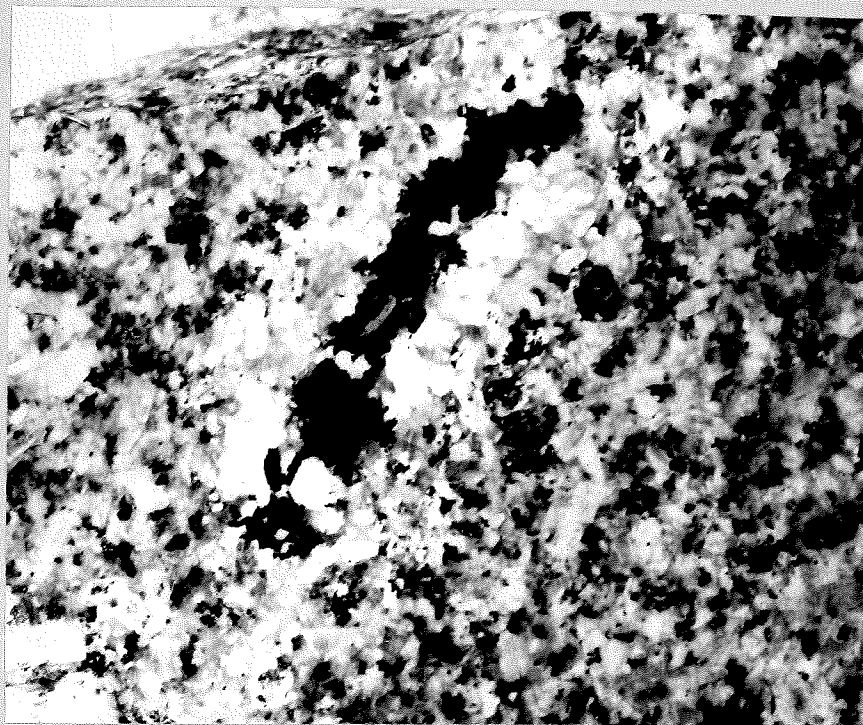


Figure 72. Hand specimen (sawed surface) showing tourmaline-pyrite segregation pod with leucocratic potash feldspar-rich selvage zone. White crystals are anorthoclase after plagioclase. X  $1\frac{1}{2}$  natural size.

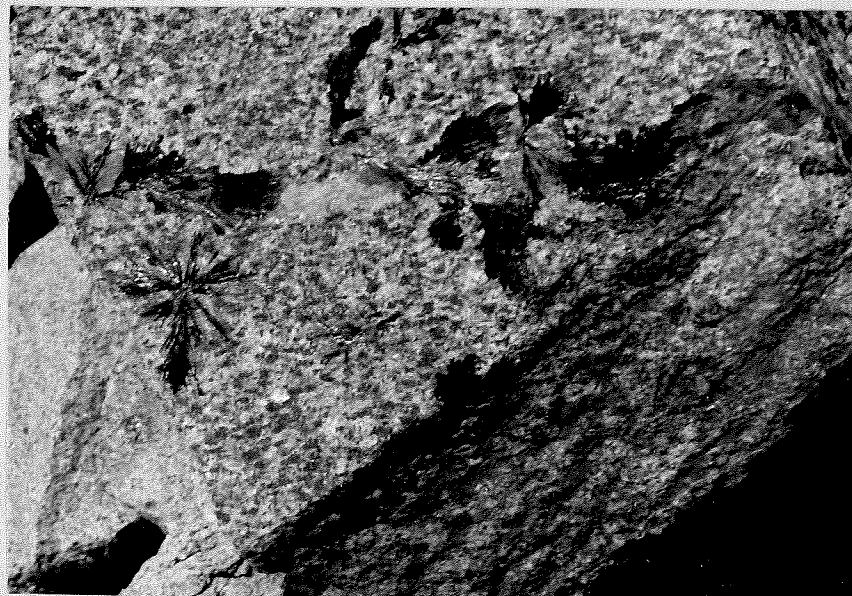


Figure 73. Tourmaline-pyrite "sunbursts" on joint surface. Note concentration of tourmaline along trace of small aplite dikelet. Sample from dump of Gray Eagle mine, Basin district.



Along some joints, tourmaline-pyrite "sunbursts" occur in discontinuous patches along the joints as shown in Figure 73. Adjacent to these joints there is a thin leucocratic selvage which is identical to that found around the segregations. As the iron, boron, and pyrite in the segregations is clearly locally derived, so the minerals in the joint walls may have derived the boron, sulfur, and iron from the walls. In fact, where a joint has cut across an early aplite, in Figure 73, tourmaline is seen to be somewhat more abundant, concentrated along the trace of the aplite with the joint, however there is little indication that the aplite provided more than a minor amount of boron to the joint plane for the formation of tourmaline. The various types of segregations and joint coatings are shown in Figure 74 as leading to vein formation.

The selvage zone along the joints which exhibit the tourmaline-pyrite sunbursts is not uniformly thick, although present where no tourmaline or pyrite occurs on the joint. Where tourmaline and pyrite are found, the selvage zone appears to be, at least in some instances, somewhat wider. The "sunbursts" represent joint controlled segregations, whereas the pod-like segregations formed along a much less conspicuous structure. The migration of materials from the quartz monzonite to the joint appears to be exactly the same as in the segregations.

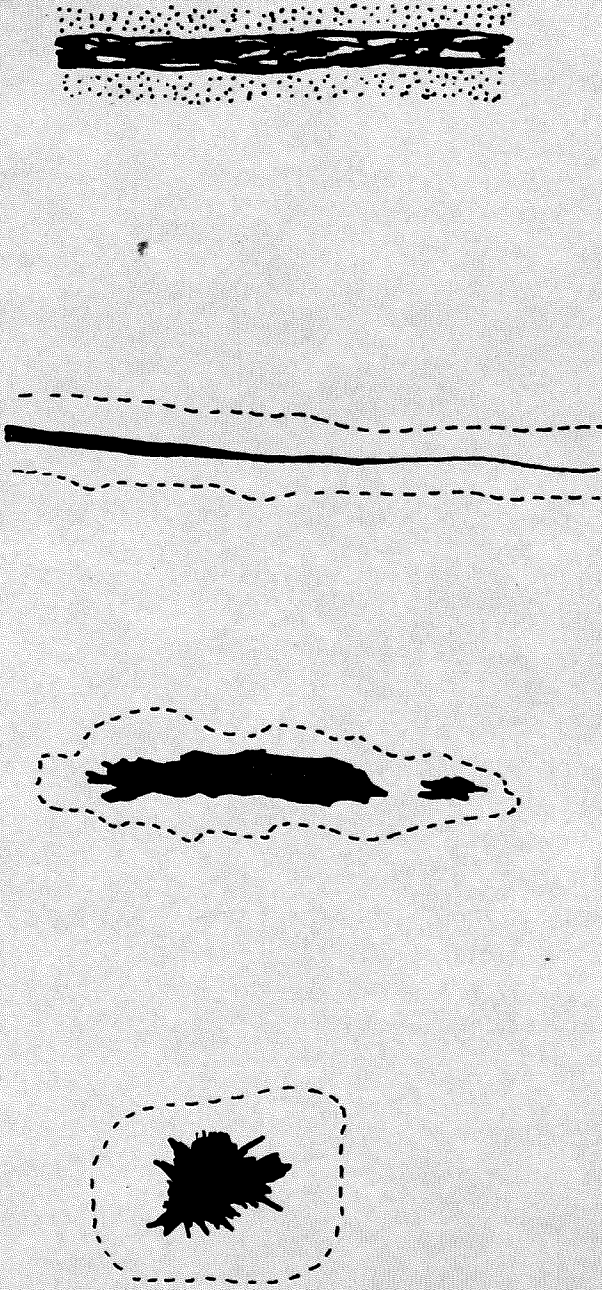
Similar exudations of iron from the walls of joints has been shown by Mackin (1947) at Three Peaks laccolith in the Iron Springs district in Utah. Magnetite, apatite and rarely a pyroxene, were derived from selvage zones of primary distinction joints.



# DEUTERIC HYDROTHERMAL

TOURMALINE ± PYRITE

+ QUARTZ



LEUCOCRATIC POTASH RICH  
FELDSPAR SELVAGE

LEUCOCRATIC  
SERICITE ALTERATION

SEGREGATIONS

JOINT COATING

VEIN

# QUARTZ MONZONITE

Figure 74



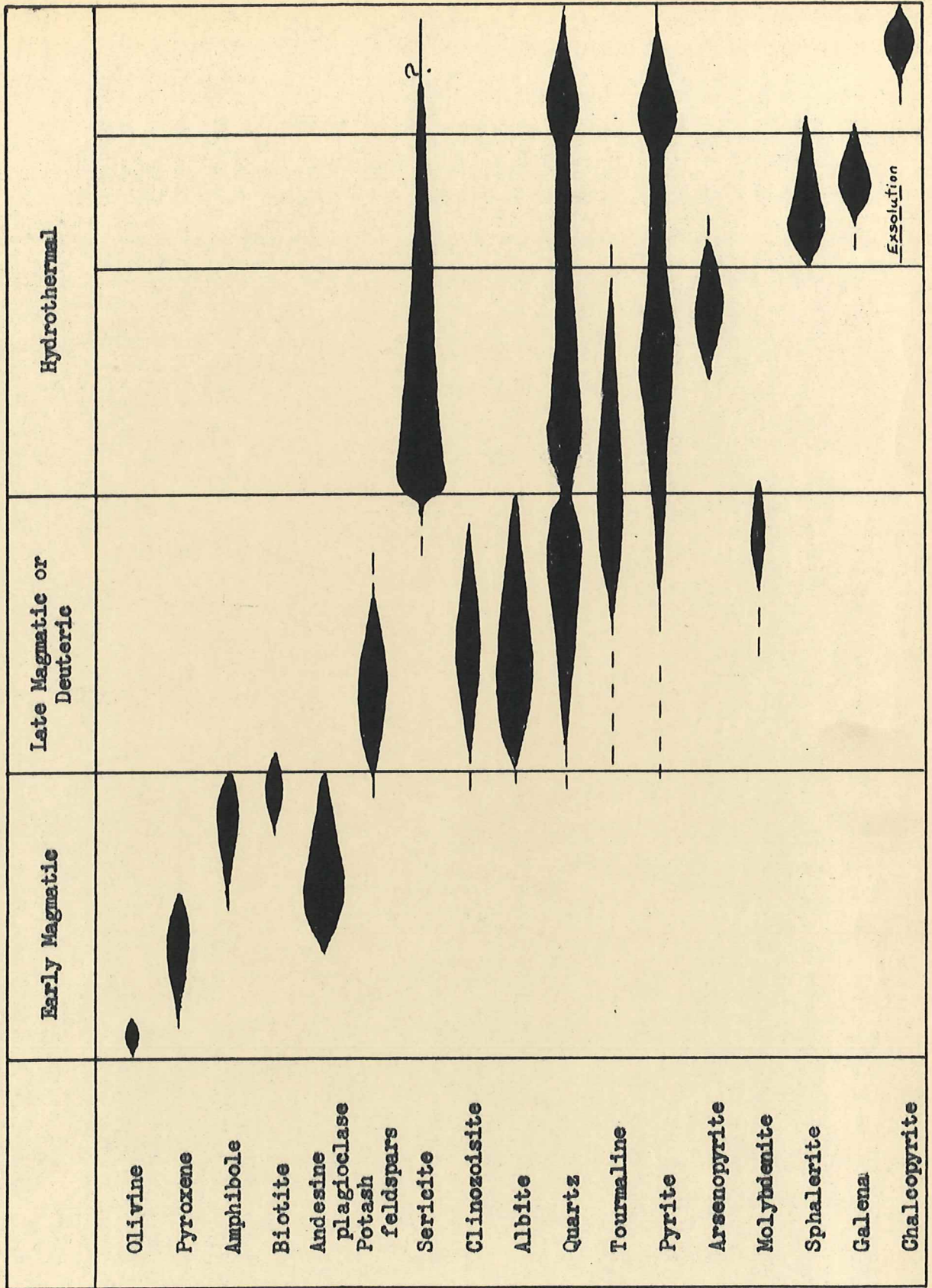
The selvage zones of the tourmaline-pyrite segregations are composed dominantly of potash feldspar and potash-bearing feldspars which have replaced the early-formed plagioclase which were formerly present. In the segregated pods, this represents a clear migration of potash toward the segregation. In the joint coatings, which actually involve some replacement of the walls by tourmaline and pyrite, the same selvage is noted. In the case of the selvage zone for both occurrences, pods and joints, some fleck-sericite alteration of plagioclase is noted at the boundary between the selvage zone and the normal quartz monzonite.

It appears to be but a short step from the tourmaline-pyrite joint-facing, without quartz, to thin tourmaline-pyrite veinlets, with quartz. Instead of a leucocratic potash feldspar-rich selvage, there is a leucocratic sericite-quartz selvage zone adjacent to the vein representing an unstable feldspar environment, possibly only a few degrees cooler than the stable feldspar stage. More complex vein structures, without base metals, are characterized by strong wall-rock sericitization adjacent to the veins. If, in the earliest vein formation and vein-like stages, significant amounts of iron, boron, and sulfur were clearly derived from the walls, then the more complex vein structures may likewise derive the same materials from the walls.

Crystallization Histories of Minerals. A review of the crystallization histories of the minerals found in vein-like structures and in the hydrothermal veins is given to indicate the paragenesis of mineral formation. Figure 75 is a paragenetic diagram showing the time and relative intensities of formation of the minerals. The periods of maxima indicate distinct "surges" of the minerals, but are not quantitative, even for



Figure 75. Paragenesis of minerals in crystallization of Quartz Monzonite



Exsolution

the same mineral. For example, quartz in the magmatic and deuteric stage is obviously more abundant than in the vein formation stages, yet the nodes on the graph might be interpreted otherwise without this explanation. Chalcedonic quartz, for example, is piddling in amount, compared to the earlier vein quartz. The nodes do, however, indicate the relative position of the more abundant periods of crystallization.

(Potash feldspar) In the deuteric stage, potash feldspar crystallized along with quartz. The earliest crystallization of potash feldspar in the groundmass of some of the quartz monzonites, and even the early porphyroblastic growths, is a distinct extension of the deuteric stage and would be regarded by many as a late magmatic stage, which it actually is. If it were possible to distinguish magmatic potash feldspar from deuteric potash feldspar by some convenient mineralogical or textural feature, such a distinction would be made.

The porphyroblastic growths of the potash feldspars has been described at some length in a preceding section. Potash feldspar occurs in aplite and pegmatite, the latter commonly displaying a distinct feldspar-rich zone with a series of quartz crystals in the core of the pegmatitic pod. It is abundant in alaskites, probably in part occupying a wide variety of mineralogical habits in the anorthoclase to normal orthoclase series.

In feldspar-tourmaline  $\pm$  pyrite segregations, potash feldspar forms a leucocratic selvage between quartz monzonite and tourmaline. The potash feldspars are in large part anorthoclase which replaces early plagioclase and is in turn partly replaced by orthoclase, and thus constitutes a concentration of potash in the selvage. At the contact of the leucocratic border with quartz monzonite, the "front" advances partly by the recrystallization of feldspar to granophyric textures, both micrographic and myrmekitic.



Similar concentrations of potash feldspars are found along some vein-like structures. From mineralogical evidence alone, there appears to be a clear migration of potash towards segregations and towards vein-like bodies.

(Sericitic) Sericite is seen to be present in flecks in plagioclase in the deuteritic stage, however the abundant sericite found in decomposed feldspar marks the hydrothermal stage with or without structural control. The wall rock alteration to sericite adjacent to veins of quartz  $\pm$  pyrite  $\pm$  tourmaline is characteristic and universal and found next to all veins except possibly the very low temperature chalcedonic veins. In the alteration of feldspar to sericite, some silica is liberated which may be utilized in quartz vein growth. Although there are no chemical data available from wall rocks to indicate the chemical transfer in the walls of veins, the mineralogy indicates a migration of potash in much the same manner suggested by Sales and Meyer (1947) in the Butte district.

(Quartz) Quartz crystallizes principally in the deuteritic stage, the early part may be considered by many to be late magmatic. Quartz is ubiquitous in the groundmass of quartz monzonites, in aplites and alaskites. It is commonly late, although potash feldspar porphyroblastic growths clearly replace some quartz, and some quartz possibly liberated in porphyroblastic growth, is younger than potash feldspar porphyroblasts. In pegmatitic segregations, quartz is clearly younger than potash feldspar, although in the early quartz crystallization stage, there is a very obvious overlap, and intimate graphic intergrowths are common. In many of the feldspar-tourmaline segregations, quartz is virtually absent or present in very minor amounts. For this reason, the quartz node at the end of the deuteritic stage, shown in Figure 75, indicates a tapering off of quartz crystallization at



the close of the deuteric stage.

Hydrothermal quartz is commonly indistinguishable from deuteric quartz except in veins. Some pegmatites have low temperature quartz crystals, and some comb-structured quartz is present in the younger hydrothermal vein stages. It is important to note that with the appearance of quartz in vein structures with pyrite and tourmaline, sericite is always found in the wall rocks and normally is relatively large in amounts with fine-grained quartz and pyrite. In the alteration of feldspar to sericite, silica is liberated. It appears clear in some instances that minute quartz veinlets could hardly have gathered silica from more than a few millimeters to a few centimeters away from the veinlet, and in some cases it appears clear that quartz from the wall rocks contributes significant amounts of silica for the construction of quartz veins. Some of the silica liberated in the feldspar alteration may migrate into the vein and through the vein to some site favorable for deposition.

Still younger chalcedonic quartz is found in many veins in the Boulder batholith, particularly in the Boulder-Basin-Wickes districts, but only a very small amount is present in the Elliston district.

(Molybdenite) Molybdenite is found in minor amounts as joint coatings in quartz monzonite at a locality on Clancy Creek, northwest of Clancy, and at 5 other localities in the batholith, none of which is in the Elliston district. (Pinkney, personal communication)

Molybdenite is a minor mineral encountered in some aplite and pegmatite bodies, notably east of Butte just north of Pipestone Pass, and at another locality towards the Highland Mountains on Roosevelt Drive. In the late deuteric stage, minor amounts of molybdenite are found in feldspar-tourmaline-



pyrite segregations and joint coatings at the Gray Eagle mine in the Basin mining district.

In the eastern part of the batholith, molybdenite is found with quartz and pyrite in moderately strongly sericitized rock, suggesting that molybdenite crosses the deuteritic-hydrothermal crystallization boundary.

(Klepper, personal communication) Molybdenite occurs in strongly sericitized feldspar in quartz porphyry at Butte, yet is distinctly younger than the copper mineralization. (Sales and Meyer 1956)

(Tourmaline) Tourmaline is found as an accessory mineral in some quartz monzonite. Klepper considers that accessory tourmaline is somewhat more calcic than tourmaline in segregations and veins. (Personal communication) Accessory tourmaline has only been recognized in the vicinity of the Julia mine in the Elliston district. Tourmaline is locally abundant, although late forming, in aplite, alaskite and pegmatite in assemblage with stable feldspars in the deuteritic stage. In pegmatites, tourmaline is clearly younger than most of the quartz, although quartz and tourmaline intergrow at the base of some tourmaline clusters. In aplites and alaskite, tourmaline occurs in small clusters in the rock and in vein-like segregations. The tendency for tourmaline to accumulate in quartz-rich segregations and vein-like bodies in alaskite has been described by Murray-Hughes (1929, p. 109) In the feldspar-pyrite-tourmaline segregations, in which magmatic quartz is very minor in amount, tourmaline is the youngest mineral, replacing feldspar and quartz, where the latter is present. In post-aplite joints, tourmaline and tourmaline-pyrite "sunbursts" occur in the joints. At one locality, a boulder on the dump of the Gray Eagle mine shows a joint surface on which a number of tourmaline-pyrite "sunbursts" are present



as shown in Figure 73. The joint clearly cuts across a small aplite diklet, about 1 in wide. Along the diklet in the joint plane, tourmaline is conspicuously more abundant. The suggestion is that the aplite was still somewhat permeable and that boron perconated through the aplite and crystallized in the joint plane in the presence of the iron available, as evidenced by the pyrite which is in the joint but not noticeably concentrated with respect to the aplite dikelt. In the Rimini district, Knopf (1913, p. 47) describes massive tourmaline "ledges" without much quartz found adjacent to quartz veins.

In hydrothermal quartz-tourmaline veins, tourmaline is clearly intergrown with vein quartz. In some veins, the intimate quartz-tourmaline intergrowths are so fine grained that the vein matter resembles anthracite coal. In others, tourmaline occurs in dense mats of needles throughout the quartz. Most of the large tourmaline-quartz veins contain pyrite.

It is uncertain if tourmaline crystallization continued into the lead and zinc mineralization stage or not. At a few localities, it is suggested that some tourmaline is present in the veins in intimate association with gold and silver, and possibly in association with zinc and lead. In general, lead and zinc orebodies are clearly younger than tourmaline-quartz.

(Pyrite) Pyrite crystallizes under a wide variety of conditions and over a long period of time. It is found with mafic mineral clusters in the igneous rocks. In such cases, sericite flecks are found in plagioclase and biotite is altered to chlorite immediately next to pyrite grains. The pyrite may be "primary" or sulfidized magnetite formed at the deuteritic stage.

Pyrite is found in many aplites, and in some pegmatites. In both it is clearly a late mineral, however, indigenous to the rock in which it is



found. In quartz-porphyry dikes in the Clancy district, marcasite and sphalerite, galena and chalcopyrite, are primary constituents of the rock. (Klepper, personal communication) In tourmaline-pyrite segregations and vein-like bodies, pyrite is intergrown with and partly replaces tourmaline. In many of the segregations, the pyrite is clearly derived from the closely adjacent rock in which it is found. In the tourmaline-pyrite joint coatings, the pyrite is in association with stable feldspars, however a flecking of sericite is commonly noted in plagioclase closely associated with the pyrite.

Quartz-tourmaline-pyrite veins and quartz-pyrite barren veins, are often large, early veins which occur in a nearly east-west trend so common throughout the batholith. Pyrite is also found in the zinc and lead mineralization stage, although it is commonly not abundant at this stage. It forms very abundantly in association with chalcopyrite at a still younger hydrothermal mineralization stage (Pinkney, personal communication). In all of the hydrothermal veins which contain pyrite, the wall rocks are altered to sericite next to the vein. In many veins where the sericite zone is wide, quartz and pyrite are disseminated throughout the wall rocks.

(Sphalerite-Galena) Sphalerite and galena commonly occur together, although in some instances, individual small ore shoots are dominantly galena or sphalerite. At the Clark breccia zone, sphalerite is present without much galena, and at the Hopkins mine, some ore shoots were massive galena or galena and pyrite without appreciable amounts of sphalerite. The base metal minerals occur in veins closely associated to or within early main-stage quartz veins. In all cases, the walls of galena-sphalerite veins are altered, however in many cases, it cannot be demonstrated that the alteration is related specifically to the base metal introduction. In all cases, the



base metal-bearing veins are small, and ore shoots are localized within, or adjacent to the early strong veins which have intense wall rock alteration throughout the course of the vein.

(Copper Minerals) Chalcopyrite is the only copper mineral of any importance, and it is minor in amount throughout the northern part of the batholith, and especially in the Elliston district. It is commonly found as exsolution blebs in sphalerite, and in many cases, it is the only copper mineral. In some of the lead and zinc deposits, and especially in the more zincy deposits, some late chalcopyrite is found. In some veins, quartz-pyrite-chalcopyrite comprises the mineral assemblage. (Pinkney, personal communication) These veins are considered to be distinctly younger than the lead-zinc veins. The age relations are reported to be clearly shown at the Crystal mine in the Basin district. In the Elliston district, the Copper King-Copper Queen vein is a large quartz-pyrite vein with some copper mineralization and some gold. Its age relations to the other base metal structures is unknown. The vein has the same region trend, about N. 70° E. that the other main veins have, and it is spaced inbetween other large veins which contain lead and zinc ore shoots.

(Carbonate Minerals) Late deuteric alteration of plagioclase produces some carbonate mineral. Small carbonate stringers are occasionally found. The earliest of the carbonate veinlets, possibly late deuteric, are composed of calcite. Ankeritic carbonate veinlets, which cut lead-zinc veins, are found in the northern part of the batholith, but are rare in the Elliston district.

Vein Structures. Fractures are able to form in nearly crystallized or completely crystallized quartz monzonite and provide channels for the



solutions which form replacement lodes and fissure fillings. Aplitic stringers apparently form in advance of complete crystallization of magma. Fractures which apparently were in part open, and into which aplitic solutions moved with facility, are shown to be sharp vein-like or dilation diklets of aplite as shown in Figure 63.

Replacement aplitic and pegmatitic bodies commonly have a tabular shape and appear to be controlled by incipient fractures. Likewise, replacement pods of feldspar-tourmaline-pyrite are similarly controlled by very obscure fractures, whereas others are clearly formed along joints, some of which cut early aplite diklets, as shown in Figure 73.

From another line of approach, small segregations of late quartz are commonly found in quartz monzonite and in alaskite. One crystal-wide strings of quartz grains occur in the rocks, as shown in Figure 76. Quartz veinlets with tourmaline, occur in segregation-like bands in alaskite as has already been described. In quartz monzonite, carbonate and quartz stringers, with or without pyrite may form. Such a stringer is shown in Figure 77. It is a small veinlet which contains a carbonate mineral and some quartz and which cuts across a plagioclase phenocryst. There is a marked zone of sericite adjacent to the veinlet but which does not persist more than about 1 mm from the veinlet. In the plagioclase crystal, the veinlet is occupied by a light greenish, pleochroic mineral which has bluish-gray interference colors. It is probably a chlorite.

All of the veins or vein-like structures which occur at the deuteric stage are, for the most part, rather indistinct and not through-going structures except for some aplites. (Through-going aplite dikes may have formed at a relatively late stage crystallization materials from some depth so that the age of the fracture, with respect to the rock in which



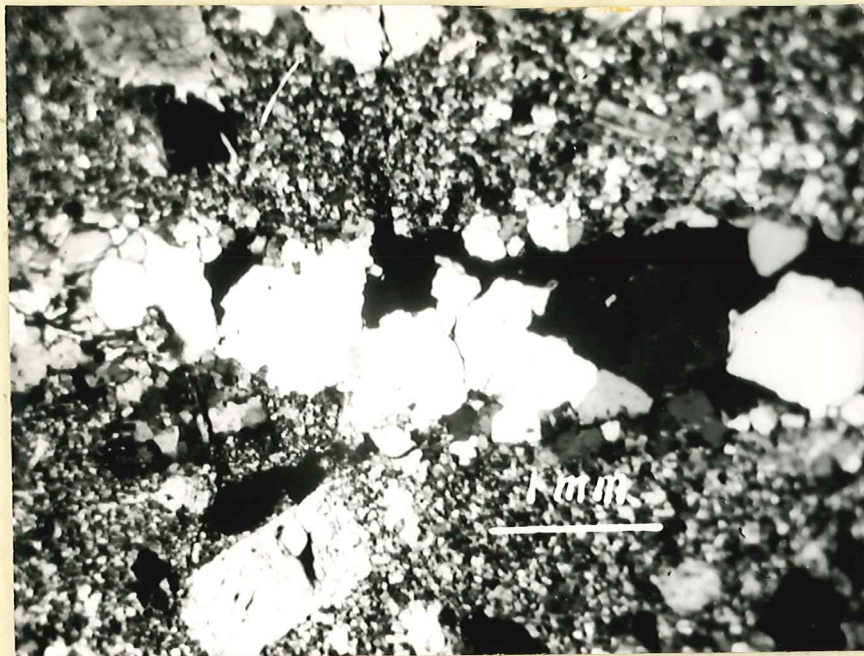


Figure 76. Photomicrograph of quartz segregation stringer in quartz monzonite. (Scale should be 0.5 mm) Crossed nicols.



Figure 77. Photomicrograph of quartz and carbonate veinlet which cuts plagioclase phenocryst. Note sericite alteration on both sides of veinlet. Veinlet occupied by chlorite in plagioclase phenocryst. (Scale should be 0.5 mm.) Crossed nicols.



the dike is found, may be quite late).

The structures in the distinctly hydrothermal stage have well defined walls which are clearly controlled by fractures. Some large fracture zones are several miles in length, however there is little evidence that more than minute openings were ever present at any one moment of time. The majority of the early veins formed by replacement, rather than by fissure filling. Some open space crystallization in brecciated veins and comb structures along small open fissures are observed, locally. Each of the larger veins which have been examined are complex lodes in which fracturing and refracturing occurred at many times along a vein zone. (See "Veins in Quartz Monzonite" under Mineral Deposits" for details). Many of the so-called "shear zones" found in the batholith are really complex joint or fracture zones in which very small movements have produced openings time and again forming complex vein structures. Some of the large fracture zones, such as the Comet-Gray Eagle structure, are as much as 100 ft wide and possibly more than 6 mi long. In these large veins, minerals have been deposited at a number of different times at different localities along the vein zone.

The earliest hydrothermal veins are simple quartz-pyrite strands, with or without tourmaline. Individual stringers are actually rare. Normally they comprise complex veins with strongly sericitized wall rocks which are impregnated with small pyrite crystals. The early quartz-pyrite-tourmaline veins are mostly barren of precious or base metal mineralization.

The lead-zinc deposits are confined to distinctly younger fractures which are distinctly related to the main vein zone but which may occur: (1) within the early vein, (2) in fissure veins more or less parallel to and along one wall or the other of the main vein, (3) in fissure veins



partly in the wall rocks, (4) in enechelon gash structures within the main vein zone, or (5) in brecciated zones within the main vein. At no locality in the northern part of the batholith has a base metal ore shoot been followed continuously in depth to as much as 1,000 ft, nor do the base metal ore shoots have considerable lateral extent along the veins. Billingsley and Grimes (1918) laid particular stress on the relationship of base metal mineralization to a zone at and near the top of the batholith which, since the day of their observations, has not been demonstrated to be incorrect. In the Elliston district, very few of the veins have been explored to any appreciable depths. At the Ontario mine, which was one of the important producers, gold, lead and zinc mineralization persisted to about 300 ft, possibly deeper, but at the bottom of the shaft at the end of the upper adit level the ores had apparently given out. The outcrop of the vein is in alaskite, possibly less than 100 ft from the roof of the batholith. At the Lilly mine, there are no base metals at the 200 ft level although the vein at that level is a wide. The outcrop of the vein may be several hundred feet below the roof of the batholith.

Wall Rock Alteration. Wall rock alteration of the quartz-pyrite-tourmaline veins, and all of the base metal-bearing veins, appear to be essentially the same. The most conspicuous feature is the strong sericite zone which may extend for many feet into the walls of the veins. The width of the zone varies more or less with the width of the mineralized structure. Many of the veins, in fact most, have strong wall rock alteration even along the barren parts of the veins.

The wall rock alteration has been incompletely studied, and most of the study has been confined to areas where base metal mineralization is present. However, at the Daniel Stanton in the Rimini district, well below



any known base metal ore shoots, and in an area where the vein is composed of quartz, tourmaline and pyrite, sericite alteration is surrounded by a distinct argillic alteration zone which appears to be similar to the relations noted by Sales and Meyer (1948) at Butte. (Klepper, personal communication) At the Lilly mine in the Elliston district, a short cross-cut at the base metal bearing part of the main vein shows sericite alteration, with kaolinitic and montmorillonitic alteration of plagioclase in successive alteration envelopes. Incomplete alteration of biotite to chlorite apparently extends beyond the argillic zone. At a crosscut on the bottom level of the mine below the base metal ore shoots, and in an area where quartz, tourmaline and pyrite comprise a strong vein but barren of valuable minerals, the wall rock alteration story appears to be much the same. About 18 in of sericite alteration was observed. Because the mine had been full of water, and as pyrite is relatively abundant in the vein, it is difficult to be certain that the argillic alteration is entirely due to alteration of the wall rocks at the time of the formation of the vein.

There is certainly no obvious relation of wall rock alteration to base metal content of the veins in which the metals are found. In fact, it seems rather that the wall rock alteration is a function of mineralization at the time the vein system was produced and has little relation to the individual base metal ore shoots. On the surface, many of the large vein zones, such as the Comet-Gray Eagle vein in the Basin district, show intense sericite alteration across a zone possibly 100 ft wide. The wall rocks are, in some localities, disseminated with pyrite. Yet base metal concentrations appear to be restricted to small portions of the vein structures and are altogether absent along long parts of the complex vein zone.



Geochemical evidence of base metals in areas of strong wall-rock alteration. During an investigation of geochemical prospecting methods relative to the detection of mineralized structures and ore shoots, soil samples were collected across numerous vein structures in the Boulder batholith, principally in areas over known base metal ore shoots. However, additional soil sample traverses were made over wide vein structures where no base metals were known. It was expected that it would be possible to detect the presence of the vein structure by minor concentrations of base metals in the soils. Over strong veins in metamorphic and sedimentary rocks, base metal anomalies were recorded, even in areas of no known mineralization. (Robertson 1956 ?) A completely unexpected result of the investigation was that, over strong vein structures in the Boulder batholith, it was impossible, in many cases, to determine the position of a vein by the occurrence of base metal concentrations in the soils. The only conclusion which appears to be justified is that the veins did not have any base metals in them at the localities investigated. If the solutions which percolated through the vein systems did contain base metals, they left no trace. An alternate proposition that base metals were formerly present but hydrothermally leached requires complete leaching of all the base metals which may be in the realm of possibility but does not seem to be in the realm of probability.

At the Comet-Gray Eagle structure, which is one of the strongest and most intensely altered fracture zones in the northern part of the batholith, three soil sample traverses were made across the structure: one just east of the Comet Mine in an area where base metal ore shoots



were known to have come to the surface, or near to the surface; and two west of the Comet between the Comet and Gray Eagle, one at the top of the Divide between the two mines, both in areas where the location of the vein structures is visible in outcrops, exploration pits or bulldozer cuts. Although the precise center and width of the structures was unknown, the general position of the structure was indicated and the soil samples were collected along a traverse line long enough on each side of the zone to be certain to bracket the structure. Table 11 shows the geochemical results of the traverse across the known ore shoot, whereas tables 12 and 13 give the geochemical data from the two traverses across the barren parts of the structures. In the first table, it is clearly evident that a high concentration of base metals were leached from the veins, and high attenuations retained in the soils not only over the vein, but over a spread adjacent to the vein structure, a relationship found to be quite general over structures in the batholithic rocks. However, the other tables clearly demonstrate that lead, copper and zinc are absent to virtually absent over the vein structure. Some of these data are shown graphically in Figure 78. It is interesting to note that in Table 12 there is some arsenic across the vein zone. In Table 13 there is not even any indication of arsenic, and lead, zinc and copper are completely absent.

In the Elliston district, base metal analyses were made of soils collected over the tops of ore shoots at a number of veins. At the Ontario vein, lead, zinc and arsenic anomalies were recorded over the vein and on the downhill slope below the vein; likewise at the Lilly and

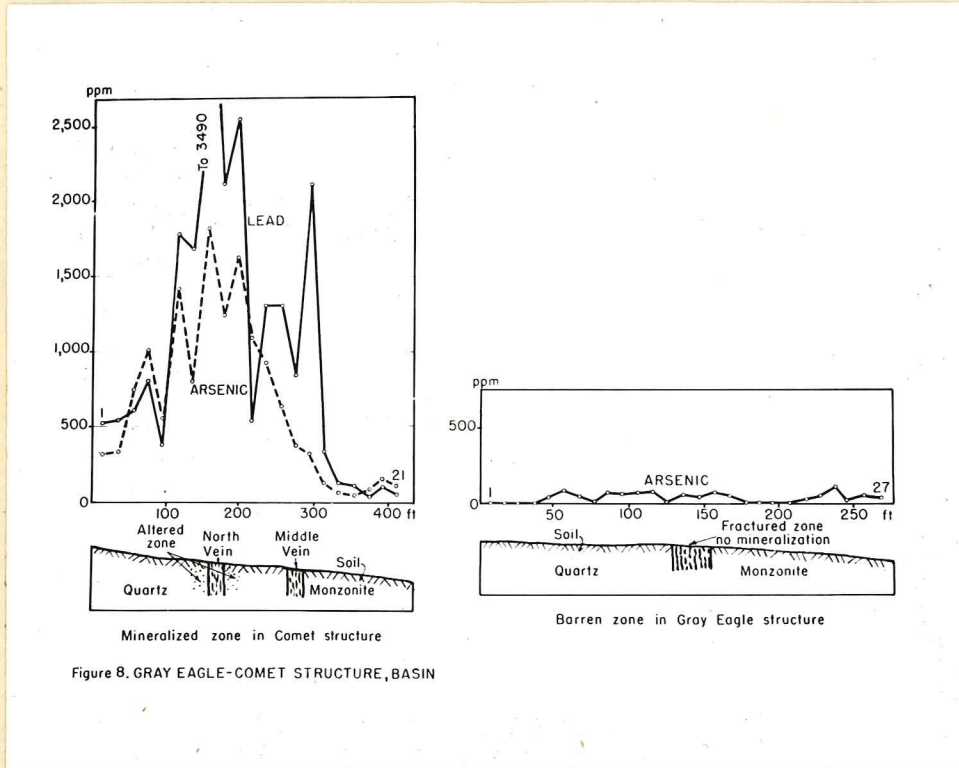


Figure 78. Geochemical soil sample traverses across mineralized zone at Comet mine, and across barren zone between Comet and Gray Eagle mines. (Robertson 1956? Figure 8)



Table 11

## Analyses of Geochemical Soil Samples at Comet Mine

Analyses of soil samples collected over mined-out area near the Comet shaft. Sample interval 20 ft, slope variable to 5°. Bearing of mineralized vein structures N. 70° W., traverse bearing N. 20° E.

(a) Colorometric chemical analyses

(b) X-ray spectrographic analyses

Sample Number	Location of Sample	Heavy Metals (a) ppm	Lead (b)	Zinc (b)	Remarks
1	120 ft N of N vein	150	565	112	
2	100 ft N of N "	200	589	215	
3	80 ft N of N "	250	1050	99	
4	60 ft N of N "	250	1350	357	
5	40 ft N of N "	175	897	25	
6	20 ft N of N "	300	1850	335	
7	Over north vein	400	1720	1310	
8	20 ft S of N vein	400	3490	494	
9	40 ft S of N "	250	2180	988	
10	60 ft S of N "	400	2620	751	
11	80 ft S of N "	200	577	530	
12	Over middle vein ?	250	1365	518	
13	" " "	250	1360	514	
14	140 ft S of N "	200	866	988	Possible contamination
15	160 ft S of N "	300	2170	341	" "
16	180 ft S of N "	200	350	152	
17	200 ft S of N "	150	123	120	
18	220 ft S of N "	100	100	90	
19	240 ft S of N "	100	62	74	
20	260 ft S of N "	100	127	144	
21	280 ft S of N "	100	96	98	

Table 11 (concluded)

Sample Number	Location of Sample	Copper (b) ppm	Arsenic (b) ppm	Manganese (b) ppm	Remarks
1	120 ft N of N vein	0	343	2000	
2	100 ft N of N "	0	384	1910	
3	80 ft N of N "	0	793	1910	
4	60 ft N of N "	0	1040	1790	
5	40 ft N of N "	0	581	1000	
6	20 ft N of N "	0	1460	1940	
7	Over north vein	0	830	2030	
8	20 ft S of N vein	0	1870	1670	
9	40 ft S of N "	0	1270	4190	
10	60 ft S of N "	0	1685	4340	
11	80 ft S of N "	0	1130	4380	
12	Over middle vein ?	0	969	5000	
13	" " "	0	685	2590	
14	140 ft S of N vein	0	400	1760	Possible contamination
15	160 ft S of N vein	0	369	2020	" "
16	180 ft S of N vein	0	146	1570	
17	200 ft S of N "	0	108	1760	
18	220 ft S of N "	0	69	1660	
19	240 ft S of N "	0	82	1705	
20	260 ft S of N "	0	172	2200	
21	280 ft S of N "	0	128	1380	



Table 12

Analyses of Geochemical Soil Samples Across Barren Zone  
between Comet and Gray Eagle Mines

Analyses of soil samples collected across strong fracture zone, but in an area of no known commercial mineralization north of the Gray Eagle mine. Sample interval 20 ft, 315 ft N. 28° W. of discovery shaft. Slope variable from horizontal to 5°. Strike of vein structure N. 70° W. Traverse bearing approximately N. 15° E. Grassy slopes, soil 4 - 6 in deep.

(a) Colorometric chemical analyses

(b) X-ray spectrographic analyses

Sample Number	Location of Sample	Heavy Metals		Zinc (b) ppm	Remarks
		(a) ppm	Lead (b) ppm		
1*	200 ft N of structure	50	0	57	
2*	180 ft N of "	100	38	31	
3*	160 ft N of "	50	0	0	
4*	140 ft N of "	50	0	0	
5	120 ft N of "	100	0	0	
6	100 ft N of "	100	0	0	
7	80 ft N of "	50	0	0	
8	60 ft N of "	50	0	0	
9	40 ft N of "	50	0	0	
10	20 ft N of "	50	0	0	
11	N side of vein zone ?	50	19	0	Small prospect
12	Over fracture zone	50	12	0	pits between
13	" " "	50	13	0	8 and 12
14	S side of zone	50	0	0	
15	20 ft S of structure	50	0	0	
16	40 ft S of "	50	0	0	
17	60 ft S of "	50	0	0	
18	80 ft S of "	100	0	0	
19	100 ft S of "	50	21	0	
20	120 ft S of "	100	52	14	
21	140 ft S of "	100	52	0	
22	160 ft S of "	50	30	0	
23	180 ft S of "	50	31	0	
24	200 ft S of "	50	0	0	
25	220 ft S of "	50	60	0	
26	240 ft S of "	50	12	0	
27	260 ft S of "	50	18	0	
28	280 ft S of "	50	40	0	
29	300 ft S of "	50	23	0	
30	320 ft S of "	50	0	0	
31	340 ft S of "	50	37	0	

\* Not shown in Figure 78.

Table 13

Analyses of soil samples collected over strongly fractured and altered zone between Comet and Gray Eagle Mines. Sample interval 20 ft. Slope variable from 20° to 25°. Strike of structure approximately N. 80° W. Bearing of traverse N. 20° E. Thick soils for 500 ft then thin granitic sands.

## a. Chemical analyses

## b. X-ray spectrographic analyses

Sample No.	Location of sample	Heavy Metals (a) ppm	Lead (b) ppm	Zinc (b) ppm	Remarks
1	240 ft N. of bulldozer cut	50	0	0	
2	220 ft "	50	0	0	
3	200 ft "	50	0	0	
4	180 ft "	50	0	0	
5	160 ft "	100	0	0	
6	140 ft "	100	0	0	
7	120 ft "	50	0	0	
8	100 ft "	50	0	0	
9	80 ft "	50	0	0	
10	60 ft "	50	0	0	
11	40 ft "	100	0	0	
12	20 ft "	100	0	0	
13	Edge of bulldozer cut	50	0	0	Bulldozer cut in altered zone
14	20 ft S of bulldozer cut	50	0	0	
15	40 ft "	50	0	0	
16	60 ft "	50	0	0	
17	80 ft "	50	0	0	
18	100 ft "	50	0	0	
19	120 ft "	50	0	0	
20	140 ft "	50	0	0	
21	160 ft "	50	0	0	
22	180 ft "	50	0	0	
23	200 ft "	50	0	0	
24	220 ft "	50	0	0	
25	240 ft "	50	0	0	
26	260 ft "	50	0	0	



Table 13 Continued

Sample No.	Location of Sample	Copper (b) ppm	Arsenic (b) ppm	Manganese (b) ppm	Remarks
1	240 ft N of dozer cut	0	60	952	
2	220 ft "	0	127	1030	
3	200 ft "	0	100	1240	
4	180 ft "	0	60	1020	
5	160 ft "	0	100	1105	
6	140 ft "	0	97	966	
7	120 ft "	0	10	923	
8	100 ft "	0	100	1055	
9	80 ft "	0	0	958	
10	60 ft "	0	18	1170	
11	40 ft "	0	18	1060	
12	20 ft "	0	68	1240	
13	Edge of "	0	62	874	Bulldozer cut in altered zone
14	20 ft S "	0	71	972	
15	40 ft "	0	117	902	
16	60 ft "	0	85	1015	
17	80 ft "	0	97	1030	
18	100 ft "	0	67	916	
19	120 ft "	0	110	511	
20	140 ft "	0	112	1070	
21	160 ft "	0	52	818	
22	180 ft "	0	46	784	
23	200 ft "	0	43	720	
24	220 ft "	0	77	1010	
25	240 ft "	0	31	930	
26	260 ft "	0	68	1055	

Bullion Mines anomalies were recorded over ore shoots. Even at the Copper King, a small copper anomaly was recorded over the vein and on the downslope side of the vein. However, at the Julia Mine, three soil-sample traverses were made across the vein structure, which is one of the large veins in the district, and one which is easily traceable on the surface because of the large number of prospects which expose the vein structure. The soil sample traverses were not taken at any locality where ore shoots were known to come to the surface, but it was anticipated that it might be possible to trace the structure in the relatively barren parts of the zone, in which case it would have offered a tool which might have served to trace the structure through a heavily timbered area where the vein had not been traced on the surface. One of the traverses was made west of the Julia shaft just west of the quartz monzonite-andesite contact in the andesitic rocks along the edge of a bull dozer cut in which stringer veins of the structure were clearly visible. (See mineral deposits for details.) The other two were taken, one east of the shaft and the other east of the lower adit level. No detectable anomalies were recorded on any traverses. The background was about 100 ppm heavy metal, with a few individual samples, not in any way clearly related to the vein structures, of 125 to 150 ppm. In the manuscript submitted for publication the following statement is taken: "The data from these three traverses add further evidence that the presence of a large vein structure with strong wall-rock alteration does not mean that the vein-forming solutions contained significant amounts of base metals. In this case, as at the Comet-Gray Eagle structure, it is doubtful whether the vein structures can be traced by geochemical



methods." (Robertson 1956 ?)

Distribution of base metal deposits. In the northern part of the Boulder batholith, and particularly in an area west of the axis of the batholith (as it is presently exposed), base metal-bearing veins are found. There are literally hundreds of veins which are moderately widely spaced. In the Elliston district, there is a reasonably systematic spacing of about one vein every one-half mile. The base metals are found in small structures which are distinctly controlled by early, barren vein structures. In no case has vertical and lateral persistence of a wide base metal ore shoot been demonstrated. Rather, the base metals, zinc and lead, are found in pods and lenses, and occasionally in minable ore shoots in veins found over an area of possibly 500 sq mi. In this area, base metal ore shoots cannot be demonstrated to persist to depths of more than 1000 to 1500 ft below the top of the batholith.

Insufficient data are available to demonstrate the concentrations of the base metals in the quartz monzonites and in the related quartz porphyry dikes. Geochemical investigations indicate that traces of zinc and lead are found throughout the area of the granitic rocks, whereas the concentrations of copper are very low to absent.

Conclusions. Data have been presented which appear to contribute to an understanding of vein formation, wall-rock alteration, and derivation of the metals and minerals in the veins.

There are apparently two lines of evidence leading to vein formation. They are:

1. Segregations of tourmaline and pyrite with stable feldspar selvage zones which lead to silica release in the feldspar destruction stage in the formation of quartz-tourmaline-pyrite veins.
2. Quartz veinlets in one crystal wide strings and as late segregation veinlets with tourmaline in alaskite. These may develop into quartz veins.

Wall rock alteration to sericite adjacent to the vein (and argillic alteration between sericite zone and fresh quartz monzonite) is produced along:

1. Barren quartz + tourmaline + pyrite veins.
2. Barren quartz + pyrite veins.
3. Quartz veins + zinc + lead + copper.
4. Carbonate veins + quartz.

There is no clear association of wall-rock alteration to base metals. In fact, the evidence appears to favor the hypothesis that the alteration is produced irregardless of the mineral composition of the vein but is more intense adjacent to the earliest vein zones.

The alteration may be regarded in two ways:

1. Auto-alteration of wall-rock minerals at the feldspar destruction or hydrothermal stage of magmatic crystallization at which time potash released from biotite and/or feldspar goes to form sericite.



2. Alteration due to egress of solutions from the vein into the walls.

In the latter case, the evidence seems to favor the hypothesis that the solutions were derived from late magmatic processes operating in the next lower segments of crystallizing quartz monzonite.

The source of iron, sulfur, and boron is demonstrated to be by local derivation from the walls of segregated tourmaline and pyrite in the deuteric stage with stable feldspar selvages. Local derivation is indicated for the tourmaline-pyrite joint coatings which have the same potash-rich feldspar stable selvage zone between the tourmaline and the quartz monzonite. With the same minerals present in the quartz veins in a feldspar unstable environment where sericite is produced, and silica is released, the evidence seems to favor the hypothesis that much of the iron, sulfur and boron were locally derived from the walls at a particular point of the vein, or from hydrothermal solutions derived from the walls of the next segments of crystallizing quartz monzonite. To consider that the iron, boron and sulfur were derived from great depth ignores the lines of evidence presented.

The possible source of much of the silica has been indicated. Exudations from the walls as a result of the destruction of feldspars to sericite is clearly indicated for very small stringers. Segregations into veinlets in the late crystallization of alaskite has also been indicated. Some of the quartz may be introduced by hydrothermal solutions, but the source of the silica need not be considered to have been derived from depths greater than the next lower segments of crystallizing quartz monzonite.

The source of the zinc, lead and copper is less clear. As veins with these metals are found over a very wide area in small shoots in widely spaced veins, and as these metals are found in minor amounts in the quartz monzonite, as indicated by the traces found by geochemical methods over fresh quartz monzonite generally; and as these minerals are found in quartz porphyry dikes as indigenous minerals; and as the wall-rock alteration adjacent to the base metal-bearing veins is the same as in the earlier barren veins, these minerals may be considered to have been locally derived. To consider that the quartz-tourmaline-pyrite veins are in large part locally derived as a result of late magmatic processes, and then to consider that the base metals are derived from sources of indefinitely deeper depth would suggest two different sources for the hydrothermal solutions which would yield the same wall-rock alteration phenomena, and form in veins which coincide with, or which are controlled by the location of the early-formed locally-derived vein structures. I think that the evidence favors very serious consideration of the hypothesis that the metals were derived from the quartz monzonite in the latest crystallization stage in the late part of the hydrothermal stage. Thus from the crystallization of the first olivine crystal in the magma to the last crystal of ore minerals, there appears to be indicated a succession of mineral formation stages which are controlled by the temperature of the cooling magma, and in part localized by fractures which form late in the crystallization history of the batholith.



### Dikes and Sills

Two stages of intrusive sills and dikes are recognized in the district. In some cases it is not possible to positively identify association to the volcanic pile or the batholithic stage.

#### Dikes In Volcanic Pile.

In the volcanic pile, sills are not uncommon. In the Little Blackfoot Creek section, (Table 4, p. 26) units 2, 4 and 7 are intrusive and sill-like. At the same locality, a dike of diorite porphyry cuts the volcanic sequence.

The hornblende andesite sill (Unit 2, Table 4) contains automorphic hornblende crystals to 10 mm in a very fine-grained groundmass of plagioclase with minor amounts of potash feldspar (?) and quartz.

Sills of porphyritic basalt with pyroxene phenocrysts is petrographically identical to the basalt flow of the same mineralogy. (See Emery basalt). Unit 4, Table 4 in the Little Blackfoot Creek section and a thin sill on the east side of Nigger Mountain in the volcanics above the roof of the batholith, are composed of porphyritic basalt.

Two andesite porphyry sills were noted in the sedimentary sequence, one northeast of the Railroad Station at Elliston and another on the west side of Elliston Creek in the Kootenai formation. The rocks in hand specimen are light brown to dark tan in which conspicuous plagioclase phenocrysts to 6 mm occur. Small knots of fine-grained mafic minerals, mostly altered hornblende, are megascopically visible. In thin section andesine and hornblende altered to chlorite are the distinguishable minerals in an aphanitic groundmass.

## Dikes in Batholithic Rocks.

A diorite porphyry sill crops out at the top of the small mountain in SE  $\frac{1}{4}$  sec. 12, T. 8 N., R. 7 W., and is Unit 7 Table 4. Plagioclase phenocrysts comprise nearly 50 per cent of the rock. Fine-grained mafic minerals are also recognized in the hand specimen. In thin section, the rock resembles the very fine-grained quartz monzonite porphyry. (See petrographic description of fine-grained quartz monzonite.)

A granodiorite porphyry dike was found on the west side of Bison Mountain (near cen. sec. 31, T. 8 N., R. 6 W., ). The dike strikes N. 75° W., and is vertical. The dike is about 100 ft wide, and shows distinct lineation of hornblende phenocrysts parallel to the walls. Epidote is abundant along the joints of the dike. The south contact of the dike is marked by a fracture zone which is about 20 ft wide in which the andesite wall rocks are intensely bleached.

In thin section, andesine phenocrysts are 1 to 3 mm long and are irregular in outline because of partial replacement by orthoclase. Quartz and potash feldspar are in the fine-grained groundmass. Hornblende, with **relics** of augite, is in ragged grains, whereas biotite is in hypauto-morphic crystals. An approximate modal analysis is as follows:

Andesine	50
Orthoclase	20
Quartz	10
Hornblende	12
Augite	3
Biotite	4
Magnetite	1



### Contact Metamorphism

The rocks intruded by the batholith are not strongly altered in the district, in contrast to the high-grade hornfelses found in contact with granitic rocks in the Helena and Marysville districts (Knopf 1913, p. 31; Barrell 1907). At most localities, the volcanic rocks at the contact are incompletely hornfelsed. At three localities, two of which overlie alaskites, the volcanic rocks are microaplite in composition. At one locality not at the contact, Colorado shale is a green, spotted, incipient hornfels.

#### Hornfelsic Contact Metamorphism

The common contact metamorphism of the volcanic rocks produced a dense, flinty, incipient hornfels. Most of the volcanic rocks intruded are andesite or basic andesite and are presumably located near the base of the volcanic pile. South of the Julia Mine, dark gray andesitic rocks are changed to brittle hornfels for about 40 to 50 ft above the contact. In thin section, the rock is observed to be recrystallized from its original felted texture to a very fine-grained hornfelsic texture. Plagioclase phenocrysts are partly altered to albite-oligoclase, and partly to fine-grained epidote and calcite. Former biotite crystals are recrystallized, and in the process, small grains of sphene were formed around the margin of the crystals. The former pyroxene was altered to a micaceous mineral, light green in color, whose birefringence is greater than normal chlorite. Figure 79 shows the texture of the hornfelsed volcanic rock.



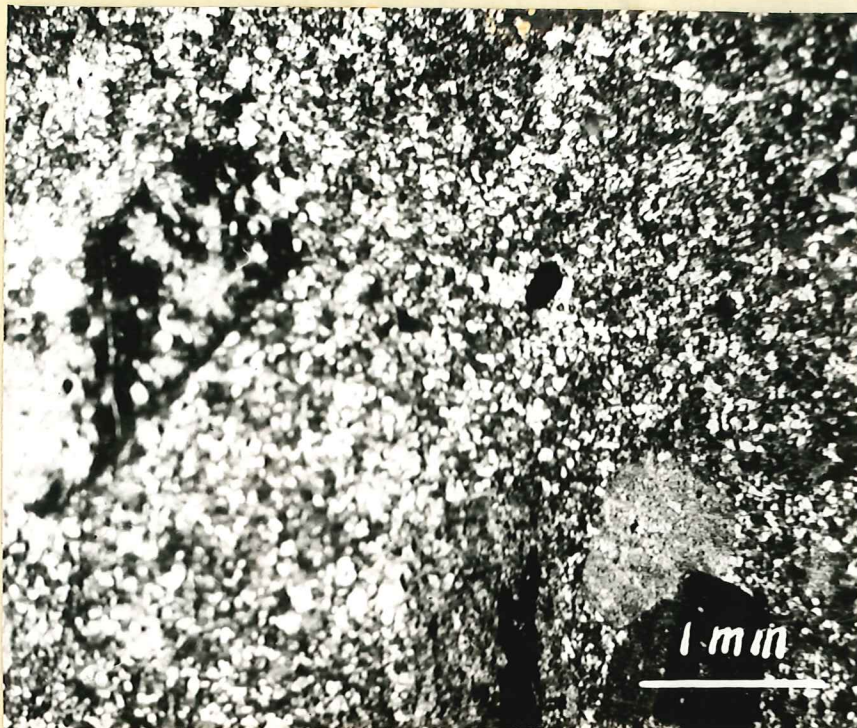


Figure 79. Photomicrograph of hornfelsic texture in andesitic volcanic rock near contact of the batholith. Crossed nicols.



Figure 80. Intrusive contact of quartz monzonite into andesitic volcanic rocks, about 1 mi west of MacDonald Pass.



A similar type of hornfelsing was noted at the base of the air beacon on the Continental Divide, south of MacDonald Pass. The rock contains a small amount of potash feldspar and may be a latite. A few miles farther south, on the east side of Mike Renig Gulch, a quartzitic water-laid tuff occurs. The tuff is presumably close to the contact, and was metamorphosed to a dense, brittle hornfels.

At the contact exposed in the road cut on the west side of MacDonald Pass, the andesitic volcanic rock is bleached to a buff color, a few inches to a few feet above the contact. Some blocks of andesite, stoped away from the contact, are bleached throughout. The unbleached rocks above the contact are incompletely hornfelsed. Figure 80 shows the contact exposed in a road cut.

Quartz monzonite under incipiently hornfelsized volcanic rocks does not show strong endomorphic alteration. The mineralogy of the alteration products in the metamorphosed volcanics is the same as in deuterically altered volcanic rocks. The feldspar alteration to albite-oligoclase and epidote is characteristic of the deuteric stage crystallization in quartz monzonite.

#### Microaplitic contact Metamorphism

The most conspicuous contact alteration is to microaplite, a distinctive type of metamorphism found at three localities, two of which overlie bodies of replacement alaskite, and the other above endomorphically altered batholithic rocks.

The microaplite alteration is called "altered cretaceous volcanics" and mapped as such by the U. S. Geological Survey field party. (Ruppel,

personal communication.) It is a very fine-grained, buff to nearly white rock which contains many small miarolitic cavities. Pyrite is commonly found as small crystals in the cavities. On the outcrop, the rock surrounding the small vugs are stained with iron oxide for several millimeters. Tourmaline is also commonly found in the vugs and crystallized with quartz in radiating clusters of crystals. Southeast of the Ontario Mine, float boulders of microaplite contain layers of tourmaline. In hand specimen, small irregular lenses of white, dense matter has the same shape and size of small "wispy" lenses found in the welded tuffs. Small plagioclase phenocrysts are inconspicuous, but in some of the micro-aplite are relatively abundant.

In thin section, the rocks are seen to be composed dominantly of sericite, potash feldspar, quartz and relics of plagioclase and mafic phenocrysts. The plagioclase phenocrysts are indicated by relics of plagioclase in masses of sericite, and in other cases, are suggested by masses of sericite which have the shape of plagioclase phenocrysts. Potash feldspar crystals are rarely as large as 0.5 mm, some are conspicuously perthitic, others show indistinct microcline-like twinning and may be anorthoclase. The mafic minerals in the original volcanic rocks are largely eliminated. Patches of a white micaceous mineral which is similar to sericite, but which has a lower relief and lower birefringence (about .008 to .010) than sericite, may represent former mafic grains. The micaceous mineral does not stain in sodium cobaltinitrite solution, suggesting that it contains little or no potash. Films of iron oxide, dusty opaque grains, and minute grains of leucoxene, are commonly found in patches which indicate the location of former mafic minerals.



The miarolitic cavities commonly have coarser grained quartz lining the cavities than quartz in the groundmass. Tourmaline is found in some vugs, and in replacement clusters in the rock. The tourmaline is somewhat lighter in color, pleochrism o = bluish green, e = light tan, than tourmaline in the segregation pods and hydrothermal veins. Some tourmaline is strongly zoned with lighter tourmaline in the core surrounded by a thin rim of darker tourmaline similar to vein tourmaline, pleochroic, o = dark bluish green, e = light olive brown.

The textures in the aplitized volcanic rocks are many and varied. Individual thin sections may exhibit a considerable range in textures across the length of the slide. The normal groundmass texture is somewhat hornfelsic in which quartz and feldspar occur in intricate intergrowths. Hornfelsic textures are partly recrystallized to microgranitic texture in which quartz and feldspar occur in microstuctured grains, micropoikilitic intergrowths and xenomorphic grains. Quartz exhibits the widest range in grain size from very small grains barely resolved by the microscope to coarse grains in segregations clusters up to 0.5 mm. Quartz also is found in elongated pods, thin discontinuous layers and in small vein-like bodies, the latter contain tourmaline in some examples. Figure 81 shows one of the textures in microaplitic rock.

The volcanic ancestry of the microaplite is clear, however the composition of the original rock is in doubt. From the best evidence available in this district, the batholith intruded the volcanic pile at or reasonably near the base of the volcanic pile. The lower units of volcanics in the Elliston district are dominantly basic andesitic breccias and andesite tuffs. Consequently, in areas where andesitic volcanics are found, and where the volcanic rocks are altered to microaplite, it was considered that the



original rock was of andesitic composition.

In regions south and east of the Elliston district, similar microaplitic volcanic rocks have been traced laterally into latite. Chemical analyses of fresh and altered rocks have demonstrated that there is very little difference in fundamental chemical composition between the two rocks. Consequently, the microaplitic rocks in those areas are considered to be the product of thermal recrystallization without significant chemical transfer. (Ruppel and Klepper, personal communication)

Both Klepper and Ruppel have examined the microaplite in the Elliston district, and both tentatively conclude that it represents recrystallized latite. I am not certain that the original rock was latite. It seems to be too much of a coincidence to have the only latite at the contact in this area over alaskite or quartz monzonite which shows deuteric differentiation. I think the question should be left open until additional chemical analyses become available.

The problem is, is the amount of chemical transfer negligible, or small but definite? Table 14 lists a rapid chemical analysis of microaplite collected about 100 yards south of the Pauper adit. (Chapman, R. W., unpublished manuscript) If the original rock was a latite whose composition was similar to the latite in Table 7 p. 54, silica and potash must have been added. That this is a reasonable possibility is indicated by the mineralogical composition of the rock. Quartz is abundant, and sericite is present in large amounts. In addition, boron and sulfur must have been added to account for pyrite and tourmaline. The same additions are required in case the original rock was andesite. (See Table 6, p. 52) The relatively greater lime content in the microaplite than in the analysis of latite is certainly suggestive that the original rock was more calcic than latite.



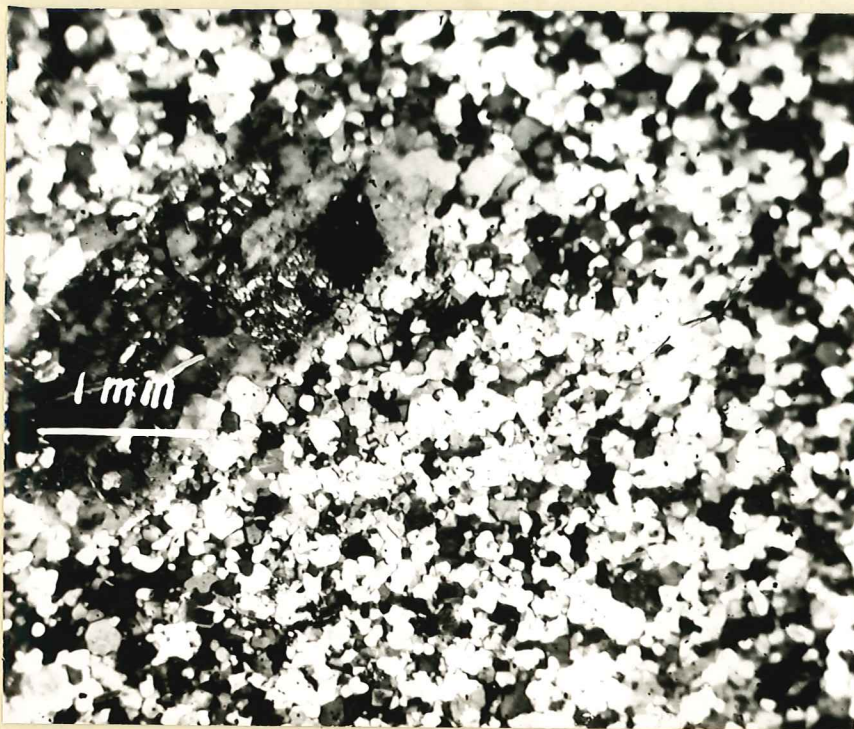


Figure 81. Photomicrograph of microaplitic texture in microaplite formed by contact metamorphism of andesitic or latitic volcanic rock. (Scale should be 0.5 mm) Crossed nicols.

Table 14

Chemical Analysis of Microaplite  
(altered volcanic rock)

	Microaplite	Latite	Andesite
SiO <sub>2</sub>	73.8	64.45	59.64
Al <sub>2</sub> O <sub>3</sub>	13.8	17.69	16.64
Fe as Fe <sub>2</sub> O <sub>3</sub>	.94	1.33 1.93	2.33 4.35
MgO	.16	.57	2.10
CaO	.99	3.73	4.59
Na <sub>2</sub> O	3.2	3.85	3.31
K <sub>2</sub> O	6.0	3.68	4.16
TiO <sub>2</sub>	.28	.69	1.08
P <sub>2</sub> O <sub>5</sub>	.04	.16	.49
MnO	.08	.05	tr.
Ignition	<u>.30</u>		
	100.00		
FeO	.51		
Fe <sub>2</sub> O <sub>3</sub>	.38		

A rapid analysis, by U. S. Geological Survey. From unpublished manuscript of R. W. Chapman.

Microaplitized volcanic rock of contact metamorphic origin, from dump of prospect about 200 yards south of Pauper workings.

Sample no. 31-111-2



From the chemical analysis of the microaplite from the Elliston district, and from the petrographic characteristics of the rock, I am of the opinion that the original rock, be it latite or andesite, was more than thermally rearranged. The association of underlying alaskites at two localities and of evidence of strong endomorphic phenomena in quartz monzonite at the other, namely the Pauper area, indicate that the materials needed to produce microaplite were precisely the same materials which transformed quartz monzonite at the top of the batholith to alaskite. It seems reasonable to consider that some of the residual deuteric or hydrothermal solutions which produced profound chemical and mineralogical changes in the batholithic rocks may have leaked into the surrounding volcanic rocks.

The mineralogy of the metamorphism, marked by the abundance of sericite, is the same as the mineralogy of the hydrothermal stage in the crystallization of the batholith. Just as sericite produces the very marked and distinctive changes in rocks adjacent to hydrothermal veins, so the conspicuous contact metamorphic rocks are the sericite-bearing rocks.

#### Thermal Stage of Metamorphism

The different types of contact metamorphism in the volcanic rocks appears to be similar, in several respects, to the different end stage crystallization phenomena in the batholithic rocks. The hornfelsic rocks have the same minerals found in the deuteric stage, and the sericite-bearing microaprites have the same mineral assemblage found in alteration of feldspar at the hydrothermal stage. Small differences in the temperature of formation produce pronounced effects comparable to the differences in mineralogy found in the deuteric stage veinlets and hydrothermal veins.

### Age of the Batholithic Rocks

The mapping in the Elliston district did not provide any additional evidence regarding the age of the batholith. From field evidence elsewhere in the batholith, quartz monzonite has intruded upper Cretaceous volcanics subsequent to their deformation by folding and thrust faulting. Knopf (1913, p. 34) considered the batholith very late Cretaceous, and Chapman (1955) indicates possibly very early Tertiary. Knopf (1913) determined that early Oligocene beds overlie eroded granitic rocks of the batholith near Blossburg. From the best evidence, the batholith is late Upper Cretaceous to very early Tertiary.

Chapman et. al. (1955) determined the age of the batholith by the zircon method of Larson (1952) as well as rocks from the Philipsburg and Idaho batholiths. The data are summarized in Table 15. The average of the 5 analyses, including a monazite age determination, is 67 million years.

### Chemical Composition of Intrusive Rocks

There are no chemical analyses of intrusive rocks from the Elliston district, however there are analyses of fresh and altered quartz monzonite from the adjacent Rimini district, and analyses from the Boulder Mountains southeast of Elliston. The published chemical analyses of gabbro from the Elkhorn district, quartz monzonite from Butte and other localities in the Boulder batholith, and aplite from the batholith are given in Tables 16 and 17.



Table 15

## Zircon and Monzonite Age Determinations

Boulder Batholith	Quartz Monzonite	Zircon	64 MY
	Quartz Monzonite	Zircon	71 MY
	Quartz Monzonite	Zircon	69 MY
	Alaskite	Zircon	61 MY
	Alaskite	Monzonite	72 MY
Philipsburg Batholith	Quartz Monzonite	Zircon	50 MY
Idaho Batholith	Gneissoid quartz monzonite	Zircon	54 MY
	" "	Monzonite	51
	" "	Zircon	72 <sup>1</sup>
	" "	" Zircon	100 <sup>2</sup>

1. Determination by Howard W. Jaffe on a "monzonite from the same gneissoid taken farther west along Lost Horse Creek" (as first two analyses).
2. Unpublished data by Larsen, Gottfried, Jaffe, from other rocks from the Idaho Batholith.

Table 16

## Chemical Analyses of Aplite

	I	2	3	4
SiO <sub>2</sub>	77.05	87.87	76.26	74.61
Al <sub>2</sub> O <sub>3</sub>	12.84	12.52	12.94	13.36
Fe <sub>2</sub> O <sub>3</sub>	.56	.67	.69	.83
FeO	.14	none	.13	.36
MgO	tr.	.09	.17	.35
CaO	.57	.49	1.10	1.35
Na <sub>2</sub> O	2.81	2.47	2.73	2.49
K <sub>2</sub> O	5.52	5.78	5.26	5.70
H <sub>2</sub> O <sup>+</sup>	.48	.52	.43	.42
H <sub>2</sub> O <sup>-</sup>	.22	.25	.17	.19
TiO <sub>2</sub>	.12	.11	.11	.11
P <sub>2</sub> O <sub>5</sub>	none	.05	.06	.06
MnO	none	tr.	tr	tr
BaO	none	none	.07	.10
SrO	none	none	none	tr
LiO <sub>2</sub>	tr	tr		
Cl		none		
CO <sub>2</sub>			none	none
S			.01	tr.
Cr <sub>2</sub> O <sub>3</sub>			none	none
NiO			none	none
	100.31	99.82		99.93



## Table 16 Continued

1. Aplite, Nettie Mine, Butte. (Weed 1899, p. 739)
2. Aplite, Nettie Mine, Butte. (Weed 1899, p. 739)
3. Aplite, Black Butte, Elkhorn district. (Barrell 1901, p. 540)
4. Aplite, Elkhorn district. (Barrell 1901, p. 540)

Table 17

## Chemical Analyses of Quartz Monzonite

	1	2	3	4	5	6	7	8	9
SiO <sub>2</sub>	67.12	65.91	64.31	64.05	64.34	63.88	63.87	61.64	64.17
Al <sub>2</sub> O <sub>3</sub>	15.00	15.32	15.44	15.38	15.72	15.84	15.39	15.63	15.25
Fe <sub>2</sub> O <sub>3</sub>	1.62	2.28	2.43	2.20	1.62	2.11	1.93	3.39	2.16
FeO	2.23	2.02	2.58	2.74	2.94	2.59	3.08	2.69	2.98
MgO	1.74	1.52	2.23	2.08	2.17	2.13	2.23	2.82	2.60
CaO	3.43	3.28	4.22	4.30	4.24	3.97	4.30	4.90	4.24
Na <sub>2</sub> O	2.76	3.08	2.17	2.74	2.76	2.81	2.76	2.64	2.62
K <sub>2</sub> O	4.52	4.80	4.09	4.00	4.04	4.23	4.18	3.72	4.34
H <sub>2</sub> O <sup>+</sup>	.58	.60	.69	.83	.76	.66	.69	.91	.28
H <sub>2</sub> O <sup>-</sup>	.09	.60	.19	.27	.25	.22	.19	.29	.16
TiO <sub>2</sub>	.48	.59	.71	.60	.53	.65	.65	.71	.67
P <sub>2</sub> O <sub>5</sub>	.15	.18	.22	.21	.14	.21	.17	.21	.16
MnO	.06	tr	tr	.11	.12	.07	.11	.04	.04
BaO	.07	.10	.07	.08	.06	.09	.07	.08	.07
SrO	.03		tr	.04	.03	.02	.04	.04	tr
Li <sub>2</sub> O	tr		tr						
Cl				.07	.03		.07		
CO <sub>2</sub>	none	.21				.34			none
S		.02	tr	.35	.03		.15	none	none
Cr <sub>2</sub> O <sub>3</sub>	none		none				tr		tr
NiO		none							
SO <sub>3</sub>	none								
Total	99.88	100.51	99.97	100.06	99.80	99.82	99.91	99.70	100.18



## Table 17 Continued

## Chemical Analyses of Quartz Monzonite

1. Granite (quartz monzonite), Boulder, Butte District (Weed 1912, p. 74)
2. Quartz monzonite, Rimini (Knopf 1913, p. 30)
3. Quartz monzonite, Elkhorn (Barrell 1901, p. 536)
4. Quartz monzonite, Gagnon Mine, Butte (Weed 1912, p. 33)
5. Quartz monzonite, Atlantic Mine, Butte (Weed 1912, p. 33)
6. Quartz monzonite, Walkerville Station, Butte (Weed 1912, p. 33)
7. Quartz monzonite, Alice Mine, Butte (Weed 1912, p. 33)
8. Quartz monzonite, Red Rock Creek, Boulder Mountains (Weed 1912, p. 33)
9. Quartz monzonite, Frohner Mine, south of Helena, Boulder Mountains (Weed 1912, p. 33)

Table 17

## Chemical Analyses Gabbro and Altered Quartz Monzonite

	1	2	3	4
SiO <sub>2</sub>	47.44	49.42	66.70	66.02
Al <sub>2</sub> O <sub>3</sub>	18.21	13.58	13.25	14.14
Fe <sub>2</sub> O <sub>3</sub>	5.37	3.98	1.34	1.53
FeO	6.05	3.59	.51	.37
MgO	5.53	5.06	.30	.67
CaO	9.98	18.15	.05	.26
Na <sub>2</sub> O	2.58	2.22	.39	.39
K <sub>2</sub> O	1.17	1.42	4.03	4.63
H <sub>2</sub> O <sup>+</sup>	1.08	.50	.37	.48
H <sub>2</sub> O <sup>-</sup>	.13	.28	.08	.10
TiO <sub>2</sub>	1.38	.71	.20	.56
P <sub>2</sub> O <sub>5</sub>	.68	1.38	.12	.17
MnO	.06		tr	tr
BaO	.08		none	.04
SrO	.08			
S	.03			
F				.02
CO <sub>2</sub>	none		.21	.25
FeS <sub>2</sub>			5.09	6.73
FeAsS			4.75	.72
PbS			.59	2.02
ZnS			1.86	.74
Total	99.86	100.29	99.84	99.94



## Table 17 Continued

## Chemical Analyses of Gabbro and Altered Quartz Monzonite

1. Gabbro, Black Butte, Elkhorn district (Barrell 1901, p. 514)
2. Gabbro, Elkhorn district (Weed 1910, p. 93)
3. Altered Quartz Monzonite, Rimini (Knopf 1913, p. 30)
4. Altered quartz Monzonite, Rimini (Knopf 1913, p. 30)

## STRUCTURE

Structural deformation in the Elliston district has been comparatively mild. Over most of the district, the sequence of events appears to be about as follows: (1) very mild warping occurred prior to the outpouring of Late Cretaceous volcanic rocks; (2) sedimentary and Upper Cretaceous volcanic rocks were tilted prior to and/or possibly during the intrusion of the batholith; (3) the batholith was intruded without deformation definitely attributable to the intrusion; and (4) faulting, jointing and vein formation occurred in volcanic and batholithic rocks alike.

In the northwestern part of the district in the vicinity of Elliston, the Paleozoic and Mesozoic sedimentary rocks dip about  $45^{\circ}$  northwest, and strike northeast. The relation of these sedimentary rocks to the Upper Cretaceous volcanic rocks is not definitely known except that they are unconformable on the sedimentary rocks. The exposures are particularly poor in this area.

South of the northeast striking fault which extends across the northern part of the district, Upper Cretaceous volcanic rocks rest unconformably upon upper Kootenai and lower Colorado beds, the stratigraphic interval of beds cut out along the unconformity is only a few hundred feet. The inclination of the beds below the unconformity in the region of the Elliston district is considered westerly at a very small angle. The hiatus is relatively insignificant in this area, but farther to the west, the volcanic rocks rest on beds 1,500 ft or more higher in the stratigraphic section. Regionally, the volcanic rocks rest on rocks as old as Madison limestone of Mississippian age.



Prior to and/or during the intrusion of the batholith, the sedimentary and volcanic rocks were tilted into generally westerly dips ranging from  $5^{\circ}$  to  $30^{\circ}$  with few evidences of reversal. There are, however, insufficient exposures of sedimentary rocks and inadequate indications of strike and dip in the andesitic volcanic rocks to more than indicate general westerly dips. Just west of the district the andesitic and basaltic volcanic rocks show quite consistent westerly dips at relatively small angles.

The mildness of the structural deformation in the Elliston district is in striking contrast to the strongly folded, overturned and thrust faulted structures in the pre-batholithic rocks on the east side of the batholith.

#### Structural Relations of Contacts of the Batholith

The roof of the batholith within the district dips gently to the west wherever the contact relations can readily be observed. The relations are indicated in the structure sections which accompany the geologic map. At two localities, the batholith appears to lens out into sill-like bodies at the margin of the batholith.

#### Roof of the Batholith

In the northeast corner of the district from MacDonald Pass to the mouth of Mike Renig Gulch (about 3 mi southeast of Elliston) a gentle westerly dip of about  $7^{\circ}$  on the roof of the batholith is indicated, if the granitic rocks are continuous under the roof. The air beacon just south of the highway on the pass is on a flat roof remnant of

hornfelsed volcanic rocks and on the knob north of the highway and the contact of the batholith all but follows the contours around the base of the knob.

The contact is particularly well exposed in a highway cut about 1 mi west of the Divide. Figure 80 (p. 209) shows the contact which is irregular in detail, but which, over all, dips to the west at about  $15^{\circ}$  to  $20^{\circ}$ .

Along the northeast slope of Nigger Mountain, the trace of the contact suggests a gentle northwesterly dip of about  $5^{\circ}$  to  $10^{\circ}$ . South of Treasure Mountain in the Pauper area, the contact is clearly exposed and although there are local irregularities from the Julia Mine to the Pauper area, the overall attitude of the contact indicates a westerly dip, probably not over  $10^{\circ}$  as suggested on structure section AA' on the geologic map.

Across the southwestern part of Bison Mountain, the contact of the roof of the Bison Mountain granodiorite dips southwesterly between  $10^{\circ}$  and  $15^{\circ}$  as indicated by the position of the contact on the topographic map. At the north end of Bison Mountain, a very gentle northerly dip is indicated, as shown in structure section BB'. Along the northwest contact of the Bison Mountain granodiorite, west of Monarch Creek, the contact is nearly parallel to the contours, and a westerly dip is suggested.

#### Floor of Batholith (?)

At two localities, there are relations which appear to indicate local lensing out of two segments of the batholith into sill-like



bodies. North of Ontario Creek north of Bison Mountain, the Bison Mountain granodiorite is but a thin layer which is underlain and overlain by volcanics. The volcanic rocks below the intrusive are strongly bleached and are the most intensely hornfelsed rocks found in the district. The volcanics above the intrusive are altered to rocks of microaplitic composition. The critical area is north of Ontario Creek in NE  $\frac{1}{4}$  sec. 20, T. 8 N., R. 6 W. Outcrops are poor, but are abundant enough to indicate the relations outlined. The dip of the roof rocks are very gently northerly, the floor dips gently to the southeast. The granodiorite thins out and disappears on the west side of the small gulch and thickens on the east side.

If the volcanic rocks herein considered to represent the floor of the batholithic segment of granodiorite are not the floor but merely a down faulted segment of the roof, three faults must be present, and the differences in metamorphism must be discounted.

The relations at the Pauper area are shown on cross section AA'. A thin lens of quartz monzonite is overlain and underlain by volcanic rocks. A fault complicates the picture, and as outcrops are scarce, the relations are not as clear as might be wished. However, the quartz monzonite lenses out completely on the west side of the fault less than  $\frac{1}{2}$  mi south of the Pauper workings. The sill (?) is probably not over 250 ft thick at the Pauper adit. The volcanic rocks underlying the sill-like body are incipiently hornfelsed, whereas the rocks overlying the quartz monzonite are strongly altered to microaplitic composition. These are the same relative metamorphic relations noted at

the previously mentioned locality north of Bison Mountain. The microplitic rocks extend around the nose of the mountain and suggest the approximate location of the plane of intrusion of the quartz monzonite. In addition, the quartz monzonite body at this locality exhibits the only evidence of vertical compositional differences due to magmatic differentiation found in the district.

### Faults

The only faults of large displacement mapped in the district strike N. 30° to 60° E. as shown on the geologic map. One N. 80° E. fault was mapped in the southwest part of the district. Numerous faults which strike northwesterly are recognized only in underground workings. They have very small displacement, however a conspicuous northwesterly lineament trend in this direction is noted on the topographic map. Immediately west of the district a north-south fault zone was mapped.

#### Northeast Fault System.

Three mapped faults and one inferred fault strike about N. 30° to 60° E. They offset the contact of the batholith as much as a mi or more. The faults appear to be steep because they maintain relatively straight courses across rugged topography.

The northernmost of these faults strikes about N. 60° E. across the northeast part of the district. The dip on the fault plane and the relative amount of displacement are not known except that the northern side is relatively upthrown. Strike slip movement may be significant.



The fault is indicated from several lines of evidence: (1) no granitic rocks are found north of the fault, and andesitic volcanic rocks are not thick on the northwest side; (2) Mississippian limestone, strongly fractured and sheared, is but  $\frac{1}{2}$  mi across glacial cover from Upper Cretaceous volcanic rocks; (3) offset is indicated in the rather sketchy mapping in the poorly exposed western part of the district.

In the Little Blackfoot Creek segment northwest of the Hopkins Mine, the valley appears to be structurally controlled by a fault or fracture zone which strikes about N.  $40^{\circ}$  to  $45^{\circ}$  E. The north side is probably the upthrown side because Colorado shale is brought into contact with the younger volcanic rocks on the south side of the creek.

Another fault with the same general trend, about N.  $40^{\circ}$  to  $45^{\circ}$  E. is found between Nigger Mountain and Treasure Mountain. The fault is probably steep because its course is straight across about 1000 feet of relief. The erosion along the southwestern segment of Little Blackfoot Creek appears to be controlled by this fault. It is probably a fault zone, suggested by the sliver of volcanic rock found along the fault. The contact of the batholith is offset possibly one half mile or more.

A similarly trending fault strikes N.  $30^{\circ}$  E. southeast of Treasure Mountain. The magnitude of the faulting is not clear, however it need not be great even if the contact of the batholith is offset over one mile, because the dip on the contact of the batholith is thought to be a small angle to the west. Northeast and southwest of the fault as mapped, it could not be traced into quartz monzonite or volcanic rocks respectively. Treasure Mountain is clearly on the upthrown block between this and the next fault to the northwest.

Just east of the Continental Divide southeast of Jericho Mountain, Ruppel (personal communication) has mapped a N.  $70^{\circ}$  E. fault. It is not clearly indicated in the Elliston district, however the contact between the altered volcanics and the quartz monzonite at this locality may be fault controlled.

#### East-West Fault.

Faulting in a N.  $80^{\circ}$  E. direction is noted along one trend line in the southern part of the district, north of Bison Mountain and on the east side of Larabee Gulch. One segment offsets Bison Mountain granodiorite. A third segment of possibly the same fault was mapped by Ruppel just west of the boundary of the district. (Personal communication.) If the three segments which line up are the same fault, it is over 4 mi long without clear indication of large displacement. It is probably pre-rhyolite in age because the rhyolite is not clearly offset. The fault has about the same trend as the vein zone, and because of its length and absence of strong offset, it is suggested that it may be related to the nearly east-west vein system.

#### Northwest Faults.

Faults which strike in a northwesterly direction are too small to map at the scale of the geologic map, but their presence is strongly suggested by the lineament pattern of the drainage, as noted on the topographic map. The drainage pattern of Telegraph Creek, one segment of Little Blackfoot Creek, and Ontario Creek, as well as many small



gulches and gaps across the ridges, particularly southwest of Jericho Mountain, suggest fracture control in a northwesterly direction.

In many of the mines, small left lateral faults of small displacement, measured usually in a few in to a few ft, have this same trend. These faults clearly cut the east-west vein structures which appear to be cut by the northeast faults as well.

#### North-South Faults.

North-south trending faults are probably the youngest faults in the region. None is mapped in the district, however about 2 mi west of the district, and extending across a zone  $\frac{1}{2}$  mi wide, there is a strong north-south fault zone. The faulting is clearly pre-rhyolite in that area, however north of Hat Creek and not  $\frac{1}{4}$  mi west of the western limit of the district, rhyolite appears to have been offset by north-south faults. Within the district, physiographic evidence suggests that the southern course of Mike Renig Gulch may be fault controlled.

Block faulting along a north-south trend is conspicuous throughout western Montana, and is thought by Pardee (1951) to be responsible for the basin-and-range type structures. The conspicuous East Ridge at Butte is considered to be a typical example of this type of faulting.

#### Joints

The jointing pattern in the pre-batholithic volcanic rocks and in the intrusive rocks are more or less the same. The most pronounced set strikes N.  $65^{\circ}$  E., and generally dip steeply south. It is parallel to the main vein system in the batholith. Two less-conspicuous sets

strike N. 5° E., and N. 50° W. A fourth set strikes nearly east-west. All are high-angle to vertical.

The joints in the volcanic rocks are commonly closely spaced, indistinct, and not consistent over a single outcrop, whereas in the intrusive rocks the joints are much more consistent and more widely spaced. Rarely, however, are joints spaced more than a few feet apart except in the coarsest-grained rocks where spacings may be as much as 5 ft or more.

### Veins

The structural pattern of the veins is touched on in this section for the purpose of indicating the probable correlations of the veins with the fracture pattern in the district. The complex history of veins formation is described in the section on ore deposits.

#### Veins in Quartz Monzonite.

Veins in the batholithic rocks are confined principally to N. 65° + 20° E. steeply south dipping fracture zones. The larger veins range from  $\frac{1}{4}$  to over two mi in length, and are spaced at roughly  $\frac{1}{2}$  mi intervals in the southeastern part of the district. Some of the veins extend from Elliston into the Rimini district. The individual veins are complex lodes which have been fractured and re-fractured many times. The amount of offset, however, is considered to be negligible.

Although many of the veins are wide mineralized structures in the granitic rocks, yet where they cross into the volcanic rocks, they



become narrow stringers. At the Julia Mine (cen. N. line sec. 8, T. 7 N., R. 6 W.) the Julia vein is about 5 ft wide in quartz monzonite, but in the volcanic rocks, it becomes a series of small quartz stringers, the largest of which are only an inch or so in width.

#### Veins in Volcanic Rocks.

On Nigger Mountain, the Big Dick vein, and several lesser veins, have the same general trend as veins in quartz monzonite. The Big Dick-Black Jack-Little Dick vein (assuming it is one vein) is nearly vertical, and may be traced for about  $1\frac{1}{2}$  mi. It contains virtually the same minerals that are found in the same trending veins in quartz monzonite.

At the Hopkins Mine, the veins which occur on the east side of Little Blackfoot Creek, are apparently related to a northeast fault zone which is presumed to control the valley development at this locality. The veins are steep and appear to become narrower where the dip becomes northwesterly. The sparsely mineralized fracture zone at the Sunset Mine is apparently along the same type of fracture zone. The breccia zone at the Clark Mine has the same trend.

On Nigger Mountain in the Big Dick Mine, there is a "flat" vein which, although thin, contained some of the richest ores produced in the district. The veins occur within the volcanics, possibly at the contact between two units, or along flat shears. The latter appears more probable because there is an indication that the thickness of the veins is greater in the flatter parts, suggesting reverse movement.

The northwest striking faults of small displacement are, in some instances, weakly mineralized with thin quartz stringers. The wall rocks may be bleached, rarely as much as an inch from the stringers.

## MINERAL DEPOSITS

Gold and silver occur with sulfide minerals in complex veins in quartz monzonite and in fissure veins in volcanic rocks. The mineralogy and paragenesis of the ore minerals is described in some detail. The complex vein formation histories of veins in quartz monzonite are also described from typical examples. Although wall rock alteration is significant, only a small amount of work has been done because of the lack of material for study. However, notes on the genesis of the wall rock alteration are included.

## Mineralogy

The mineralogy of the veins in the Elliston district is similar to the mineralogy of veins found in the dozen or more mining districts in the granitic and volcanic rocks in the region of the northern part of the Boulder batholith. The minerals found in the veins are listed in Table 18. Each of the minerals is briefly described and their associations with other minerals discussed.

## Non-Metallic Gangue Minerals.

Tourmaline. Tourmaline is abundant in many of the veins. Its association with base metal deposits is quite uncommon and was first described by Knopf (1913). Actually in the northern part of the Boulder batholith, tourmaline is a characteristic gangue mineral, found with quartz-pyrite and other sulfides.

Tourmaline occurs in masses of radiating and interlocking needles without significant amounts of quartz, replacing the wall rocks in some



of the earliest fractures and in at least three different associations with quartz: (1) in intimate fine-grained intergrowths so fine that the minerals are not resolved with the unaided eye, in which the quartz and tourmaline are in about equal amounts; (2) tourmaline intergrown between the crystal grains of quartz; and (3) in needles of tourmaline penetrating crystals of quartz, the latter type is shown in Figure 82.

Tourmaline is always black in hand specimen. In thin section it is strongly pleochroic, o = bluish green, e = olive green.

Tourmaline is intimately intergrown with and replaced by pyrite at a number of localities, similar to the tourmaline-pyrite intergrowths found in the late deuteritic vein-like bodies, but it has not been observed intergrown with any of the base metal sulfides. Knopf intimates that the tourmaline is related to the base metals, however the reference is not specific with regard to the individual minerals. He states: "That the development of the sulphides and the tourmalinization were contemporaneous is proved by the mutual intergrowth of the sulphides with the tourmaline." (Knopf 1913, p. 113)

Quartz. Quartz is commonly the most abundant gangue mineral in the veins. It is clearly introduced into the veins at several different stages, often with different physical characters, and with different suites of vein minerals. It occurs with tourmaline in intimate intergrowths so fine that the minerals are not detected with the naked eye; in massive "vein" quartz which may be white, translucent, or strongly milky; and in minor amounts in combs of crystals lining small open spaces either along fractures or in open spaces between breccia fragments.

Table 18

## VEIN MINERALS

## Non-Metallic Gangue Minerals

quartz	ankeritic carbonates
tourmaline	chalcedonic quartz
calcite	

## Hypogene Minerals

pyrite	tennantite-tetrahedrite
arsenopyrite	tennantite
sphalerite	chalcopyrite
galena	bornite
bournonite	millerite
boulangerite	hematite
tetrahedrite	magnetic iron oxide

## Supergene and Secondary Minerals

covellite  
 chalcocite  
 cerussite  
 blumbojaresite  
 limonite





Figure 82. Photomicrograph of quartz with tourmaline needles in tourmalinized quartz monzonite wall rock at the Lilly mine. x 200 (Regnier 1950)

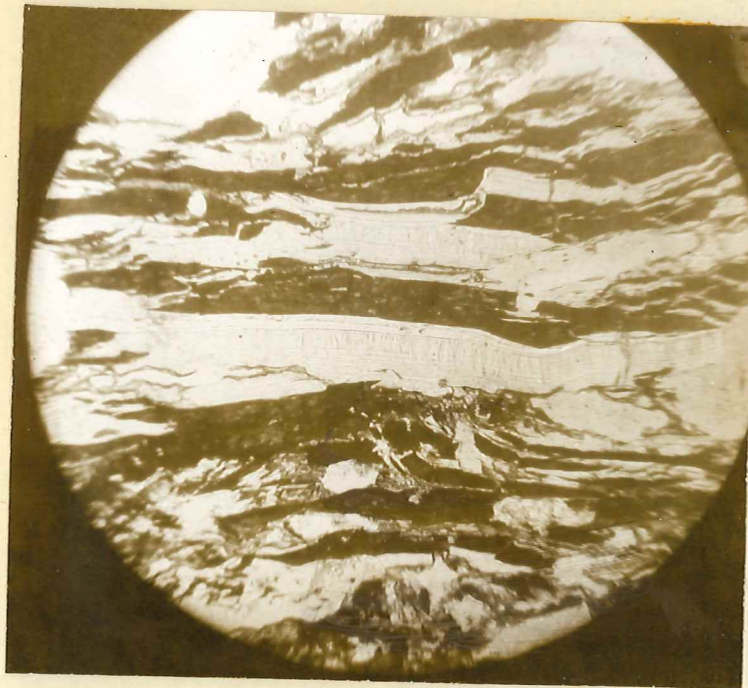


Figure 83. Photomicrograph of chalcedonic quartz veinlets in sphalerite. Plane polarized light. x 20 (Regnier 1950)

Early hydrothermal quartz and tourmaline replace the wall rocks of some of the veins, and also is found in fissure fillings. The larger veins are made up of a succession of fractures in which quartz was introduced at several different stages. Milky white "bull" quartz is commonly found in some of the large veins with pyrite, the only other vein mineral. Small amounts of quartz accompanied the introduction of lead and zinc minerals, however the amount of quartz occurring with sphalerite and galena is small.

Calcite. Calcite is very minor in amount in some of the quartz veins. A small amount occurs at the Lilly Mine and undoubtedly is found elsewhere. The mineral is white and fills open spaces between quartz, sphalerite and galena.

Ankeritic carbonates. Ankeritic carbonate is not abundant in the youngest veins in the quartz monzonite, however it is an important gangue mineral in veins in the volcanic rocks. The colors range from light buff to light brown and have some pinkish hues which suggest the presence of manganese. Commonly ankeritic carbonate veinlets are somewhat younger than the base metal-bearing veins since they transect the early sulfide bearing veins.

Chalcedonic quartz. Chalcedonic quartz is rarely observed in the veins in the Elliston district. It is always the youngest primary mineral and found in very late fractures. At the Lilly Mine, some sheared sphalerite contains chalcedonic quartz in the fractures as shown in Figure 83.



Arsenopyrite. Arsenopyrite is the common gold-bearing mineral in the veins. Bladed or blunt needle-like crystals are found in quartz, but arsenopyrite is very irregular in outline in contact with tourmaline. Arsenopyrite and pyrite commonly occur together and the two minerals appear to have formed more or less simultaneously where pyrite and arsenopyrite have good crystal form against each other as in Figure 84. Figure 85 shows arsenopyrite replacing pyrite.

A spectrographic analysis of hand-picked arsenopyrite shows iron, arsenic, silver, lead, cobalt, nickel, bismuth and antimony. (Regnier 1951, p. 12). A fire assay of selected arsenopyrite from the Lilly Mine shows .53 oz gold and 5.1 oz silver per ton. (Aiken 1950.)

Gold is associated with arsenopyrite throughout the district, and in the region generally. It is thought to occupy open spaces in the arsenopyrite structure. (Stillwell and Edwards 1946.)

Pyrite. Pyrite is the most common sulfide mineral in the veins. It is found in barren quartz-tourmaline veins, with quartz in otherwise unmineralized veins and with the base metals. Most of the pyrite appears to be isotropic, however at several localities in the southern part of the district, the pyrite is lighter in color than ordinary pyrite and optically anisotropic. An unusual anisotropic pyrite was noted at the Orphan Boy claim. Flat, pseudo-orthorhombic crystals consist of combinations of cubes and pyritohedrons. Regnier (1951) determined by spectrographic analysis that the mineral contains iron, arsenic, nickel, cobalt, lead and silver. As arsenic is relatively insensitive, spectrographically, a relatively large amount is indicated.

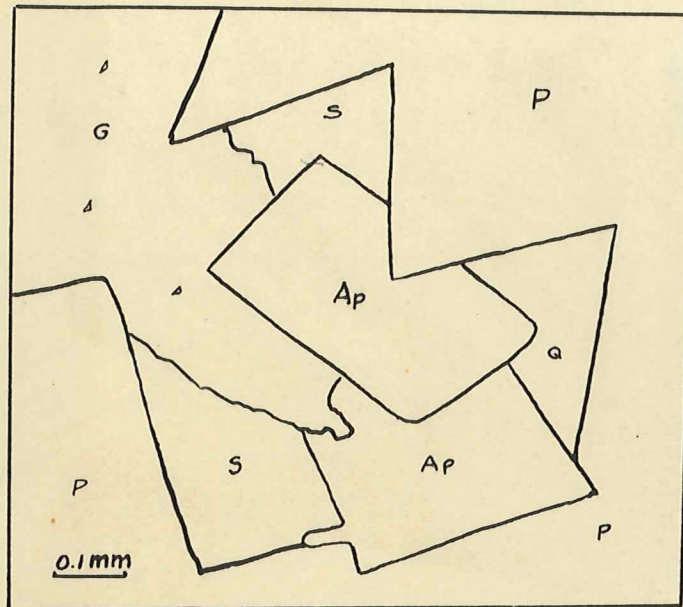


Figure 84. Camera lucida sketch of pyrite and arsenopyrite in contact along possible crystal faces. Partial replacement of arsenopyrite by galena indicated. Q is late quartz. x 100 (after Regnier 1950)



Figure 85. Photomicrograph of pyrite partly replaced by arsenopyrite. Pyrite is pitted, has high relief; arsenopyrite has low relief. x 100 (Regnier 1950)



The arsenic may be responsible for the anisotropism, as suggested by Schneiderholm (1937).

Fire assay of hand-picked pyrite from the Third Term and the Lilly Mines showed a trace of gold, .06 oz and 4.6 oz silver per ton, respectively.

Sphalerite. Sphalerite is most commonly dark brown to nearly black, however some sphalerite is light resinous brown. The mineral most commonly fills open spaces, but it also occurs as replacement of quartz, pyrite and arsenopyrite. Figures 86, 87 and 88 show sphalerite replacing quartz, pyrite and arsenopyrite respectively. Sphalerite is the most easily replaced mineral of all the ore minerals. Chalcopyrite is usually found in dark sphalerite in exsolution blebs, however, some light sphalerite is completely devoid of chalcopyrite. In some instances, late chalcopyrite replaces sphalerite. A spectographic analysis by Regnier (1951) indicates the presence of iron, zinc, copper, silver, cadmium, indium, gallium, lead and arsenic. These are typical metals found in sphalerite (Stoiber 1940 ). At the Lilly Mine, selected sphalerite was assayed and found to contain .12 oz gold and 18.2 oz silver per ton. It is probable that the sphalerite contains some tetrahedrite which is responsible for the large amount of silver. (Aiken 1950). Two types of sphalerite were assayed from the Third Term claim, the black sphalerite contains a trace of gold, and 5.32 oz silver, and the resin-brown sphalerite a trace of gold and 1.04 oz silver (Johns , 1952).



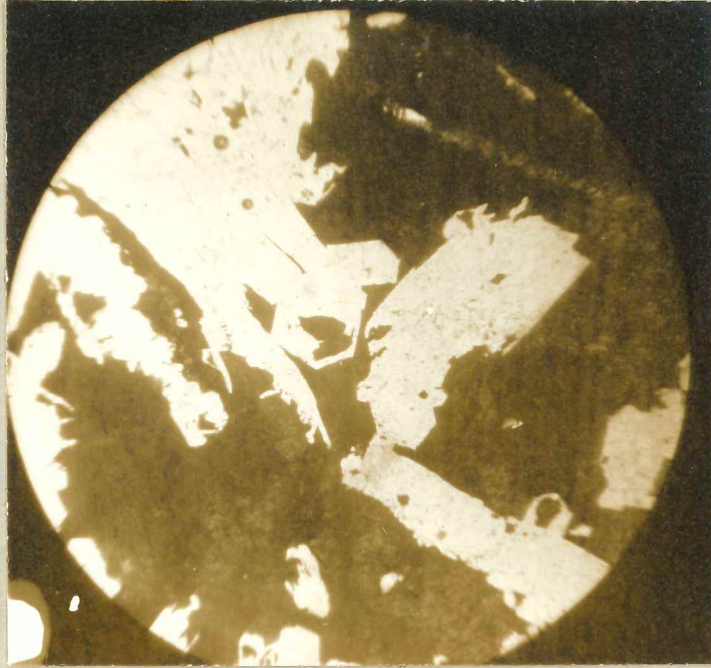


Figure 86. Photomicrograph showing replacement of quartz (black) by sphalerite (white) x 100 Reflected light. (Regnier 1950)

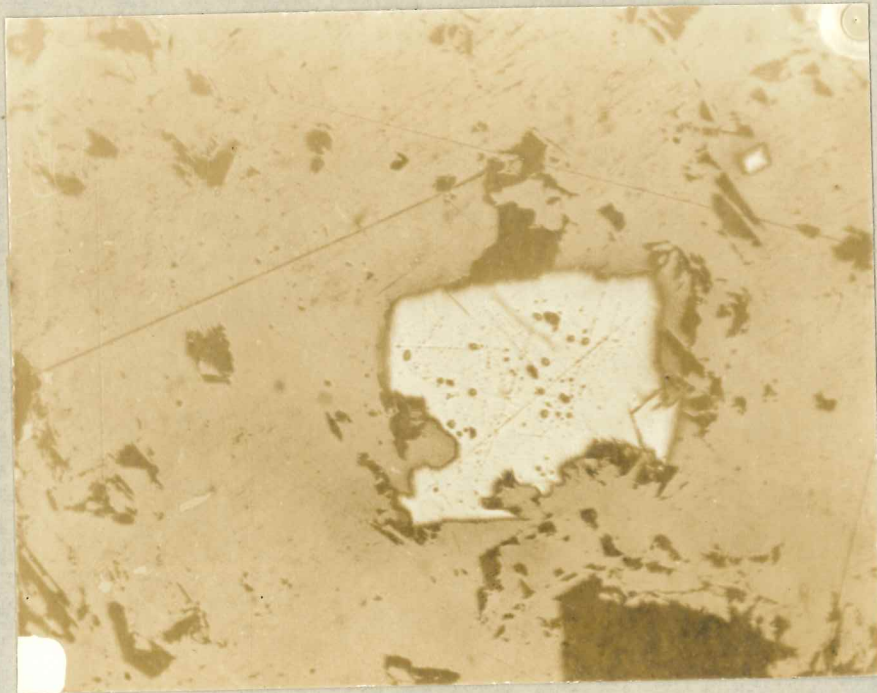


Figure 87. Photomicrograph of pyrite (white) partly replaced by sphalerite (gray). x 100 Reflected light. (Regnier 1950)



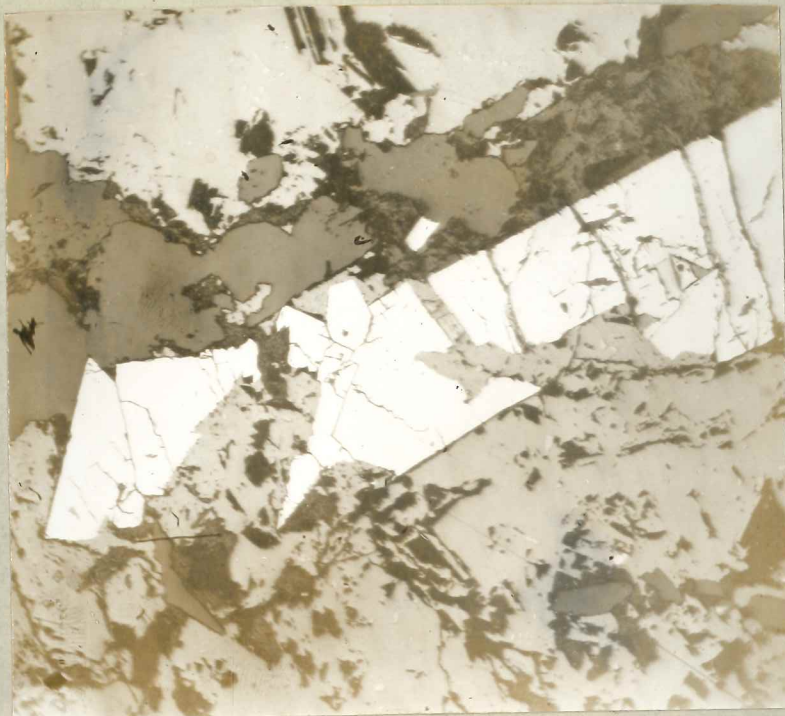


Figure 88. Photomicrograph of arsenopyrite (white) partly replaced by sphalerite (light gray). Quartz is dark gray. x 100. Reflected light. (Regnier 1950)



Figure 89. Photomicrograph of sphalerite (gray) partly replaced by bournonite (lighter gray) through center of photograph. Light gray (diamond shaped grains) arsenopyrite. x 100. Reflected light. (Regnier 1950)



Tetrahedrite-Tennantite. Tetrahedrite is the most important silver mineral. It is a minor but common mineral, almost always found with sphalerite, and replacing it. Tennantite was identified at the Third Term claim and Sure Thing Mine, and an intermediate tetrahedrite-tennantite mineral at the Third Term claim. A fire assay of hand-picked tetrahedrite from the Third Term prospect shows .16 oz of gold and 6.60 oz silver. (Johns 1952.) Tetrahedrite is probably responsible for the 18.2 oz silver per ton in the sphalerite of the Lilly Mine.

Hand-picked samples of tetrahedrite, tetrahedrite-tennantite, and tennantite were analyzed by powder X-ray methods by Johns (1952). The data are presented in Table 19. At the Lilly Mine, the tetrahedrite was checked for arsenic, but none was found (Regnier 1951).

Bournonite. Bournonite was identified by Regnier from polished sections of ores from the Lilly Mine (1951) and checked by me. It occurs as irregular replacement grains in sphalerite as shown in Figure 89. In contact with galena, the mineral has a bluish white color. The mineral is weakly anisotropic, pinkish brown to brownish gray. Multiple twinning, similar in appearance to that found in plagioclase, is a characteristic and distinctive feature.

Boulangerite. Boulangerite is not a common mineral but has been positively identified at the Lilly Mine and the Moonlight claim. It is moderately abundant at the latter locality associated with sphalerite, as shown in Figure 90. The mineral is noted in polished sections replacing sphalerite and often asserting its own crystal form in the sphalerite. The mineral occurs in elongated bladed needles and is



Table 19

X-ray "d" values for Tetrahedrite  
and Tennantite, Third Term Mine

Tennantite (standard)	Tennantite- Tetrahedrite Third Term	Tennantite- Tetrahedrite Third Term	Tetrahedrite Third Term	Tetrahedrite (standard)
4.16				4.2
3.70		3.71	3.70	3.68
		3.29	3.30	3.29
		3.14		
2.95*	2.95	2.98*	3.00*	3.00
2.74				2.78
2.56	2.57	2.57	2.59	2.60
		2.46	2.44	2.45
				2.33
2.09		2.09	2.12	2.12
2.10		2.02	2.04	2.04
1.87		1.92		1.89
1.81		1.82	1.83	1.83
1.76	1.79		1.78	1.78
1.71				1.73
		1.64	1.68	1.68
1.65				1.64
				1.605
	1.52	1.55	1.57	1.56
1.54				



Figure 90. Photomicrograph of sphalerite (gray) partly replaced by boulangierite (white). Boulangierite tends to develop hypautomorphic to automorphic crystals in sphalerite. x 200. Reflected light. (Regnier 1950)



Figure 91. Photomicrograph of sphalerite (light gray) partly replaced by galena (white). Both minerals replace quartz (dark gray). x 100. Reflected light. (Regnier 1950)



strongly anisotropic. It resembles the boulangerite found in a similar mineral association in the Emery district. Careful microchemical tests failed to detect iron, nitric acid stains black, and potassium hydroxide is negative--all tests which serve to distinguish the mineral from Jamesonite.

Galena. Galena is an important mineral found mostly in fissure fillings, but it also replaces pyrite and sphalerite. Figure 91 shows replacement criteria for replacement of sphalerite by galena. Most commonly, the galena is quite fine-grained and in some cases may be described as steel galena, however, at other localities, the galena is coarse grained and shows good cleavage. Spectrographic analysis of the fine-grained galena from the Lilly Mine shows lead, copper, silver and iron. (Regnier 1951). Fire assays at the Lilly and Third Term Mines show .08 and .25 oz gold, and 2.1 and 9.72 oz silver per ton respectively. (Aiken 1950, Johns 1952.)

Chalcopyrite. Chalcopyrite is a relatively minor mineral found in two distinctly different occurrences. It is present in practically all dark sphalerite in exsolution blebs which are clearly controlled by interval structures in sphalerite. It also occurs as a distinctly younger mineral than sphalerite or galena, and it is found to replace both of these minerals, however the replacement of sphalerite is usually more conspicuous. Figure 92 shows chalcopyrite exsolution in sphalerite.



Figure 92. Photomicrograph of exsolution blebs of chalcopyrite (light gray) in sphalerite (gray). Note concentration of blebs along trace of dodecahedral cleavage planes. x 200. Reflected light. (Regnier 1950)

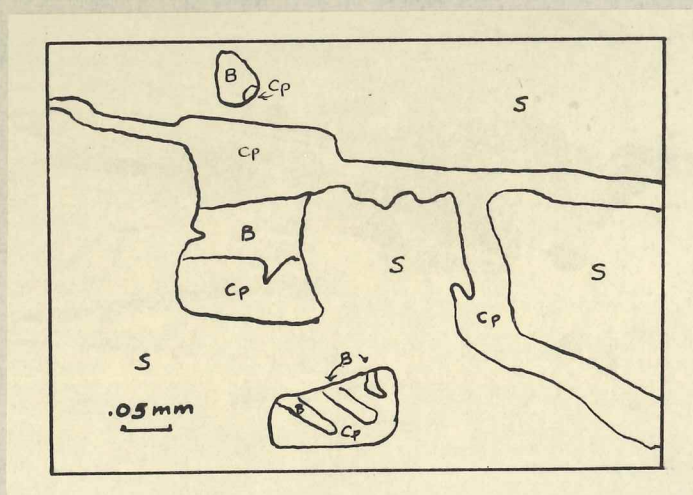


Figure 93. Camera lucida drawing of chalcopyrite exsolution strings and blebs partly replaced by bornite. Host mineral is sphalerite. Sketched from polished section of ore from Beatrice mine. About x 500. After Regnier 1950)



Bornite. Bornite is found in very small amounts at the Bullion Parks area on the Continental Divide. It was observed to have replaced exsolution blebs of chalcopyrite in sphalerite as shown in Figure 93.

Millerite. The presence of millerite was first noted by the writer at the Lilly Mine. It occurs in quartz-lined vuggy openings in the vein. The hair-like crystals are characteristic. Nickel was detected by microchemical tests.

Hematite. Hematite is not common, but at one prospect pit northeast of Telegraph Mine, quartz and hematite are found without any sulfide minerals. Massive chunks of hematite 20 to 30 pounds are found on the dump. The hematite is partly oxidized to goethite.

Magnetitic Iron Ore. In the massive hematite at the prospect northeast of Telegraph Mine (see hematite above) some of the hematite is slightly magnetic. Magnetite, or possibly martite may be present with the hematite. None of the specimens was examined in detail in polished section.

Chalcocite. Supergene chalcocite was noted only at the Beatrice Mine in the Bullion Parks area. (The mine is a few hundred yards east of the Divide.) Chalcocite is observed as a replacement mineral in sphalerite, chalcopyrite, and pyrite.

Covellite. Covellite is a rare supergene mineral noted in polished ore samples from the Beatrice Mine on the east side of the Divide. The deep blue color and strong anisotropism serve to identify the minute flecks found with supergene chalcocite in pyrite, chalcopyrite and sphalerite.

Oxidation minerals. The secondary minerals have not been studied in detail. Cerrussite, plumbojarosite, limonite and goethite have been identified. Cerrussite and plumbojarosite were noted in the oxidized part of the workings at the Hopkins Mine. Limonite is universally associated with fine-grained silica in the oxidized gossans. Distinctive boxworks are commonly preserved. The sphalerite boxworks are open and very siliceous, whereas the pyrite boxworks contain massive limonite.

#### Paragenesis

#### Stages of Vein Formation.

The succession of vein formation is about as follows:

Tourmaline-rich veins, minor pyrite and quartz (barren)

Tourmaline-quartz-pyrite veins (barren)

Tourmaline-quartz-pyrite-arsenopyrite (gold bearing)

Quartz-pyrite veins (barren)

Quartz veins with:

a. pyrite, arsenopyrite

b. sphalerite

c. sphalerite-galena, silver-bearing minerals

d. chalcopyrite

Quartz-carbonate veins

Sphalerite and galena in volcanic rocks

Chalcedonic quartz veins.



### Criteria for Paragenetic Relations.

The age relations of two minerals found in contact are not always clearly indicated. Some minerals exert a strong power of crystallization and form, by replacement, good crystal outlines. Thus the presence of an automorphic form is not necessarily a sign of early crystallization. Pyrite, arsenopyrite and boulangerite tend to form good crystal form under favorable circumstances. Other minerals exert only a slight tendency to develop crystal form. The replacement textures between two minerals which show only a slight tendency to develop crystal form are distinctly clearer than for the minerals which exert their own crystal form.

Criteria for replacement are many and not always dependable. One indication between a pair of minerals, such as galena and sphalerite, is insufficient. Usually several criteria were used to determine the relations before they were introduced as evidence on a paragenetic diagram prepared for a given mine or prospect. The sum of all of the sections was compiled into separate diagrams and these brought together to form the diagram in Figure 94. The criteria described by Newhouse (1928), Bastian (1931), Edwards (1947) and others have been used as a basis for distinguishing age relations.

Paragenesis diagram. Line diagrams are commonly employed to show paragenetic relations, as in Figure 94. They may be satisfactory to show the general order or sequence of mineral crystallization, but they fail to indicate significant details in mineral inter-relationships, and likewise tend to imply, by over-simplification, relations not known to exist. Obscure and indefinite relations appear to be as distinct and positive as all other mineral relationships.

In the diagram used, lines connect each pair of minerals found in contact. The precise relationships are indicated. In the case of replacement textures, an arrow head points toward the mineral which has been partially replaced. If the indications are inconclusive, double pointed arrows may be employed, and if one tendency is dominant over another, the size of the arrowhead may serve to indicate the relative degree of certainty or preponderance of observations. Simultaneous deposition, and exsolution are likewise indicated. (Robertson and Vandever 1952.)

The preparation of this type of diagram requires that a very thorough examination be made to determine all of the possible inter-relationships. From such a study, it becomes apparent that sphalerite is very easily replaced, whereas pyrite, for example, is a much less hospitable host to the younger minerals.

Pyrite-Arsenopyrite. Pyrite replaces quartz much more easily than tourmaline, as shown in Figure 95. Arsenopyrite may form good crystal forms in quartz, but is commonly ragged in outline against tourmaline as shown in Figure 96. Arsenopyrite may replace pyrite, and pyrite may replace arsenopyrite. In some instances, the minerals appear to crystallize more or less simultaneously. Figure 85 indicates criteria for nearly simultaneous deposition, whereas Figure 83 shows arsenopyrite replacing pyrite. Both arsenopyrite and pyrite are replaced by sphalerite, but only rarely are other minerals found to replace pyrite. Galena, chalcopyrite, and bornite replace pyrite to some extent, and bornite replaces arsenopyrite. (Bornite is so uncommon that its precise position



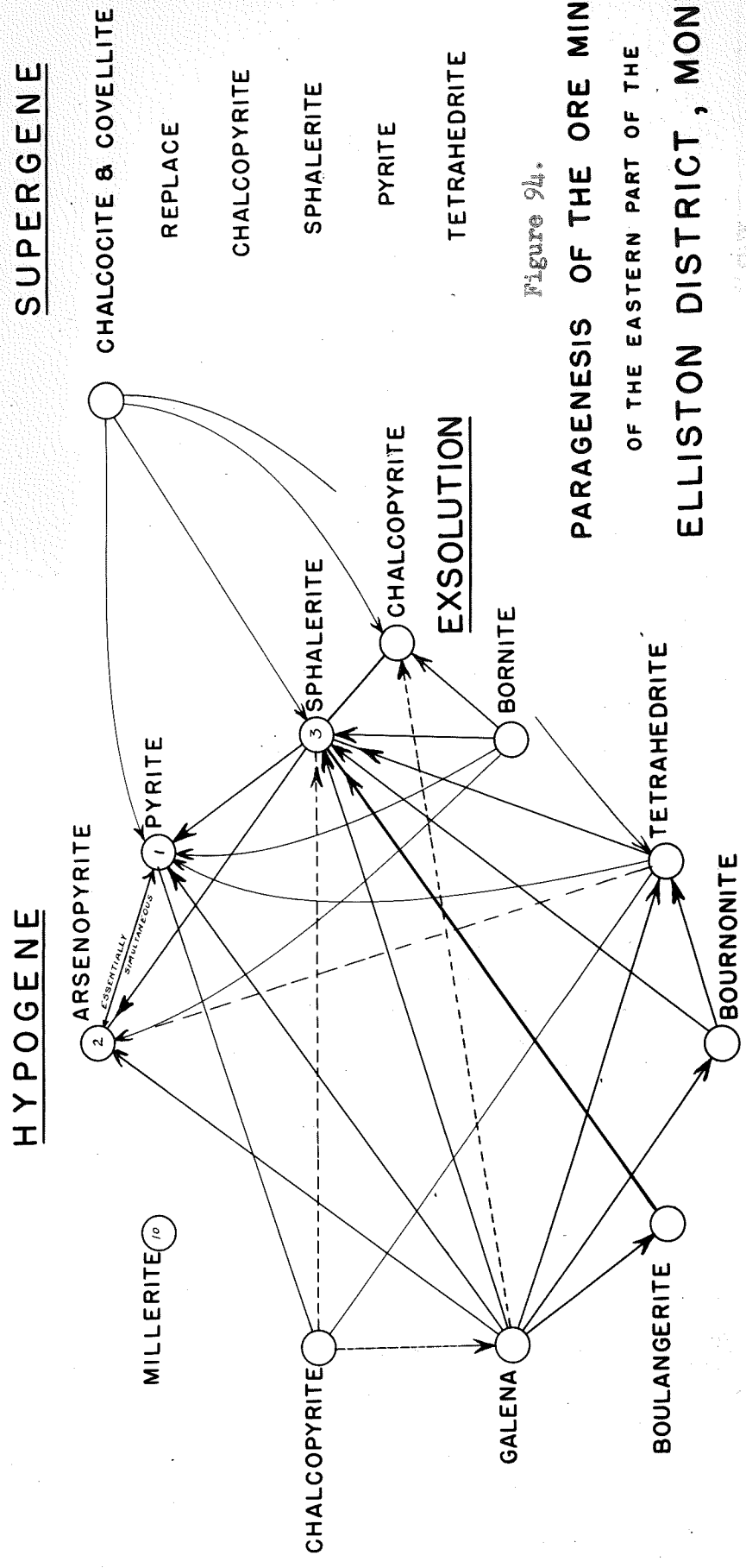


Figure 94.

PARAGENESIS OF THE ORE MINERALS  
 OF THE EASTERN PART OF THE  
 ELLISTON DISTRICT, MONTANA

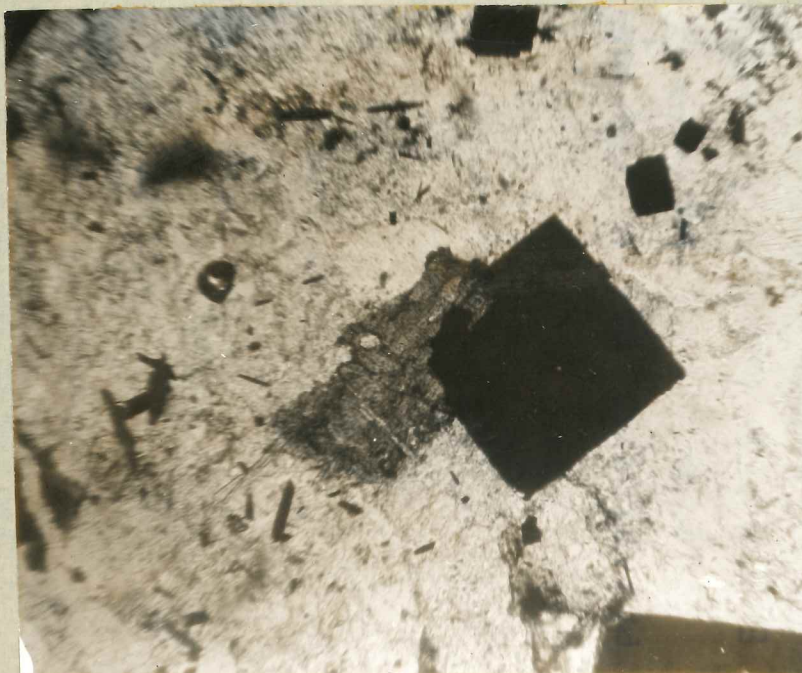


Figure 95. Photomicrograph of pyrite cube (black) formed by replacement in altered quartz monzonite. Plane polarized light. x 40. (Regnier 1950)

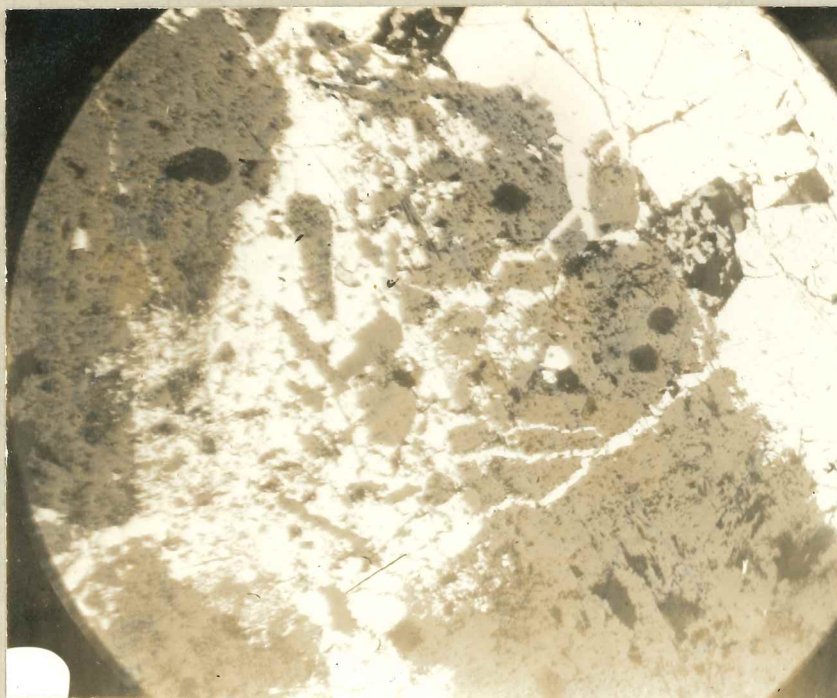


Figure 96. Photomicrograph of tourmaline (gray) partly replaced by pyrite (white). Note ragged boundaries of pyrite against tourmaline compared to automorphic pyrite replacement in sericitized wall rocks as shown in Figure 95 above. x 100. Reflected light. (Regnier 1950)



in the sequence of minerals has not been established.) Fractured arsenopyrite may be cemented with late quartz as shown in Figure 97.

Sphalerite. Sphalerite is the common host for all younger sulfide minerals. Bornite, tetrahedrite, tennantite, bournonite, boulangerite, galena, and chalcopyrite replace sphalerite. Boulangerite replaces only sphalerite, with but one possible exception that it appears to have replaced some bournonite. Chalcopyrite is found in much of the dark sphalerite in exsolution blebs which are clearly crystallographically controlled, as shown in Figure 92. This may indicate that the sphalerite formed at a temperature greater than  $350^{\circ}$  to  $400^{\circ}$  C. (Buerger 1934, p. 528). Some sphalerite, notably at the Third Term prospect, contains no exsolution chalcopyrite.

Galena. Galena is distinctly younger than the sulfo-salts and complex lead and copper sulf antimonides. Galena replaces boulangerite, bournonite, and tetrahedrite.

Chalcopyrite. Chalcopyrite is distinctly a late mineral in the paragenetic sequence. This is not its common position in the sequence of minerals observed by Newhouse (1938) and Lindgren (1937) and Edwards (1947). Chalcopyrite in the position observed more nearly agrees with the theoretical position of the iron-bearing sulfides, according to Verhoogen (1938, p. 776), and with the decreasing hardness scale of Gilbert (1924, p. 668) in which early chalcopyrite is an anomaly. The copper-bearing veins have been described as younger than the lead-zinc veins in the section on the hydrothermal stage of crystallization discussed in quartz monzonite. In the Elliston district,

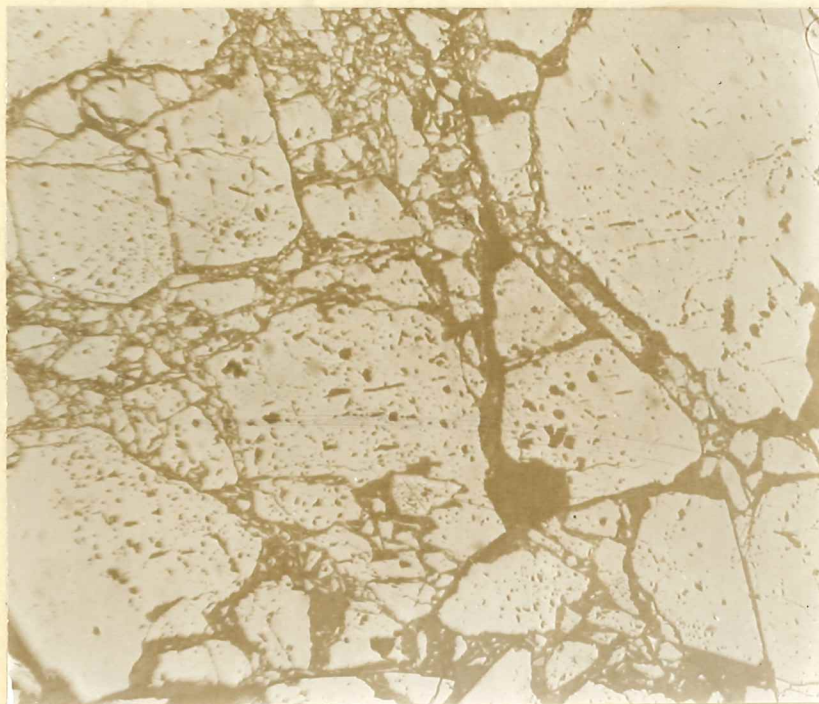


Figure 97. Photomicrograph of brecciated arsenopyrite cemented by younger quartz. x 100. Reflected light. (Regnier 1950)



the evidence that chalcopyrite-bearing veins are younger than lead-zinc veins is lacking. Pinkney (personal communication) is convinced that the copper-bearing structures are generally younger than the lead-zinc structures. Certainly in the Elliston district, chalcopyrite occurs late in the paragenetic sequence of the common hypogene sulfide minerals.

Evidence of its position late in the paragenetic sequence is shown in Figure 98 which is a line diagram showing replacement of galena by chalcopyrite. Figure 99 shows chalcopyrite filling fractures in tetrahedrite, which is, in turn found as a replacement mineral in sphalerite.

### Veins

The principal mineral deposits are found in veins in quartz monzonite and volcanic rocks. Steeply dipping, complex replacement lodes and fissure veins, which contain the valuable mineral deposits, strike about N.  $65^{\circ}$  + E. in quartz monzonite. In the volcanic rocks, there are three kinds of veins which have been productive: (1) Fissure veins which have the same trend as the veins in the quartz monzonite are mostly narrow and of limited importance; (2) Nearly vertical fissure veins which strike N.  $45^{\circ}$  E. and which are parallel to the main fault pattern of the district; and (3) "Flat" veins which are thin, less than 1 ft wide, which have yielded some of the highest tenor ores mined in the district. A few steep east-west striking fissure veins occur in sedimentary rocks.

Although there are probably several miles of underground workings in the district, only a few thousand feet were open at various times during the period of this investigation. The largest mines are inactive,

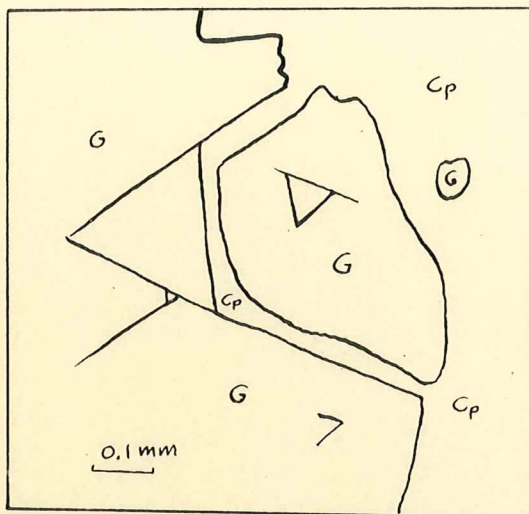


Figure 98. Chalcopyrite replacing galena. Cp chalcopyrite, G galena. (After Regnier 1951.)

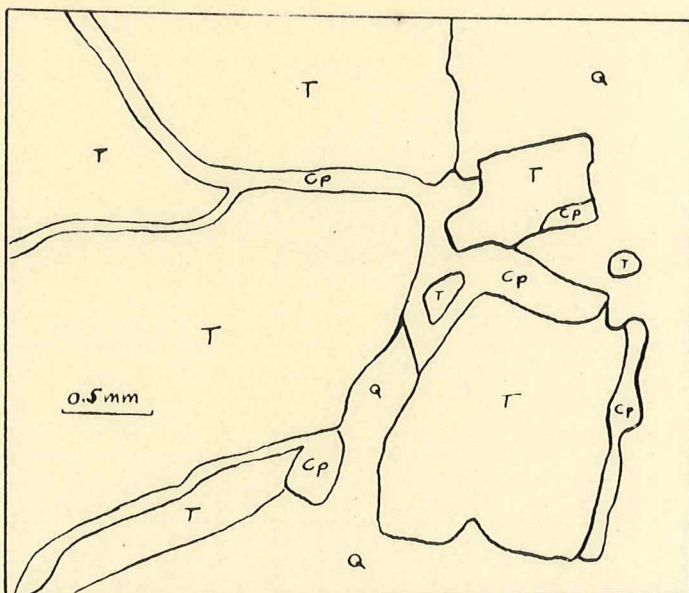


Figure 99. Chalcopyrite veinlets in and replacement of tetrahedrite. Lilly mine. T tetrahedrite, Cp chalcopyrite, Q quartz. (After Regnier 1951)



and the underground workings are inaccessible. In other mines, parts of the main haulage ways are open, but the stopes are caved. In others the underground is completely inaccessible. Because of the limited amount of information available, the structures which localize the ore shoots are very imperfectly known. The relations which have been observed in some of the accessible mines, and which are indicated on old mine maps and reports, give indications of the controls of the ore shoots in the veins.

#### Veins in Quartz Monzonite.

All of the mines which are known to have produced valuable ores are located close to the margin (and presumably a short distance below the roof) of the batholith. The veins are shown on the geologic map. Several of the veins or vein-zones are more than 1 mi long and some are as much as 20 ft wide. The veins are also spaced at about  $\frac{1}{2}$  mi intervals throughout the southeastern part of the district; most of the large veins occurring in the southern end of the mineralized area as shown on the geologic map.

At the Moonlight, Sure Thing and Anna R., the wall rock along a part of each of the veins is aplite. At the Anna R., the rock is a rather coarse aplite, whereas at the other two properties the aplite is very fine grained.

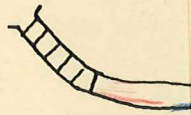
All the veins have the same general structural and mineralogical features, however there are two types distinguished by the presence or absence of tourmaline. In some large veins, tourmaline is abundant, whereas in other equally large veins, tourmaline is absent. No

systematic differences in the ore mineralogy or wall-rock alteration is found in the two types of veins. Rather, it is thought that the veins which do not contain tourmaline formed at a slightly younger stage than the tourmaline-bearing structures, but prior to the introduction of the base metals. The general sequence of hydrothermal stages has been discussed in the section on the hydrothermal crystallization stage.

Quartz-tourmaline veins. Of the tourmaline-bearing veins, only the Lilly Mine was opened during a part of the time of this study. It is located in the southeastern part of the district in NE  $\frac{1}{4}$  SW  $\frac{1}{4}$  sec. 15, T. 8 N., R. 6 W. The tunnel level and two lower levels were mapped for the main vein structures, as shown in Figure 100. (Note: The history and productive record of this and other mines are given in the appendix.) The vein is something of a maverick because it strikes N.  $45^{\circ}$  E., and dips  $80^{\circ}$  N., nearly  $20^{\circ}$  off the main course, and dips north instead of south. The mapping was not in sufficient detail to show the structural relationships between each of the successive stages of vein formation. However, sufficient observations were made to indicate eight stages. They are:

1. Quartz-tourmaline replacement in walls of fracture in quartz monzonite. Sericitization at this stage. A small amount of pyrite is disseminated at the wall rocks. No gold or valuable minerals introduced at this stage.
2. Fine-grained black quartz-tourmaline vein matter with pyrite in large part introduced as fissure filling. There were possibly





ADIT LEVEL 80

MINE DISTRICT

50 FEET

several periods of fracturing and vein filling at this stage. It is barren of valuable minerals.

3. Quartz-tourmaline with pyrite and arsenopyrite. Knopf (1913, p. 51) noted the relation of gold with arsenopyrite in similar veins in the Helena region and in the Rimini district specifically. "Horses" of early tourmalinized quartz monzonite (stage 1) occur both in the hanging wall and footwall of the vein. Gold occurs in the arsenopyrite.
4. White vein quartz in small, parallel fractures cutting black tourmaline - quartz. Many small fractures up to 1 in wide contain minor amounts of pyrite. (At other mines, this is a main mineralization stage, whereas in the Lilly Mine it is relatively minor. No valuable minerals introduced at this stage.)
5. Lead-zinc-silver mineralization. Pyrite, sphalerite, galena and sulfosalts and minor amounts of quartz. Two types of occurrences are as follows:
  - (a) Base metal minerals cement and replace brecciated black tourmaline quartz (2 above) with white quartz veinlets (4 above) as shown in Figure 99.
  - (b) Fissure fillings in veins which cut diagonally across main vein zone in an enechelon pattern of "gash" veins. Similar ore shoots occur at Julia and Anna R. veins, based on observations drawn from old report at Julia and observations made by the Newman brothers at the Anna R.

There were several different surges of base metal mineralization which followed fractures formed at different times and which occur in or adjacent to the main veins. Sphalerite veins are cut by galena-bearing veins, etc.



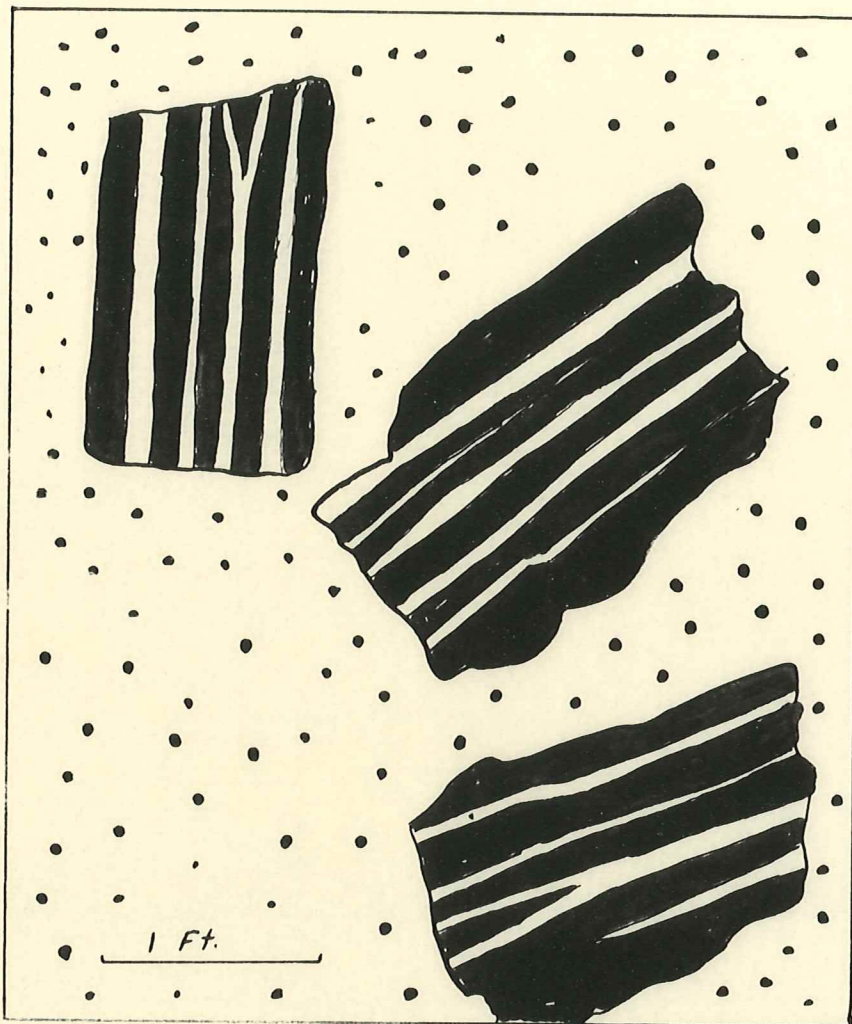


Figure 99a. Black tourmaline-quartz vein matter fractured and filled with white quartz veinlets, became brecciated and cemented with younger sphalerite, galena and quartz (dotted area). Lilly mine. (After Regnier 1951)

6. Quartz occurs in small fractures and breccias, the latter involve brecciated base metal mineral fragments cemented with quartz. Open spaces are incompletely filled; comb structure quartz crystals occur at this stage. A few hairs of millerite were found in vugs at the Lilly Mine.
7. Chalcedonic quartz stringers are minor at the Lilly Mine. At the Third Term Mine, veinlets of chalcedonic quartz cut sphalerite as shown in Figure 83.
8. Carbonate veinlets are not known at the Lilly Mine but are found in other mines in quartz monzonite. At the lower tunnel below the Pauper workings, ankeritic carbonate veinlets were noted cutting base metal minerals in boulders on the dump.

The width of the vein zone at the Lilly Mine is not precisely known. It is not less than 4 ft wide, and in some parts may be as much as 10 ft wide. Other tourmaline-quartz veins are even wider. The Twin Cities structure is, at one locality near the shaft, about 18 ft wide. Several different vein stages are indicated in the boulders found on the dump and in the ore bin. They include:

1. Quartz-tourmaline rock with tourmaline crystals as much as 3 in long.
2. Quartz with pyrite.
3. Quartz with pyrite, sphalerite and galena.

The Ontario vein is from 5 to possibly 20 ft wide at the outcrop. The wallrocks are impregnated with tourmaline. Chunks of massive quartz with pyrite are found on the dump along with base metal minerals



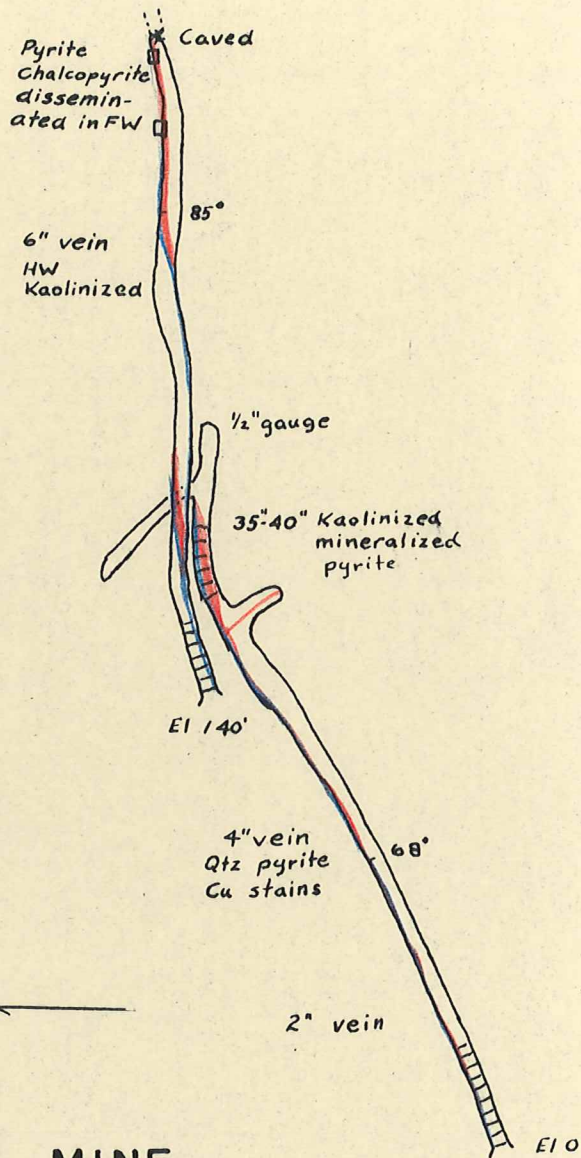
in chunks of ore. At the Pauper workings, pyrite and arsenopyrite occur with massive quartz-tourmaline vein matter, whereas galena which occurs with very little pyrite, is in distinctly younger fissures.

Quartz Veins. Quartz veins which contain no tourmaline appear to have somewhat less complex histories. Many of the veins are small, however the Anna R.-Copper King structure and the Iron Giant are wide veins, the former is one of the longest veins in the district. At the Julia Mine, the early barren vein matter is a massive quartz-pyrite vein in which younger fractures contain the base metals. An old mining report indicates that the valuable minerals were in ore shoots which cut diagonally across the main vein structure but pinch out on either side of the main quartz-pyrite vein. At the Iron Giant, only quartz and pyrite are known.

Variations in thickness of veins with changes of strike and dip. In the large veins, it is difficult to determine variations in the width of the vein along different parts which have variations in strike or dip. Some of the smaller veins, however, show the relations quite clearly. At the Hubcamp Mine the vein in the N. 65° E. zone is from 2 to 4 in, whereas in the east-west zone, it attains a maximum thickness of 35 to 40 in. See Figure 100. The same generalization was noted at the Keystone workings. The more nearly east-west segments of the veins apparently approach the tensional direction.

#### Veins in Volcanic Rocks.

Three different types of veins have been found in the volcanic rocks. They are:



HUBCAMP MINE  
ELLISTON MINING DISTRICT

0 50 FEET

Figure 100.



1. N. 65° E. striking veins, which have the same trend as the veins in quartz monzonite.
2. Northeast striking veins. These are subsidiary fractures to the large northeast faults.
3. Flat veins, found only on Nigger Mountain, possibly at the contact between two volcanic units or along flat shears.

N. 65° E. Veins. There are only a few veins of this trend. Actually, where a vein in the batholith crosses the margin of the batholith into the volcanic rocks, the strong veins disperse into small stringers and fractures. This is the case at the Julia Mine. Uphill from the Julia shaft, a bulldozer cut exposes the vein zone not more than 50 ft west of the contact. A few iron stained stringers occur in a zone about 10 ft wide. Just south of the saddle between Nigger and Treasure Mountains, there are several small stringers of quartz which have the same general trend as the veins in the quartz monzonite and have strongly bleached zones, but there are no indications of sulfide minerals other than pyrite.

At the Treasure Mountain claims, the vein pattern is not known, but because of the presence of tourmaline in the oxidized vein matter on the dump, as well as the very intense sericitization of the rocks, the mineralization is conceivably related to the underlying quartz monzonite.

The Blue Bird-Big Dick-Little Dick-Black Jack vein zone on Nigger Mountain extends for more than 1 mi across the mountain and is exposed in a vertical section of about 1500 ft. In a tunnel just above Little

Blackfoot Creek, a very complex, narrow fracture zone was noted as shown in Figure 101. The veins curve and split, and range in width from a fraction of an inch to about two feet. The veins widen out in the flatter parts of the vein, suggesting a component of reverse movement along the structure. (Newhouse 1942, p. 16.) The vein matter contains grayish quartz, pyrite and minor amounts of galena and sphalerite. The latter minerals are more abundant in the flatter parts of the veins. Thin gouge zones mark the borders of the larger veins segments.

The big Dick vein on the eastern end of the vein zone, ranges from 1 ft to nearly 5 ft wide. It changes strike and dip along the line of strike and line of dip, but because the workings are inaccessible, the structural controls of the wide parts of the veins are not known. They probably have the same features as those found in the smaller N. 65° E. striking veins.

At the Peerless prospect (NW cor. NE  $\frac{1}{4}$  sec. 13, T. 8 N., R. 7 W.) the portal of which is caved, a narrow steeply dipping quartz vein, which contains pyrite and some galena, strikes about N. 65° E. The wall rock is a dark andesite porphyry which is intensely bleached for a few inches adjacent to the vein, as determined from material collected on the dump.

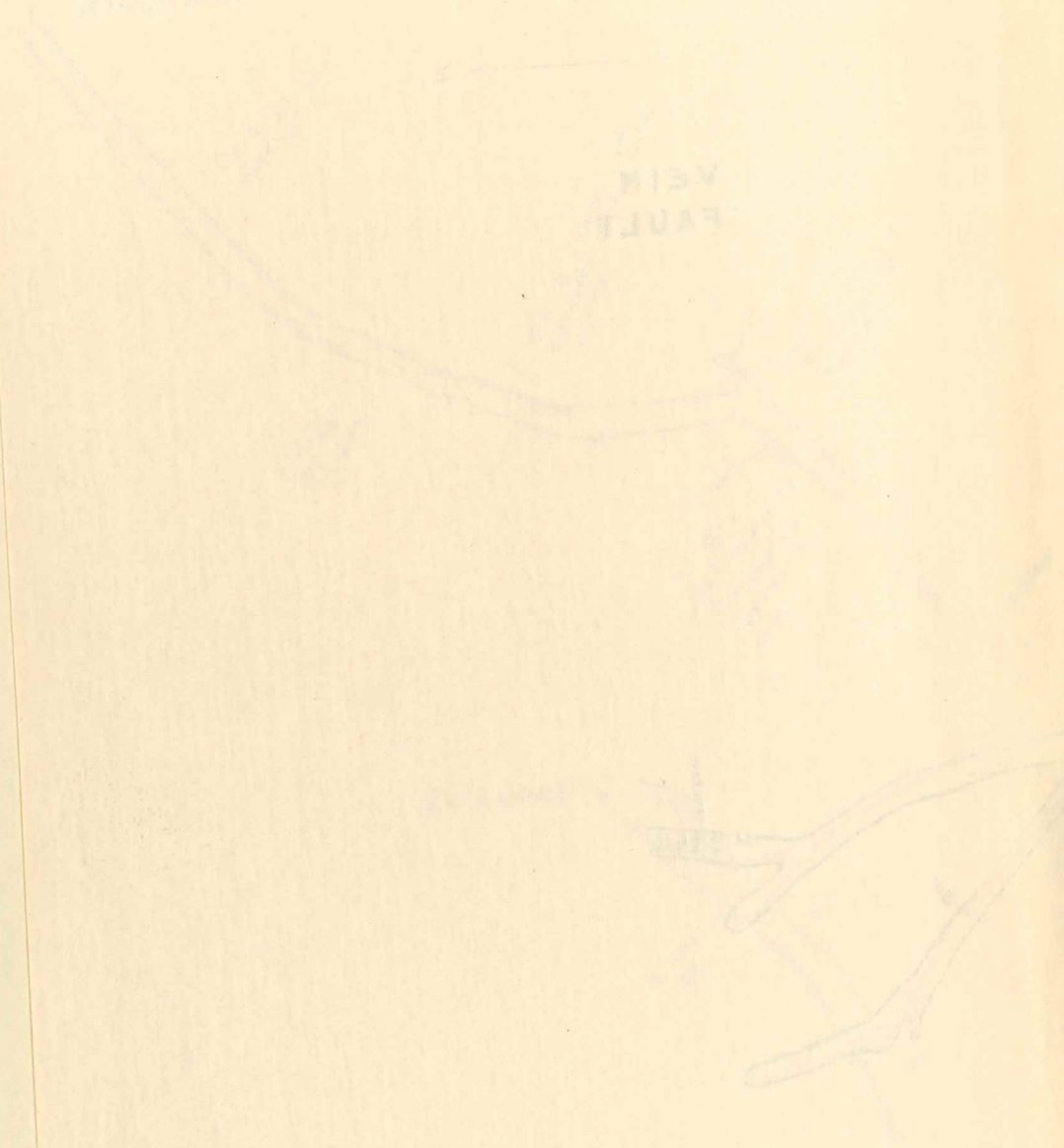
Northeast Veins. Veins which have the same strike as the major fault system, are along the east side of Little Blackfoot Creek and at the Kimball Mines (Kirstead Claims). The geological map shows several veins which strike parallel to the Little Blackfoot Creek at the Sadie, Hopkins, Flora and Negroes workings. These are complex fracture zones



200

200

VEIN  
FAULT



BLUEBIRD MINE  
ELLISTON MINING DISTRICT

50 FEET

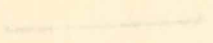


FIGURE 101

in which the basic andesite wall rocks are very strongly bleached and altered to clay minerals. Galena, sphalerite and pyrite occur in fissures and brecciated zones. At the Hopkins Mine there are two parallel veins each 4 to 8 ft wide, and at the Sadie claim the fracture zone must be 20 ft wide. None of the veins is persistently mineralized more than a few hundred feet along the strike.

At the Hopkins Mine, the two veins which crop out high on the steep hillside are not persistent to the tunnel located near the base of the hill. The veins pinch out at depth less than 200 ft. Two westerly dipping stringers were encountered in the tunnel. It is suggested that where fractures change in dip along the line of dip, they pinch out. This is typical of normal faults with dip slip movement. These veins may be in small normal faults along a major fault zone which is postulated in the valley of the Little Blackfoot Creek.

Some indication of the mineralogy of the ores that were mined out of the veins at the Hopkins Mine was obtained from a faulted segment of one of the veins which was located by bulldozing southwest of the small gulch that is south of the old workings. The cut opened a 30-in vein which was filled with massive galena. The galena gave way to pyrite and sphalerite in a matter of about 10 ft or so, but some excellent ore was shipped.

At the Sadie Mine, only pyrite was found in the wide intensely altered fracture zone. Silver was reported, but not observed, in the iron oxide crusts along fractures in a splintery breccia zone at the Negroes.



At the Kimball Mines, a brecciated zone of variable width, from a few feet to more than 20 ft, strikes about N. 30° to 35° E. The andesitic rocks are bleached, and traces of pyrite and galena were observed. Silver was reported in assays. The fracture zone is along a major fault, or closely related to one and was reported to be persistent for about 3/4 mi since it was exposed in a bulldozer cut on the west limb of Nigger Mountain.

Flat Veins. The only important flat vein is at the Big Dick Mine on Nigger Mountain. The vein is variable in dip and strike, but in its most productive area the dip is slightly to the west 5° to 10°. Where the ores are thickest, the vein is the flattest. It is probably a reverse fracture (Newhouse 1942.). The vein is distinctly older than the Black Jack vein (N. 65° E.) which clearly offsets the flat vein.

The mineralogy of the flat veins is similar to the steep veins, except that the gold content is appreciably greater. Arsenopyrite is minor in amount in the ores actually examined, however it is probable that the mining was advanced to non-economic fringes. There may have been more arsenopyrite in the ores mined than in the ores left. Pyrite, galena, sphalerite, quartz and carbonate are the minerals in the veins.

At new workings at the Hopkins Mine (south of the gulch) some high-grade gold ores were shipped from one part of the vertical vein at the intersection with a very small, flat quartz seam. The operator attributed the higher-than-average gold values to the small flat vein.

#### Veins in Sedimentary Rocks.

Only at the Carbonate King Mine south of Elliston (NW 1/4 sec. 13,

T. 9 N., R. 7 W.) have veins been found in the sedimentary rocks. There are two caved shafts, the collar of each is apparently in basal Kootenai quartzite. Siliceous carbonate rocks were found on the dump. The structure strikes N. 80° E. and is nearly vertical. The vein minerals on the dump are quartz, some pyrite, iron oxide and minor indications of lead and copper in gossan. About 150 yds north of the Carbonate King, another similar trending vein is indicated by a line of prospect pits. Irregular quartz veinlets  $1\frac{1}{2}$  in wide were noted on an outcrop.

#### Breccia Deposits.

In many of the fissure veins, the sulfides cement brecciated vein matter, and in a sense, constitute breccia deposits. However, they are not so classified because they have already been described in the discussion of vein formation. Only one mineral deposit is classified as a breccia deposit. At the Clark Mine (SW  $\frac{1}{4}$  sec. 19, T. 9 N., R. 6 W.) andesite is brecciated, bleached and cemented with pyrite and sphalerite and minor amounts of quartz, galena and tetrahedrite.

The breccia zone is in the hanging wall of gouge layer a few in wide which dips 10° to 30° southwest. Where the dip is about 10°, there is only a 1-in stringer of sulfide minerals above the gouge; where the dip is 18°, the breccia zone is about 1 ft wide; and where it dips 26° to 30°, the breccia zone is about 6 in to  $4\frac{1}{2}$  ft wide, and from the geochemical soil sampling, assumed to be at least 200 ft long. The breccia zone is distinctly wider where the gouge zone is steeply dipping.



The mineralized zone has a distinct linear trend in a northeast-southwest direction. The trend was determined by a geochemical soil survey made of the area. The results are shown in Figure 102.

#### Replacement in Limestone

At one locality on the west side of Little Blackfoot Creek opposite the road to the Big Dick Mines (NE  $\frac{1}{4}$  NW  $\frac{1}{4}$  sec. 12, T. 8 N., R. 7 W.) two adits were driven into the hillside. On the lower dump limestone, with pyrite and galena, was found. The adit was partly open, and a sketch map was made of the open workings. Black, silty limestone was seen in the open adit, and in a partly caved drift it was possible to see strongly bleached limestone over the cave. The structural control for the deposit was not determined.

The only limestone which seems to be reasonable at this locality is the upper Kootenai "gastropod" limestone bed, however, no fossils were found. About 2 mi west, the volcanics rest on Colorado shale.

#### Wall Rock Alteration

The formation of sericite in the walls of the hydrothermal veins has been discussed as typical of hydrothermal stage crystallization of quartz monzonite. Its formation is characteristic of all the hydrothermal quartz veins whether or not they contain valuable ore minerals. An argillic alteration zone occurs between the sericite zone and fresh rock.

The argillic zone is not well studied because completely satisfactory samples of argillic wall rock alteration are not available.

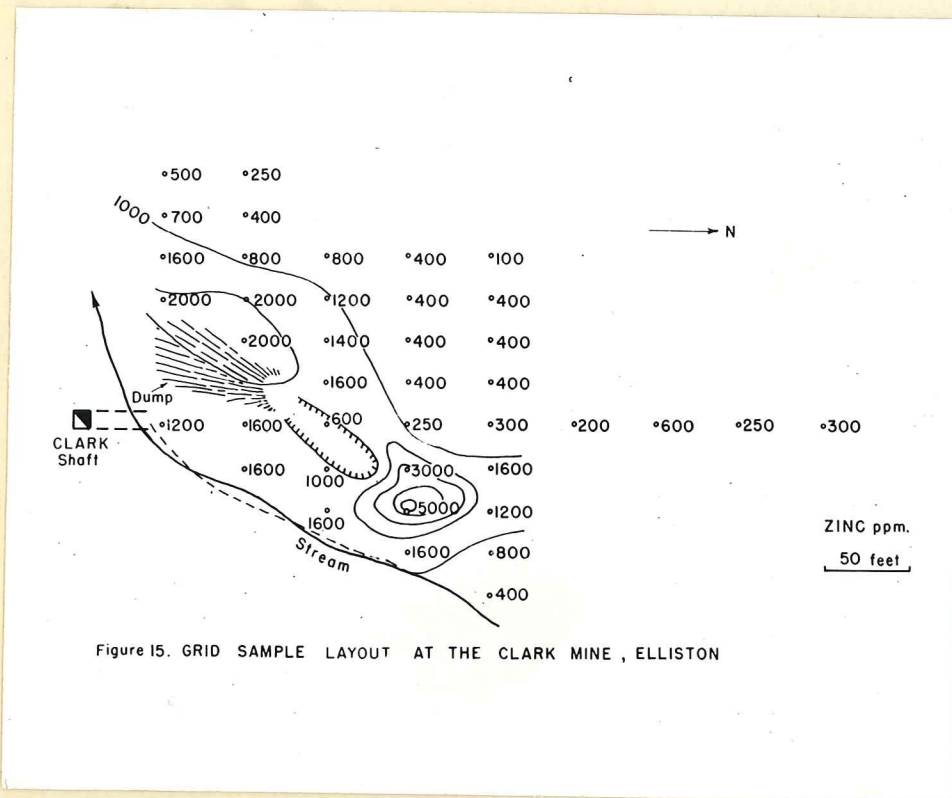


Figure 102. Geochemical soil sample grid at Clark Prospect. Note northeast trend of high zinc concentrations.



Because argillic minerals form by normal weathering processes and clay mineral formation may be stimulated by acid mine waters percolating through permeable wall rocks, the greatest possible care must be taken in the collection of wall rock samples for detailed petrographic studies. At no locality in the Elliston district were ideal conditions found because no new cross cuts were made in fresh rock below the zone of weathering at any time during this investigation.

Suites of wall rock samples are difficult to obtain in any event because there are so few crosscuts in the district anywhere. The best samples were collected from the lower level of the Lilly Mine in a crosscut. This suite of samples was far from ideal because the lower levels had been under water for many years, and the altered zone was subject to acid mine waters during the several months that the lower level was open to the atmosphere after dewatering, and undoubtedly was subject to equally unfavorable conditions at the time the lower level was opened originally.

#### Veins in Quartz Monzonite

The wall rock alteration along veins in quartz monzonite show the following zones:

1. Vein matter-quartz + pyrite + tourmaline + base metal minerals
2. Sericite zone + quartz + pyrite
3. Argillic zone--plagioclase of quartz monzonite altered to clay minerals. White and light gray alteration probably kaolin group minerals, greenish alteration contains swelling clays of montmorillonite group. The greenish clays occur at the outer edge of the alteration zone.

4. Fresh quartz monzonite--partial alteration of biotite to chlorite at the fringe of the argillic alteration zone.

At the margin of the veins, nearly white to buff sericite occurs with quartz and some pyrite. Intense and virtually complete sericitization of the minerals is common at the edge of the vein, but grades outward from the vein to sericitized quartz monzonite in which relics of potash feldspars are preserved. Figure 103 shows sericite alteration in nearly completely sericitized wall rock.

At the edge of the sericite zone, white, chalky alteration products are noted and the texture of the original rocks is preserved. Plagioclase is the most intensely kaolinized feldspar, and hornblende and biotite are strongly altered to chlorite and a chlorite-like mineral with higher interference colors than normal chlorite.

Nearly fresh quartz monzonite beyond the bleached zone contains altered plagioclase crystals which have a greenish hue. Thin sections made of the rock must be cut and ground in kerosene because the clay minerals swell in water. They are probably montmorillonite clay minerals which are noted only in plagioclase feldspar. Biotite is incompletely altered to chlorite, and potash feldspar is fresh.

At the Lilly Mine, the two main alteration zones were noted in the crosscut at the bottom level, and in a small crosscut off the tunnel level in the oxidized zone. Similar alteration is indicated at each of the properties where observations were possible. The same appears to be true for veins in quartz monzonite throughout the northern part of the Boulder batholith (Robertson, Klepper and Pinkney, 1956).



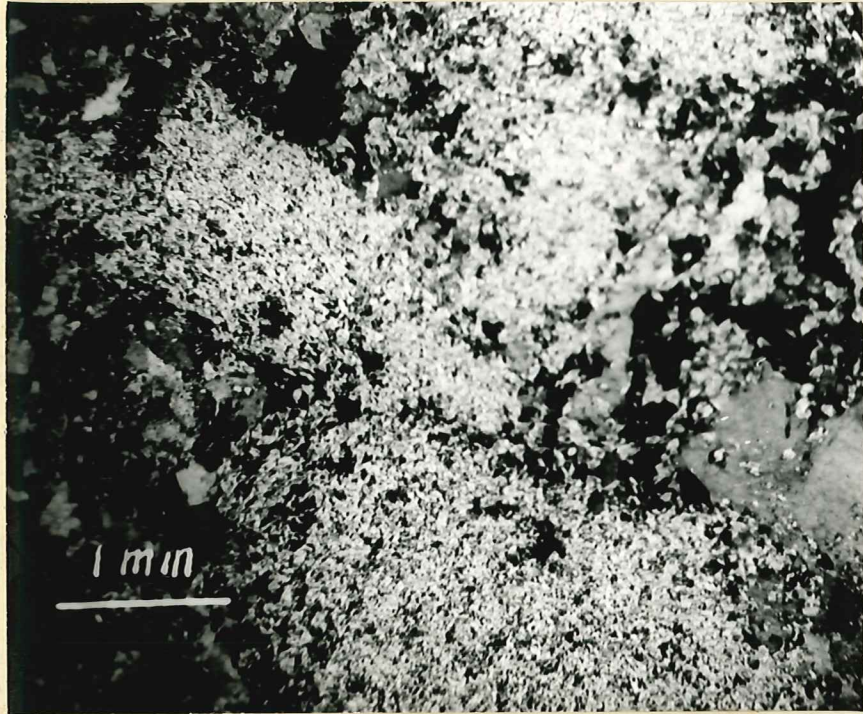


Figure 103. Photomicrograph of sericite wall rock alteration next to Lilly vein. Crossed nicols.

## Volcanic Rocks

The alteration of the volcanic rocks is conspicuous because of the bleaching of the dark rocks at the margins of the veins. The bleached zone ranges from a fraction of an inch to many feet along small stringers and wide fracture zones. Alteration of feldspars to sericite and clay minerals, and the liberation of lime for the formation of carbonate minerals, are the important mineralogical changes. Sericite is commonly found at the margin of many of the veins with bleached argillic alteration away from the vein. At the Treasure Mountain workings, tourmaline and pyrite are found in intensely sericitized volcanic rocks of the main workings.

In the veins on Nigger Mountain and at the Monarch Mine, sericite has been observed next to the veins and argillic minerals farther away from the vein. The argillic zone contains an abundance of ankeritic carbonate and the mafic minerals are altered to chlorite and allophane.

### Origin of Wall Rock Alteration.

In the quartz monzonite rocks, the sericite alteration at the margins of the hydrothermal veins has been described as a possible potash migration towards the vein structure at the hydrothermal stage which is analagous to the potash migration to the selvage zone in tourmaline-pyrite segregations and joint fillings discussed in the section on the hydrothermal stage of magmatic crystallization. The alteration products are not clearly linked to hydrothermal solutions containing base metals percolating through the veins. This is true



not only in the Elliston district, but in the whole of the northern part of the Boulder batholith (Robertson, Klepper and Pinkney, 1956).

A considerably more detailed investigation of wall rocks will be necessary before the relations of the argillic zone to the sericite zone is understood, and the origin of the wall rock alterations known without the possibility of multiple interpretations.

The sequence of alteration envelopes in the Elliston district is the same as that found surrounding the hydrothermal veins at Butte. Sales and Meyer (1948) found that sericite, surrounded by two argillic zones, encased every vein, no matter what the nature of the mineralization in the vein. They concluded that:

"Successive zones of sericitized and argillized quartz monzonite occur around every ore-bearing fracture, regardless of size, attitude, or relative age; provided the lower intensity alteration zones are not flushed out by coalescence of the alteration envelopes of adjacent structures, as is the case in the Central zone. Furthermore, the two types of alteration always occupy the same relative positions: sericite adjacent to the ore-bearing vein; clay minerals always between sericitized rock and fresh quartz monzonite. And among themselves, the clay minerals always are arranged with the high kaolinite end of the series nearer the vein and the montmorillonite end toward fresh rock.

"According to any theory based on the premise that an epoch of sericitization followed an epoch of argillization, the geometric pattern of alteration which envelopes every mineralized fissure in the Butte district would require that the invading zone of sericitization never advanced completely through the argillized zone to attack fresh rock directly. It would have to advance only far enough out from every part of the respective channel walls to leave unchanged the characteristic band of argillization between the advancing sericite zone and unaltered rock. Furthermore, if the solutions which produced sericite were not capable of causing argillization, and if sericitization therefore wholly followed all argillization, sericitization could not have commenced until the very end of circulation in the channel maze after all mineralized fissures of all ages had been formed and sulphide deposition had all but ceased completely; for in areas where a complete structural sequence may be observed, argillization is always present, and in its customary zonal position, in the wall



rocks of Northwest-faulting and Northeast-faulting fissures and even in the last formed fractures of the district along which there was any mineralizing activity. This would place sericite in the untenable position of postsulphide age and run counter to the generally observed ubiquitous close association of sulphides and sericite at Butte and in other deposits of its type." (Sales and Meyer 1947, p. 32.)

If the alteration zones in the Elliston district are produced in part by migration out of the walls towards the veins, the argillic alteration zones probably formed at the same time as the sericite zone because the geometric relations of the zones, so far as is known, is always the same, not only for the Elliston district, but for veins throughout the Boulder batholith. If the alteration zones are formed by circulating solutions from the vein, the same arguments employed in Butte apply in Elliston for the origin of the successive envelopes.

The conclusions at Butte, and these derived in the Elliston district, are in striking contrast to those arrived at by Lovering (1949) who concluded that the different alteration phenomena found in the East Tintic district in rocks overlying intrusive igneous rocks were produced by progressively changing composition of the hydrothermal solutions "at the source as well as during their journey toward the surface. . . these solutions intermittently from the magmatic source and gave rise to a series of successive semi-independent stages of alteration of which the last and least extensive was metallizing." (Lovering, et.al. 1949, p. 62)

If wall rock alteration is an auto-alteration process in the Elliston district, as has been suggested, then the uniform sequence of alteration around all fissures throughout the northern part of the batholith may be reasonably explained. If the solutions which produced



the different mineralogical changes in the main vein structures got out into the volcanic and sedimentary rocks at different times, due to differences in the time of formation of through-going fractures, a variety of alteration products in the overlying rocks might be expected.

If, however, the wall rock alteration in the batholithic rocks is due to hydrothermal solutions derived from some deep-seated reservoir below the limits of observation, then it seems impossible to account for the same type of wall rock alteration along all of the structure, produced at different times, in such widely scattered areas throughout the batholith. The alteration adjacent to the tourmaline structures might be expected to be somewhat different from the quartz-pyrite structures, and different from the base metal stage.

The points which seem to correlate the observations of Lovering with those in the Elliston district are: (1) the sequence of stages at which intense alteration was produced prior to any base metal introduction, and (2) the late stage of the base metals in the vein formation sequences in the two areas.

In the volcanic rocks, the wall rock alteration needs to be studied with great care from each of the vein types and from different stages of vein formation, particularly from veins which contain no valuable minerals, to determine if the alteration is essentially the same in all veins or if differences do occur which might be correlated to different stages in vein formation in the batholithic rocks.

The contribution of this investigation to wall rock alteration studies is that virtually unmineralized veins have intense wall rock alteration, veins which contain no trace of base metals are strongly altered, and veins which do contain base metals are likewise altered.

The presence of base metals may be of negligible importance in the formation of an alteration envelope around a large barren vein structure. The need is to investigate more thoroughly the wall rock alteration around barren veins to compare with the alteration along mineralized veins.



## OUTLINE OF PHYSIOGRAPHIC HISTORY

## Upland Surface

The most prominent physiographic feature in the region is a rolling upland surface which is dissected by narrow valleys 1,000 to 3,000 ft deep. Pardee and Schrader described the surface as follows: "In most of the mountain groups the main summit areas are flat or gently dipping and accord in form and height with a surface of moderate relief and gentle contour that formerly extended over the whole region." (Pardee and Schrader 1933, p. 3) They indicated that the level is at about 7,000 ft. In the southern part of the Elliston district, the upland surface is at about 7,200 ft  $\pm$  200 ft. Bisen Mountain at elevation 8028 ft is interpreted as a monadnock standing above this surface.

Little Blackfoot and Telegraph Creek have dissected the surface to such an extent that only small remnants are preserved in the northern part of the district.

Remnants of rhyolite are found at the general level of the upland surface and extend to elevations as low as 6,600 ft along Ontario Creek. Actually there are two rhyolite units and their relative age relations are not known, nor has the geologic age of either been established. The simplest explanation for the relation of the rhyolites to the upland surface seems to be: (1) development of the surface; (2) uplift followed by partial dissection of the surface; (3) extrusion of rhyolite onto the surface and into some valleys; (4) further dissection which removed much of the rhyolite and produced the present topography. If a substantial age difference is established between the two rhyolite units, the outline will have to be modified.

The uncertainty regarding the age of the rhyolite is emphasized by the conflicting information. Pardee and Schrader (1933, p. 21) consider the rhyolite exposed near Avon to be pre-Oligocene. Knopf (1913, p. 41) considers the age of the rhyolite near Blossburg (north-east of Elliston) to be possibly upper Miocene. In the discussion of the youngest rhyolite in the district, the possibility that it may be Pliocene in age has been mentioned. (See p. 82)

An accurate dating of the rhyolite would be of value in placing a younger limit on the age of the upland surface. Fenneman (1931, p. 192-194) discusses two hypotheses for the formation of the upland surface. The alternatives indicate an Eocene or Pliocene age. Fenneman apparently favors the younger alternative.

### Glaciation

#### Older Stage

Western Montana was subjected to glaciation during several stages during the Pleistocene. Within the district, two stages have been recognized. Evidence of the earlier stage is found in isolated remnants of glacial drift high on the west side of the Continental Divide east of Elliston and by patches of drift on the divide farther south in the district. Alden mapped a patch of older drift east of Elliston (1953, p. 87 and Plate I).

Probably during this early stage, a local ice cap centered in the Thunderbolt-Cliff Mountain area just southwest of the district. (Ruppel, personal communication) Evidence of glaciers which radiated north and east from the cap is conspicuous in the Elliston district. Glacial deposits are found on the Continental Divide northeast and southeast of



the Ontario mine. Northeast of the mine (in sec. 12) a large, weathered erratic of andesitic tuff is found as shown in Figure 105. Similar andesitic tuff crops out on Cliff Mountain, but is not known to crop out anywhere inbetween. In order to have come from Cliff Mountain, or possibly Thunderbolt Mountain, it would have had to travel across Little Blackfoot, Ontario and upper Telegraph Creek valleys. Conditions for the formation of an ice cap could have been set up by ice damming of Little Blackfoot Creek northwest of the Elliston district. As ice continued to accumulate, it flowed northeast and east ( as well as south according to Ruppel) across the drainages mentioned and across what is now the Continental Divide. The patches of high level drift east of Elliston may have been derived from glaciation at this stage.

Evidence of ice movement across interstream divides is found at a number of localities in the district. They include: (1) andesitic boulders in glacial drift 1 mi north of Bullion park (NE  $\frac{1}{4}$  sec. 2, T. 8 N., R. 6 W.), the narrow ice-grooved notch across the west ridge of Jericho Mountain (SW  $\frac{1}{4}$  sec. 34), and probably in the northeast trending grooves occupied by small lakes, situated at different levels, on the divide between Telegraph and Ontario Creeks southeast of the Pauper area.

The age of the older drift is uncertain. Horberg (1954, p. 1102) noted somewhat similar remnants of drift in the Waterton Lakes region and designated them as pre-Wisconsin. The older drift mapped by Alden (1953, p. 87 and Plate I) is called Illinoian or early Wisconsin (Iowan).





Figure 104. Older stage glacial drift in road cut on U. S. Highway 10 N about 4 mi east of Elliston.



Figure 105. Glacial erratic of andesitic tuff on Continental Divide northeast of Ontario mine. Quartz monzonite underlies the glacial drift.



### Younger Stage

During the younger stage, glaciers were confined to the valleys. Little Blackfoot and Ontario Creeks occupy typical U-shaped valleys that contain thick deposits of glacial outwash. Patches of glacial debris are found high on the sides of the valleys, 500 ft above the valley floor at the Hopkins mine. The glaciers extended downstream to Elliston.

Two terminal moraines are recognized. The larger is located about 1 mi northeast of Elliston and contains boulders obviously derived from the headwaters of Little Blackfoot Creek. The moraine is about 100 ft thick above the valley floor outwash gravels and has a subdued hummocky surface. The end of this moraine was cut off by Little Blackfoot Creek so that it is not an obvious feature from the highway. It is much the larger of the two moraines and the only one large enough to account for the glacial erosion found in Ontario and Little Blackfoot Creeks.

A small, hummocky moraine, located about 2 mi southeast of Elliston is mapped by Alden (1953, p. 87 and Plate I). An early Wisconsin age is indicated. Because of the very minor erosional modification of the younger moraines, and because of their relation to the patches of older drift, these deposits may be correlated with Horberg's early Wisconsin drift. (1954, p. 1103)



Figure 106. Oblique air photo showing cirques on Cliff Mountain. Note westerly dip on the andesitic volcanic rocks.

The cirques on Cliff Mountain were probably occupied by the younger stage glaciers. The valley floor is at about 8,000 ft. As may be noted in Figure 106, the cirques are not sharp and clean like cirques found at higher elevations in the Elkhorn Mountains and in the Flint Creek Range (west of the Deerlodge Valley). It is probable that no glaciers occupied the Cliff Mountain cirques in the most recent, or Late Wisconsin, glaciation recognized in western Montana.



## PETROGENESIS

This section discusses ideas about the petrogenetic relations of the volcanic and intrusive rocks, the emplacement of the Boulder batholith and the origin of the mineral deposits. These ideas have developed as a result of investigations made for this report and from observations made throughout the area of the batholith. The ideas are based on the interpretation of the known facts. The conclusions on the subject are tentative and are in part controversial. In order to be able to attempt to deal with the problems of rock petrogenesis, a number of separate topics bearing on the topic must be brought together.

## The Volcanic Pile

The volcanic pile over the region of the batholith is comprised principally of pyroclastic volcanics (tuffs, lapilli tuffs and pyroclastic breccias which are, in part, welded), some flows and flow breccias and minor amounts of volcanic-derived sediments. The composition of the pile is dominantly andesitic, but includes rocks as acidic as latite and as basic as olive basalt. Sills and a very few dikes are mostly of basaltic composition.

The chemical composition of the pile can only be approximated because there are insufficient chemical data. Judging the composition of the pile to be dominantly andesitic, with latitic and basaltic rocks, three analyses considered typical of the three principal rock types were weighted--6 parts andesite, 3 parts latite, and one part basalt--to provide an approximation of the composition of the volcanic pile based on my impression of the relative amounts of the different rock

types found throughout the Upper Cretaceous volcanic pile. The composite analysis is given in Table 20. Note that the composition is not greatly different from the average of 10 analyses of quartz monzonite which are also included in the table. With respect to alumina, lime soda and potash (the feldspar ingredients) the similarity is virtually an identity.

#### Sills in Sedimentary Rocks

In the underlying sedimentary rocks, basaltic sills are abundant. There are some sills in the Elliston district in the Cretaceous rocks, and even more are exposed in the sedimentary rocks to the west. On the east side of the batholith, particularly in the lower Paleozoic rocks, there are a great many sills. At some localities, practically each shale unit in the section is the site of a sill which was intruded by simple dilation, hence the thickness of the sedimentary section is increased by the total thickness of the intrusives. The shale beds were somewhat baked and thermally metamorphosed along the contacts. Most of the sills are basaltic and one rather thick sill is a fine-grained granogabbro. Because of the basic composition of the sills and because the basaltic flows are the youngest rocks in the volcanic pile, the sills are considered to have intruded relatively late in the pre-batholithic volcanic period.

By the time the pile reached its maximum thickness, and after the sills were intruded, the physical properties of the crust in this general region had been materially changed. The crust obviously became thicker and stronger and the pile served to increase the lithostatic



Table 20

## Chemical Composition of Volcanic Pile

	1	2
SiO <sub>2</sub>	61.2	64.3
Al <sub>2</sub> O <sub>3</sub>	16.6	16.5
Fe <sub>2</sub> O <sub>3</sub>	2.1	2.3
FeO	3.6	2.7
MgO	1.9	2.4
CaO	4.6	4.6
Na <sub>2</sub> O	3.4	3.3
K <sub>2</sub> O	4.0	4.3

1. Chemical composition of volcanic pile based on weighted average of 6 parts andesite, 3 parts latite and one part basalt.
2. Average of 10 analyses of quartz monzonite of Boulder batholith.

head on the volcanic vents. The sills increased the resistance to intrusion by welding the easily intruded horizons.

#### Structures Prior to Emplacement of the Batholith

Throughout the period of extrusive activity, there was no significant deformation of the accumulating pile. At least no such evidence has been recognized in the Elliston district.

The strong structural deformation came after the volcanic pile had formed. On the east side of the batholith, the sediments and volcanics were strongly folded, overturned and thrust faulted. Two prominent anticlinal axes and the large thrust fault zone curve out and around the batholith as shown diagrammatically in Figure 104. The orientation of the anticlinal arcs and the displacement eastward of the northwest-southeast trending thrust zone suggests a "push" from the west. On the west side of the batholith, and over the batholith itself (judging from the attitude of large roof remnants) the structural deformation was very mild. These structures are older than the intrusive rocks now exposed, but these structures are considered to be intimately related to the stresses which not only produced the folds and thrust faults, but also forced the batholith into its present location.

#### Emplacement of the Batholith

##### Marginal Relations of the Batholith

In most areas in the northern part of the batholith, volcanic rocks comprise the roof of the batholith. The gentle westerly dips of



of the contact in the Elliston district and large roof pendant in the Wickes district indicate that the roof of the batholith was remarkably flat over much of the batholith.

The northern contact of the batholith, mapped by Knopf (map unavailable), and parts of the eastern side, mapped by the U. S. Geological Survey, show quite straight intrusive contacts. In some areas on the east side, steeply dipping sedimentary beds are parallel to the dip of the contact and in other areas, the sedimentary rocks are truncated by the contact along straight lines. The straight line contacts which appear to cut across pre-batholithic structures, may be intrusive fault contacts.

At two localities in the Elliston district, thin sill-like margins of the batholith are indicated. They are found in both the granodiorite and main stage quartz monzonite. Elsewhere no bottom of granitic rocks has been observed. Where the roof rocks are only slightly disturbed, the batholith has a flat, concordant roof. Where the rocks are deformed, the contacts are partly conformable and partly may be interpreted as intrusive fault contacts.

#### Magmatic Differentiation.

Magmatic differentiation is indicated in the different rock types found in the extrusive volcanic rocks which range in composition from latite to olivine gabbro, as well as in the intrusive rocks which have a range in composition from granogabbro to alaskite. In the volcanic series, the rocks grade upward from basic andesite to andesite to latite to basalt. There are interlayered basalts more or less throughout



of the contact in the Elliston district and large roof pendant in the Wickes district indicate that the roof of the batholith was remarkably flat over much of the batholith.

The northern contact of the batholith, mapped by Knopf (map unavailable), and parts of the eastern side, mapped by the U. S. Geological Survey, show quite straight intrusive contacts. In some areas on the east side, steeply dipping sedimentary beds are parallel to the dip of the contact and in other areas, the sedimentary rocks are truncated by the contact along straight lines. The straight line contacts which appear to cut across pre-batholithic structures, may be intrusive fault contacts.

At two localities in the Elliston district, thin sill-like margins of the batholith are indicated. They are found in both the granodiorite and main stage quartz monzonite. Elsewhere no bottom of granitic rocks has been observed. Where the roof rocks are only slightly disturbed, the batholith has a flat, concordant roof. Where the rocks are deformed, the contacts are partly conformable and partly may be interpreted as intrusive fault contacts.

#### Magmatic Differentiation.

Magmatic differentiation is indicated in the different rock types found in the extrusive volcanic rocks which range in composition from latite to olivine gabbro, as well as in the intrusive rocks which have a range in composition from granogabbro to alaskite. In the volcanic series, the rocks grade upward from basic andesite to andesite to latite to basalt. There are interlayered basalts more or less throughout



the section, but they are more abundant in the upper part of the pile. In the batholithic rocks, somewhat the same trend is shown from granogabbro to basic granodiorite to quartz monzonite to alaskite. Within each individual unit, there is little evidence of vertical or lateral differences in mineralogical composition except the formation of alaskite in quartz monzonite.

In quartz monzonite, the separation of aplite-alaskite is the conspicuous differentiation phenomenon. The process is described in some detail in the section on the deuteric stage of crystallization of quartz monzonite. The formation of aplite-alaskite is considered to be the product of deuteric differentiation whereby early-formed quartz monzonite is replaced by deuteric solutions to produce alaskite.

Only at one locality have vertical variations in composition been recognized within a body of quartz monzonite. The lower part of a small sill-like marginal apophysis is marked by the presence of olivine, pyroxene and labradorite. Grading upward, the rock becomes normal quartz monzonite with conspicuous deuteric phenomena. The differentiation is a deuteric feature. There is no clear evidence of concentration of mafic minerals at the base due to crystal settling. Chemical Petrology.

All of the chemical analyses of igneous rocks from the region of the Boulder batholith were assembled and variation diagrams compiled to see if the compositions of the various rock types have any correspondence with each other. Variation diagrams compiled by Larsen (1938)



are probably more successful than most in showing relations of rocks in a petrogenetic province.

Figure 105 shows a Larsen-type diagram which relates  $1/3 \text{ SiO}_2 + \text{K}_2\text{O} - \text{CaO} - \text{F}_2\text{O} - \text{MgO}$  to  $\text{CaO} + \text{FeO} + \text{MgO}$ . On this diagram, the basic rocks are at the left side and the acidic rocks, such as aplite, are at the right side. Practically all of the points, each of which represents a chemical analysis of a rock, fall into a straight line. Both intrusive and extrusive rocks fall into the same line. This suggests that the chemical changes, brought about by magmatic differentiation and by deuteric alteration and differentiation, take place systematically in the same chemical sense. Different petrogenetic provinces have lines of somewhat different slope and position.

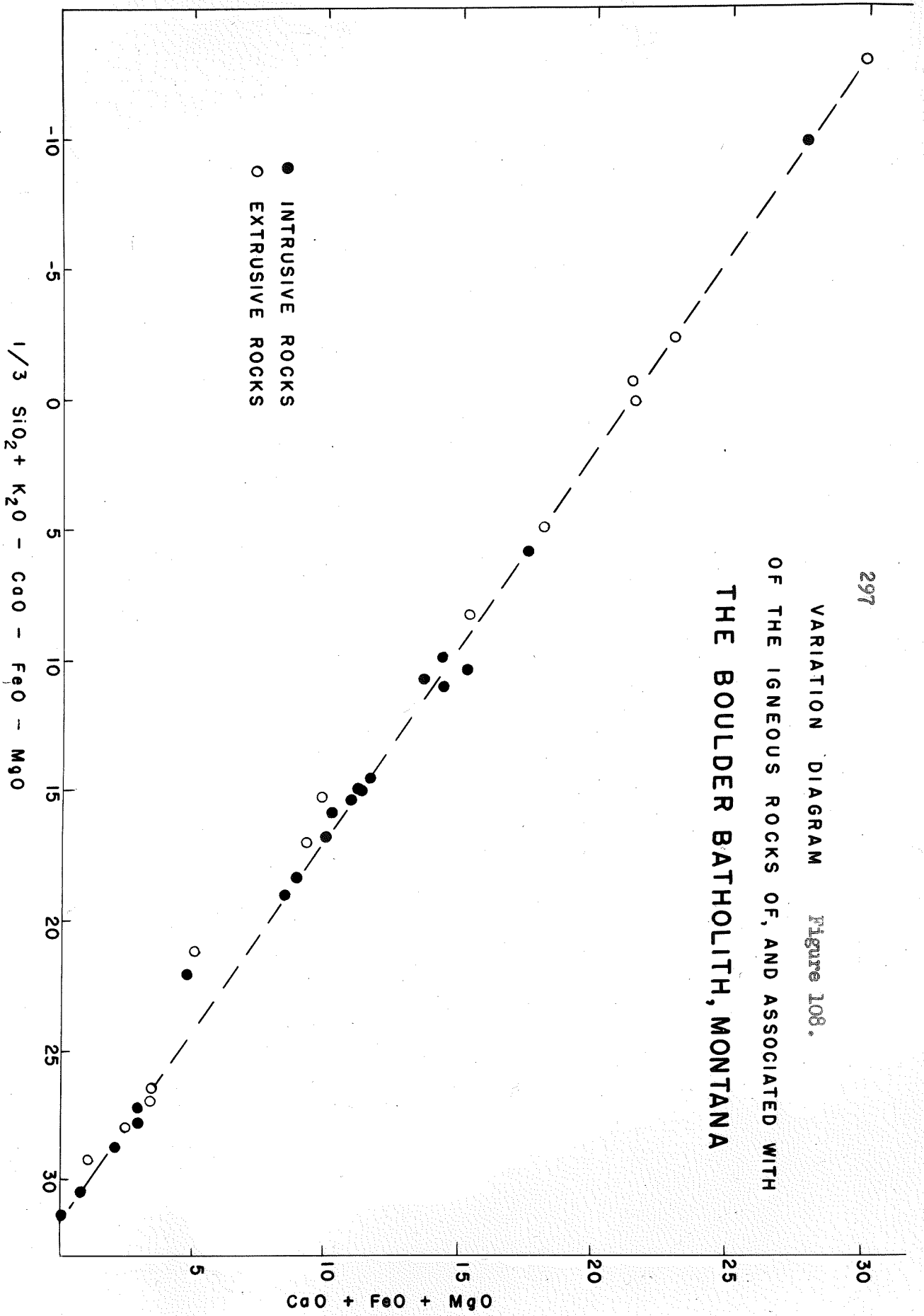
Figure 106 is a series of variation diagrams which have the same abscissa as in Figure 105, but the individual oxides of a chemical analysis are plotted on different ordinates. The great majority of the analyses fall into smooth curves. This indicates that each of the chemical constituents in the rocks responded to changes in the equilibrium relations of the magma in the same chemical direction sense in both intrusive and extrusive rocks. Curves of other petrogenetic provinces (not herein shown) have similar trends, but the details are strikingly different, particularly in the  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  curves.

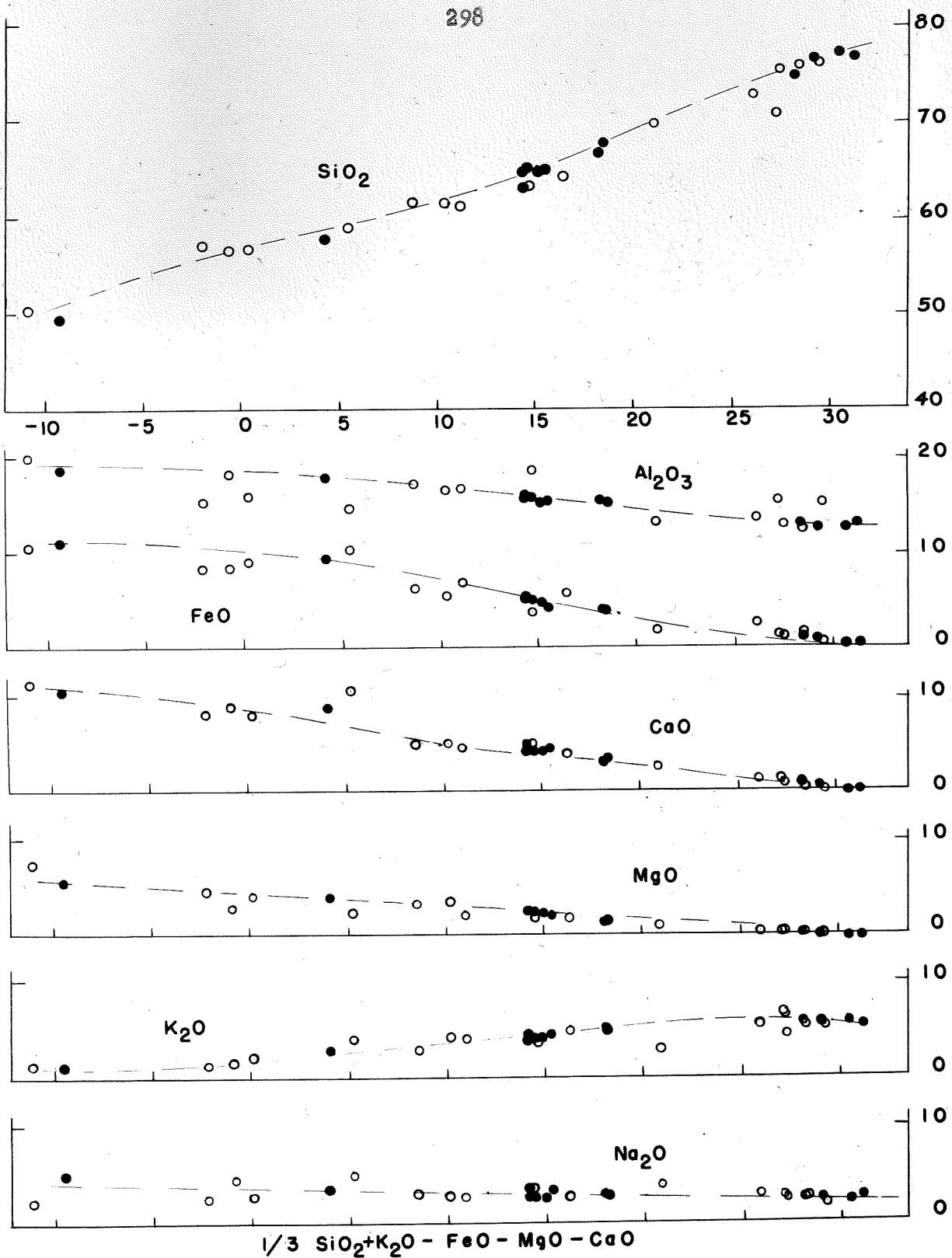
I consider these data to be significant because they seem to indicate the following:

1. Both the intrusive and extrusive rocks have the same relative proportions of the chemical elements for the entire range of chemical compositions of the rocks. Thus the volcanic rocks were but the first phase of a magmatic petrogenetic epoch.



VARIATION DIAGRAM  
OF THE IGNEOUS ROCKS OF, AND ASSOCIATED WITH  
THE BOULDER BATHOLITH, MONTANA





VARIATION DIAGRAMS OF THE IGNEOUS ROCKS OF, AND ASSOCIATED WITH  
THE BOULDER BATHOLITH, MONTANA

● INTRUSIVE ROCKS

○ EXTRUSIVE ROCKS

Figure 109.



2. The chemical data are negative evidence for assimilation. If any significant amount of assimilation of sedimentary rocks took place in the intruding batholithic rocks, thorough mixing must have been accomplished prior to the outpouring of the volcanics! The argument for assimilation is discussed under the next sub-heading.
3. The chemical data support the field and petrographic evidence that the Boulder batholith is magmatic in origin.

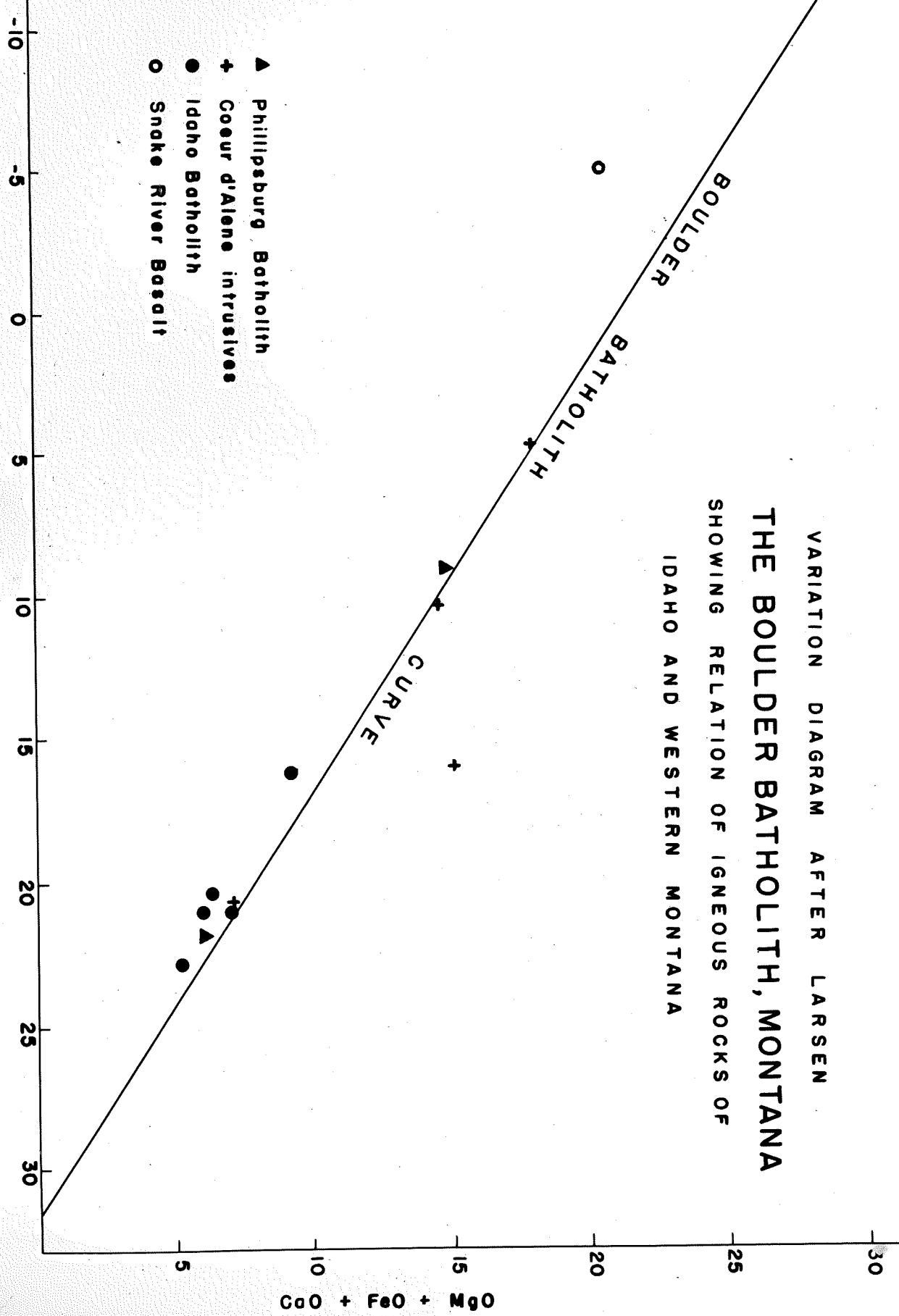
In Figure 107, the straight line from Figure 105 is used as a reference. Chemical analyses of intrusive rocks from the Coeur D'Alene district, the Phillipsburg batholith and the Idaho batholith are plotted. A single analysis of Snake River basalt is also included. The Tertiary (?) intrusives in the Couer d'Alene district and the Philipsburg batholith are petrographically similar to the Boulder batholith. The chemical data suggest that they may be genetically related. Four of five of the analyses of the Idaho batholith do not fall onto the same line, but seem to suggest the position of a different line. They are clearly not genetically related to the Boulder batholith. The Snake River basalt is in another world, chemically.

#### Method of Emplacement of the Batholith

Three possible mechanisms of emplacement of the batholith have been considered. They are: Injection of a fluid magma by (1) magmatic stoping or (2) dilation; and granitization. Chemical evidence does not support wholesale stoping and field and petrographic evidence belay significant granitization.

VARIATION DIAGRAM AFTER LARSEN  
**THE BOULDER BATHOLITH, MONTANA**  
 SHOWING RELATION OF IGNEOUS ROCKS OF  
 IDAHO AND WESTERN MONTANA

$$\frac{1}{3} \text{SiO}_2 + \text{K}_2\text{O} - \text{CaO} - \text{FeO} - \text{MgO}$$





### Dimensions of the Boulder Batholith.

Up to the present, only the northern outcrop area of the Boulder batholith has been considered. In order to discuss the problem of emplacement, the dimensions of the batholith as a whole must be considered. It is no easy task to place dimensions on the batholith because there have been insufficient detailed investigations to settle the question and opinions of prominent authorities are in variance.

Billingsley (1915, p. 33) regarded the Boulder batholith as a much larger body than the outcrop area from Butte to Helena. He considered the batholith continuous from the Phillipsburg batholith, about 30 mi west, to the east margin of the Boulder batholith. Pardee and Schrader (1933, p. 4) point out that there are many similar intrusives still farther to the west which are virtually continuous into the Idaho batholith. (See Figure 104.) Knopf (1913) considered that the Boulder batholith was merely an outlier of the Idaho batholith. Ross (1938) on the other hand, considered that the Idaho batholith is older than the Boulder batholith and contains different rocks.

From the chemical data presented, it appears probable that the petrogenetically related intrusive rocks occupy an area many times the outcrop area of the exposed Boulder batholith. Likewise from the chemical data, an intimate link between the Boulder batholith and the rocks exposed in the Idaho batholith is not indicated.

Magmatic Stopping. The prevailing opinion among geologists appears to be that the Boulder batholith was emplaced by some stopping mechanism, although no one has spelled out the conditions. Billingsley (1915, p. 33)

and Barrell (1907, p. 21) considered that the Boulder batholith is a dome-shaped mass that made room for itself by stoping, with possibly some elevating of the roof.

Lawson (1914) proposed that the Boulder batholith is a laccolith, a conclusion with which Cloos (1923, p. 13) was in general agreement. Daly (1933, p. 112) expresses his obvious disagreement with these words: "Even if Lawson's evidence for his view were compelling, Cloos's deduction lacks cogency."

If the areal exposure of the Boulder batholith is considered to be the approximate limits for the batholith, the body, which is 20 to 25 mi wide and over 60 mi long, is represented as having been stoped from "abyssal depths" (Barrell 1921, p. 6). Its upward progress, based on field observations, was halted within one mile of the earth's surface, over the entire area which, just prior to the emplacement of the batholith, was subjected to strong folding and thrust faulting. The intrusion came to within a mile of the surface, did not break through, and did not materially alter the enclosing rocks or roof pendants for more than a few inches to a few hundred feet from the contact.

Two of the problems appear to be: (1) how did the batholith maintain its roof? and (2) why did the stoping stop at or near the base of the volcanics across the entire region under such a uniformly flat roof? If assimilation is significant, as has been maintained, then the composition of the magma must have been modified by the addition of the sedimentary rocks to the composition of the existing rocks, namely, quartz monzonite. Some stoping and assimilation is indicated by the number of xenoliths found in the batholith, and by the evidence of



their partial to nearly complete (?) digestion. Over large areas of the batholith, it is often impossible to examine an outcrop in detail without finding a xenolith. In some localities near the border of the batholith, the number of xenoliths is great, in other areas they are few in number. In the Elliston district, xenoliths are very scarce. The question is, how much assimilation took place? The chemical data do not support wholesale assimilation of carbonate rocks such as those which underlie the volcanic pile where no intrusive rocks occur.

It is not difficult to compute the composition of the magma that would be required to assimilate the rocks in its path. The approximate composition of the more than 2 mi of Late PreCambrian, Paleozoic and Mesozoic sedimentary rocks, through which the batholith is supposed to have stoped, is given in Analysis 1, Table 21. This is a generalized composite analysis, based on the relative proportions of the sedimentary rocks found in the region of the batholith. It undoubtedly is not precisely correct, nevertheless the large amount of limestone and dolomite in the sedimentary section is indicated.

It would be theoretically possible to produce a rock of quartz monzonitic composition from an 8 mi thick layer of granitic magma of the composition given in Analysis 3, Table 21. If an acidic granitic magma is required to stope its way to the surface and digest the sediments to produce quartz monzonite in a body 10 mi thick, (involving, of course, thorough mixing) then the problem is only partly dealt with, because the origin of the granitic magma must be explained.

Table 21

	I		II
SiO <sub>2</sub>	39.5		71.2
Al <sub>2</sub> O <sub>3</sub>	10.1		17.5
Fe <sub>2</sub> O <sub>3</sub>	3.2		4.8
MgO	6.1		1.0
CaO	18.2		2.4
Na <sub>2</sub> O	1.2		3.3
K <sub>2</sub> O	1.9		2.6
CO	19.3	negative	3.2

I Chemical composition of sedimentary rocks (Late Pre-Cambrian through Mesozoic. Estimated 2 mi thick).

II Theoretical composition of granitic magma that would be required to assimilate the sedimentary rocks to produce quartz monzonite.



If assimilation took place after the extrusion of the volcanic rocks, then the chemical compositions of the volcanic and batholithic rocks would be different, but the chemical data show them to be strikingly similar the evidence of differentiation (compositional changes) progresses in the same relative chemical proportions for both volcanic and batholithic rocks.

One additional point is quite obvious. The sedimentary rocks through which the magma is presumed to have stopped are of widely divergent types, chemically, whereas only rather narrow mineralogical and chemical limits are found within the main body of quartz monzonite. This seems to suggest to me that whatever assimilation took place was accompanied by very thorough mixing.

Dilation. Throughout the region of the batholith, the intrusive rocks (other than the batholith) are dominantly sills. Sills are abundant in the sedimentary rocks and many occur in the volcanic rocks. The earliest stage of intrusion of granitic rocks are sills of granogabbro. The more or less concordant roof of the batholith suggests the possibility that the main intrusive is likewise a concordant batholith of the type described by Iddings (1914, p. 205). Lawson recognized the significance of the flat roof over 40 years ago (1914). I arrived at the conclusion that the batholith is a concordant sill-like body before I was aware of Lawson's paper. *Bully!*

The intrusion of the batholith was immediately preceded by strong folding and thrust faulting. As shown in Figure 104, the trend lines of the folds and faults swing sharply out and around east of the perimeter

of the outcrop near the batholith. The batholith is at the node of a very strong structural discontinuity in the thrust fault zone at the Rocky Mountain front which extends from north of Glacier Park to eastern Idaho. The batholith is about at the cross roads of the thrust fault zone and the more nearly east-west Osborn fault zone.

On the regional map, the Boulder batholith looks like a small appendage in front of the Idaho batholith. The "push" which preceded the batholith, and which is considered to have been responsible for the emplacement of the batholith, appears to have come from the west. Misch postulates a north Pacific arc with the Idaho batholith at the eastern nose of the arc. The "push" may have originated from this same westerly source which formed the north Pacific arc. (Peter Misch, unpublished A.A.P.G. distinguished lecture.)

A simple dilation mechanism for emplacement of quartz monzonite magma along a possible thrust plane, or at or near the base of the volcanic pile (which may be a shearing-off plane) seems to satisfy many of the field, petrographic and chemical conditions, including:

1. Flat and/or concordant roof over broad area, without break-through.
2. The relatively weak contact metamorphism favors a thin, not strongly superheated intrusive.
3. A minimum amount of kinetic energy is required for emplacement of sill.
4. The chemical evidences support close chemical consanguinity of volcanic and intrusive rocks. (Negative evidence for assimilation.)
5. The regional structural picture is explained.



If the batholith has a floor, then the batholith is a very thick composite sill, probably no more than a mile or two in thickness at most localities. The quartz monzonitic magma, moving along a thrust fault, may simply have elevated the roof over a thrust or the plane at the base of the volcanic pile. The phenomenon is considered to be similar to the squeezing of toothpaste from a tube, or the forcing of grout into a foundation. The minimum amount of intrusive kinetic energy, namely the overcoming of the lithostatic head, is all that is required.

Granitization. The absence of regional metamorphic zoning surrounding the granitic rocks, the absence of gneissose structures generally, and the static nature of the contact metamorphism, what there is of it, do not support a granitized origin. Both Goodspeed and Misch, ardent proponents of granitization, consider that the bulk of the Boulder batholith is characterized by igneous textures, and that it is an orthomagmatic intrusive. (Personal communications.)

#### Origin of the Regional Fracture Pattern

The vein pattern within the Boulder batholith has been discussed and considered to be tectonic, produced by very mild deformation which took place during and following the late consolidation stages of the batholith. If the batholith is more nearly a sheet-like body than a plug, the all-over pattern of fractures in the batholith and enclosing rocks could be produced by a couple such as was demonstrated by Mead (1920, p. 50). Many years ago Mead obtained a fracture pattern by a laboratory demonstration which appeared at that time to fit the pattern

in the Butte mining district, and which now appears to fit the pattern throughout the batholith.

The orientation of the couple which is thought to have produced the structures is shown in Figure 111. The nearly east-west veins represent the tensional direction as evidenced by the wide vein zones and the widening of fractures in the more nearly east-west direction of some of the veins. The northeast faults probably represent one of the shear directions, and the northwest fracture zone, the other shear direction. The latter zone is not well developed in the Elliston district and is the least important fracture system in Butte which is in the southern part of the batholith.

#### Mineralization

In the section on the crystallization of the quartz monzonite, it is shown that the sequence of mineral formation in vein-like structures appear to be controlled by the timing of fracturing. The fracturing is clearly related to late crystallization stages of the batholith, rather than to a separate, much younger deformation. In the section on the mineral deposits, the local derivation of the ores is suggested. It has been partly demonstrated that the quartz-pyrite-tourmaline bearing veins are derived from the crystallizing quartz monzonite, and the local derivation is a strong possibility for the base metals as well.

If the batholith is a sheet-like body instead of a plug, the amount of crystallizing magma under the frozen crust of the batholith would be relatively small, consequently there would be no great body



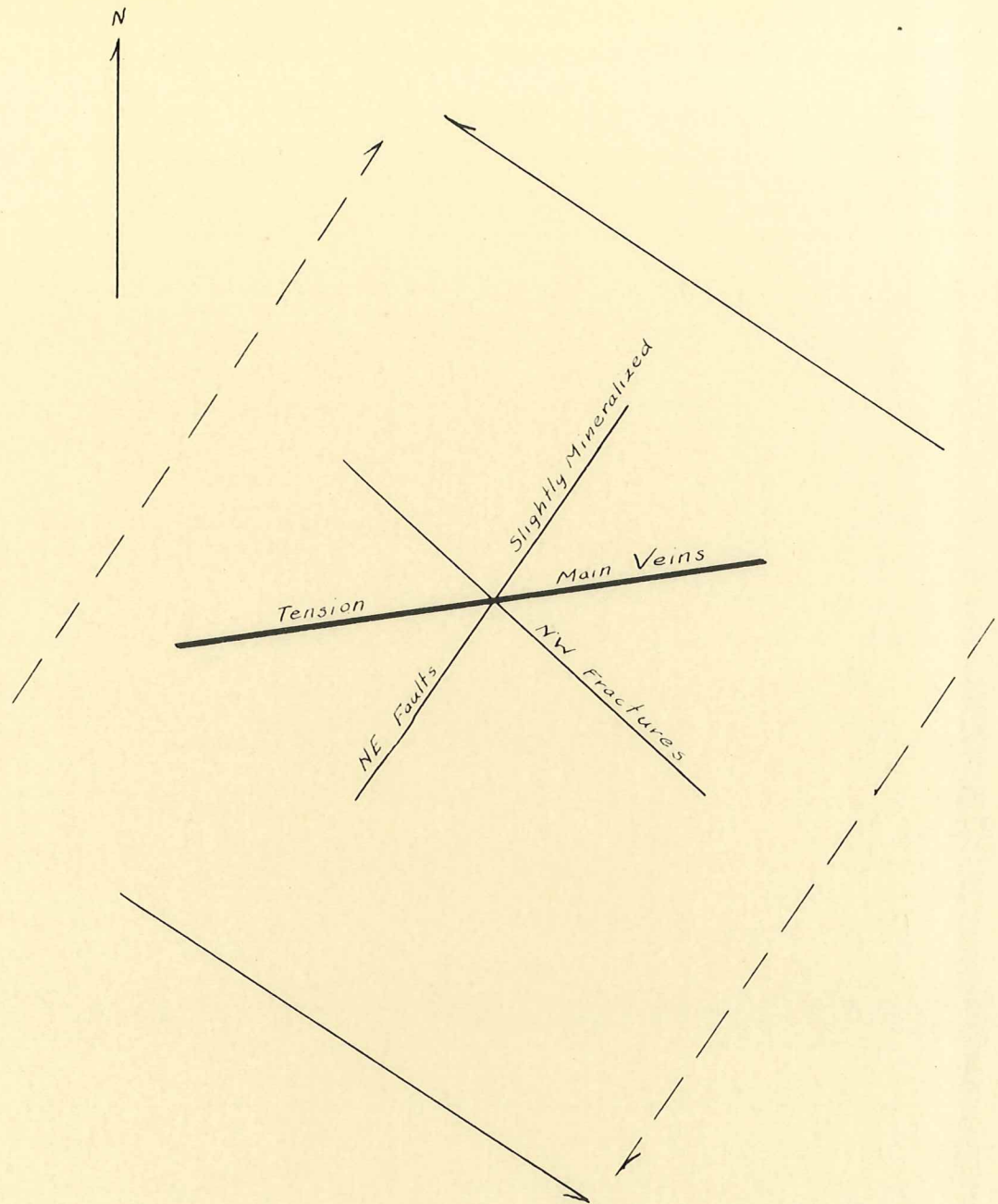


Figure 111. Orientation of Couple which may have acted upon the region into which the Boulder batholith was intruded and which may have produced the mineralized structures.

of magma from which to obtain vast quantities of the ore minerals. Thus the ore minerals might be expected to be spotty in distribution. The uniform mineralization of the veins over a large area, the localization of small ore shoots of the same base metals in the veins throughout the area, may be partly explained by the lack of a large source reservoir.

At Butte, however, the situation is entirely different. The mines at Butte are nearly a mile into the batholith, and are undiminished in valuable mineralization. The district is strongly zoned with respect to a central area. It is a type of mineralization and zoning which is foreign to the rest of the batholith. Butte may well be over the keel or throat of the batholith, and as such, may have derived ore solutions from a much greater thickness of magma, or through a much more direct tap of ore-forming solutions of deeper origin. Butte is a distinct mineralogical anomaly in many respects, and does not fit the picture of the mineralization found throughout the northern part of the batholith. Yet the structural pattern and wall-rock alteration are the same as those found throughout the batholith.

#### Summary

As a result of this investigation, and of observations made in the area of the Boulder batholith, several tentative conclusions have been reached which are not in strict conformity to some previously published conclusions. The conclusions are tentative because there are distinct and large gaps in the knowledge of the history of the batholith. The picture which appears to me to best fit the evidence presently available is briefly this:



1. The volcanic pile formed along some structural zone which is not now clearly indicated, however its position suggests control by the intersection of the thrust fault zone along the Rocky Mountain front and the Osborn fault zone.
2. As the pile grew, the lithostatic head increased until the succeeding volcanic surges were unable to rise to the surface. The thickness of the pile was from 5,000 to possibly 10,000 feet or more.
3. Further magmatic activity was represented by the emplacement of many basic sills in weak horizons within and below the volcanic pile. Basaltic magma, because of its high fluidity, was physically suited for emplacement as sills. Some basalt flows were extruded onto the surface and are the youngest rocks in the volcanic pile.
4. Freezing of the sills and the channels through which the volcanics had been vented, possibly at a dormant period, served to plug the previously established avenues of eruption. The bulk of the pile itself served to balance the forces which caused the outpourings in the first place.
5. Continuation or renewal of stresses (presumably from the west) were offset by a greatly strengthened and somewhat thickened crust. A considerable amount of energy was stored in the build up of these stresses.
6. When the external stresses overcame the resistance of the crust, strong deformation resulted. First the crust yielded by folding, then by overfolding and then by thrust faulting.

7. When the thrust faults formed, (possibly shearing off part of the volcanic pile) the magma, a potentially dilatant mass (Mead 1925, p. 697) moved along the thrust fault or faults. No great stress was applied to the magma after the thrust faulting. Rather the release of pressure may have caused the magma to "liquify" and move into and along a weak zone (either the plane of a thrust or the base of the volcanic pile).
8. The magma elevated the roof of the batholith by very slight warping and by elevation along intrusive faults. It was emplaced at several stages, but the main stage of quartz monzonite came in response to a minimum amount of energy, sufficient only to overcome the lithostatic head. The contact metamorphism was normally weak.
9. Slight movement involving local disturbance of the roof of the batholith after partial crystallization caused a streaming of late magmatic solutions and/or gases (?) towards the roof of the batholith. Aplite and alaskite bodies formed at this stage. Some of the aplites formed by dilation in relatively stronger zones of the batholith; some alaskite formed by metamomatic replacement of the early-formed quartz monzonite.
10. As the magma crystallized to the point where fractures could be maintained, mild tectonic forces acting on the body, produced a uniform, batholith-wide, fracture pattern of the type found in thin sheets. (Mead 1920)
11. The vein matter (quartz, tourmaline and pyrite) was derived, in large part from the walls of the fractures. As re-fracturing took place,



solutions liberated from the incompletely crystallized walls at somewhat greater depths, circulated into the veins and migrated towards the roof. Whereas the first fractures are marked by replacement of the walls, the succeeding fractures are more and more clearly accompanied by fissure fillings. (The significance of successive depth zones for solutions was first spelled out by Klepper in personal communication.) The wall rock alteration was mostly accomplished along the veins prior to the base metal stage of crystallization, and as a consequence of normal, latest stage magmatic (hydrothermal) crystallization.

12. The base metals were derived very late and represent the concentrations of the latest residual magmatic solutions collected from the quartz monzonite magma. As successively deeper and deeper segments of the batholith crystallized, successive "spurts" of base metals were introduced into fractures which were controlled by the large, early-formed, vein structures.

(I think that Klepper is responsible for clarifying my ideas about the multiple base metal injections and the derivation of solutions from successively deeper crystallizing parts of the magma.)

13. At a much later date, renewed magmatic activity produced minor amounts of rhyolite and dacite in and adjacent to the area of the batholith.
14. The youngest rhyolite may not be related to the batholith at all and may have been derived from a foreign source.

## CONCLUSIONS

The petrographic observations of volcanic and intrusive rocks and the petrologic interpretations which have a bearing on the petrogenesis of the igneous rocks, are summarized. Successive stages of vein formation have been recognized and the vein and wall rock mineralogy and paragenesis has been determined. Tentative conclusions have been reached on the petrogenesis of the part of the Boulder batholith exposed in the district and the derivation of the mineral deposits.

## Upper Cretaceous Volcanic Rocks

1. Although the volcanic rocks were not mapped in detail, the general composition and thickness of the volcanic pile in this district is determined. The volcanic rocks are dominantly andesitic and latitic tuffs, lapilli tuffs and tuff breccias. Basalt flows are the youngest rocks in the volcanic pile. The thickness of the composite section of volcanics is about 5,000 ft.
2. Andesitic and latitic welded tuffs have been recognized.
3. The age of the basalt flows is uncertain, but they are tentatively placed with the pre-batholithic volcanic rocks.

## Intrusive Rocks of the Boulder Batholith

1. Two magmatic intrusive units of the Boulder batholith occur in the district, (a) a previously unmapped and distinctive type of granodiorite herein called the Bison Mountain granodiorite, and (b) the main stage or normal quartz monzonite.



2. The quartz monzonite exhibits a distinctive sequence of crystallization stages from early magmatic through late magmatic or deuteritic and hydrothermal.

(a) The early magmatic stage is recognized by the crystallization of andesine, olivine (rare), pyroxene, hornblende and biotite in a normal magmatic sequence.

(b) The late magmatic or deuteritic stage is characterized by a distinctive mineral assemblage and the formation of aplite-alaskite. Crystallization of potash feldspars and quartz and alteration of early formed plagioclase define the late magmatic or deuteritic stage.

(i) The potash feldspars include a series of minerals including orthoclase, anorthoclase and orthoclase microperthite. Orthoclase crystallizes in xenomorphic grains in the groundmass of quartz monzonite, in a succession of porphyroblastic growths from incipient crystalloblastic grains to well formed porphyroblasts. Anorthoclase is a common intermediate step in the replacement of plagioclase by orthoclase.

Orthoclase microperthite is a distinctive type, herein called reorganization perthite. It may have formed by reorganization of albite liberated in the replacement of andesine, first by anorthoclase and then by orthoclase, at a relatively low temperature.

(ii) Quartz occurs as xenomorphic grains in the groundmass and in granophyric intergrowths with potash and rarely plagioclase feldspars.

(iii) Andesine which crystallized in the early magmatic stage is altered to clinozoisite, albite and flecks of sericite in the deuteritic stage.

(c) The hydrothermal stage is marked by feldspar destruction to sericite, with or without structural control. Alteration of feldspar to sericite is characteristic along the quartz ± tourmaline ± pyrite veins. A nearly complete sequence of tourmaline and pyrite segregations without much quartz and with potash feldspar-rich selvage zone in pods and vein-like structures has been recognized in the late deuteritic stage. The alteration of feldspar to sericite, with the accompanying liberation of silica, is characteristic of quartz + tourmaline + pyrite veins in the early hydrothermal stage.

In addition, a complete sequence of crystallization events from early magmatic through the deuteritic and into the early hydrothermal stage, without structural control, has been recognized.

The base and precious metal minerals are thought to represent a somewhat younger hydrothermal stage. Sericite alteration persisted throughout the hydrothermal stage.

3. Volcanic rocks adjacent to the contact of the intrusive quartz monzonite were changed to incipient hornfelses above normal quartz monzonite and to microaplite above quartz monzonite which exhibits strong endomorphic alteration, especially over alaskite.
4. The roof of the intrusive rocks dips gently to the west. At two localities along the western margin, the intrusive rocks appear to lens out into sill-like bodies.



### Mineral Deposits

1. The mineralized veins in quartz monzonite strike N. 65° E., and dip steeply south. The veins formed by a succession of fractures accompanied by mineralization, first replacement lodes of quartz ± tourmaline ± pyrite, followed by successive stages of fissure fillings by the same minerals and later by base metal-bearing minerals and quartz. The paragenetic sequence of the important vein minerals is: tourmaline, quartz, pyrite, arsenopyrite, sphalerite, galena, tetrahedrite, boulangerite, chalcopyrite and ankeritic carbonates.
2. Wall rock alteration consists of successive zones from the vein outward: (a) sericite zone with quartz and pyrite, and (b) argillic zone in which kaolinitic alteration of plagioclase is next to the sericite zone and montmorillonitic alteration of plagioclase next to fresh rock. Alteration of biotite to shreds of chlorite occurs at the perimeter of the wall rock alteration zone.
3. The mineral deposits in the volcanic rocks are intimately related to the batholithic rocks. Deposits occur in narrow fissure veins of three orientations: (1) veins with the same trend as veins in the batholithic rocks; (2) veins parallel to the main northeast faults, and (3) "flat" veins. Except for tourmaline, the essential mineralogy is the same as in the batholithic rocks.
4. Throughout the district, gold occurs with arsenopyrite and silver with galena and tetrahedrite.

### Petrogenesis

1. The Boulder batholith is a composite magmatic intrusive body which, on the basis of mineralogical, petrographical and chemical evidence, is closely related to the Upper Cretaceous volcanic rocks. The physical and optical characteristics of plagioclase feldspars, hornblende and biotite are strikingly similar in intrusive and extrusive rocks, the deuteric alteration of feldspars is the same, and chemical variation diagrams suggest close chemical consanguinity between the volcanic and intrusive rocks.
2. The outpouring of the volcanic rocks is considered to be but the first event in the magmatic epoch in which the Boulder batholith formed.
3. The batholith is considered to be a concordant body, emplaced mainly by dilation without a considerable amount of stoping and assimilation.
4. The structural pattern of veins and faults was produced by tectonic forces (possibly a couple) during and following crystallization of the batholith.
5. The mineralization in the veins is considered to be intimately related to the batholith and to have been derived from latest magmatic (hydrothermal) solutions. The wall rock alteration may have been formed by potash migration towards the vein from the walls at a temperature lower than that required for stable feldspars. Silica released in the alteration of feldspar to sericite contributed to vein quartz.



## References

- Aiken, Wilbur, 1950, The Lilly mine of Powell County, Montana: B. S. Thesis, Montana School of Mines.
- Alden, W. C., 1953, Physiography and glacial geology of western Montana and adjacent areas: U. S. Geol. Survey Prof. Paper 231.
- Alling, H. L., 1921, The mineralogy of the feldspars: Jour. Geology v. 29, p. 193-294
- Anderson, A. L., 1952, Multiple emplacement of the Idaho batholith: Jour. Geology v. 60, p. 255-265.
- Anderson, Olaf, 1928, The genesis of some types of feldspar from granite pegmatites, Norsk geol. tidsskr. v. 10, p. 148.
- Baier, Ernst, 1930, Lammellenbau u. Entmischung Struktur der Feldspäte: Zeitschr. Kristallographie v. 73, p. 465.
- Balk, Robert, 1937, Structural behavior of igneous rocks: Geol. Soc. America Memoir 5.
- Barksdale, J. D., 1951, Cretaceous glassy welded tuffs--Lewis and Clark County, Montana: Am. Jour. Science v. 249, p. 439-443.
- Barrell, Joseph, 1907, Geology of the Marysville mining district, Montana: U. S. Geol. Survey Prof. Paper 57.
- Barth, T. F. W., 1931, Mineralogical petrography of Pacific lavas: Am. Jour. Sci., 5th ser. v. 221, p. 377-405, 491-530.
- Bastin, E. S., et. al., 1931, Criteria of age relations of minerals: Econ. Geology v. 26, p. 561-610.
- Becraft, G. E., 1955, New field classification of the quartz monzonite and granodiorite of the Boulder batholith, Montana: Geol. Soc. Am. Bull. v. 66, p. 1642.
- Beliankin, D., 1927, Anorthoclase: Acad. sci. U.R.S.S. Bull. p. 1115-1124.
- Berry, G. W., 1943, Stratigraphy and structure at Three Forks, Montana: Geol. Soc. Am. Bull. v. 54, p. 1-30.
- Billingsley, Paul, 1916, The Boulder batholith of Montana: Am. Inst. Min. Met. Eng. Trans. v. 51, p.
- and Grimes, J. A., 1918, Ore deposits of the Boulder batholith of Montana: Am. Inst. Min. Met. Eng. Trans. v. 58, p. 284-368.



- Bowen, N. L., 1928, The evolution of igneous rocks: Princeton University Press.
- and Tuttle, O. F., The system  $\text{NaAlSi}_3\text{O}_8 - \text{KAlSi}_3\text{O}_8 - \text{H}_2\text{O}$ : Jour. Geology v. 58, p. 489-511.
- Brammhall, Alfred, and Harwood, H. F., 1927, On the temperature range of formation of tourmaline in Dartmoor granite: Mineralog. Mag. v. 21, p. 205.
- 1932, The Dartmoor granites, their genetic relationships: Geol. Soc. London Quart. Jour. v. 88, p. 171-237.
- Buddington, A. F., 1939, Adirondack igneous rocks and their metamorphism: Geol. Soc. America Memoir 7.
- , 1933, Correlation of kinds of igneous rocks with kinds of mineralization: Am. Inst. Min. Met. Eng. Ore Deposits of the Western States, Chapter 8, p. 350-385.
- Buerger, M. J., 1934, Unmixing of chalcopyrite from sphalerite; Econ. Geology v. 19, p. 525.
- Butler, B. S., 1913, Geology and ore deposits of the San Francisco and adjacent districts, Utah: U.S. Geol. Survey Prof. Paper 80.
- Camsell, Charles, 1910, Geology and ore deposits of Hedley mining district, British Columbia: Canada Geol. Survey Memoir 2, p. 100.
- Chaisson, Ursula, 1950, The optics of triclinic adularia: Jour. Geology v. 58, p. 537-547.
- Chayes, Felix, 1950, On the relation between anorthite content and  $\gamma$ -index of natural plagioclase: Jour. Geology v. 58, p. 593-595.
- , 1952a, On the association of perthitic microcline with highly undulant or granular quartz in some calcalkaline granites: Am. Jour. Sci., 5th ser. v. 250, p. 281-296.
- 1952b, Notes on the staining of potash feldspar with sodium cobaltinitrite in thin section: Am. Mineralogist v. 37, p. 337-340.
- 1954, A test of the revised determinative chart for plagioclase: Am. Jour. Sci., 5th ser. v. 252, p. 172-180.
- 1955, Potash feldspar as a by-product of the biotite-chlorite transformation: Jour. Geology v. 63, p. 75-81.
- Chao, S. H., Smare, D. L., and Taylon, W. H., 1939, An X-ray examination of some soda-potash feldspars: Mineralog. Mag. v. 25, p. 338-350.



- Emmons, W. H., 1933, On the mechanism of the deposition of certain metalliferous lode systems associated with granitic batholiths: Chapt. 7 in ore deposits of the western states, Am. Inst. Min. Met. Eng. O. 327-349.
- Faessler, Carl, and Tremblay, L. P., 1946, Perthite as age indicator in Laurentian gneiss and Pine Hill intrusives: Canadian Inst. Min. Met. Trans. v. 405, p. 58-70.
- Fenneman, N. M., 1931, Physiography of western United States: McGraw-Hill Book Co., New York.
- Fenner, C. N., 1948, Incandescent tuff flows in southern Peru: Geol. Soc. America Bull. v. 59, p. 379-893.
- Franco, R. R., and Schairer, J. F., 1951, Liquidus temperatures in mixture of the feldspars of soda, potash and lime; Jour. Geology v. 59, p. 259-267.
- Friedman, Irving, 1951, Some aspects of the system  $H_2O - Na_2O - SiO_2 - Al_2O_3$ : Jour. Geology v. 59, p. 19-31.
- Fritsch, K. von, and Reiss, W., 1868, Geol. Beschrieb. der Insel Teerife, Teneriffe, (Canary Islands). (See Mansfield and Ross.)
- Fuller, R. E., 1938, Deuteric alteration controlled by the jointing of lavas, Am. Jour. Sci. 5th ser. v. 235, p. 161-171.
- Gabriel, Akton, and Cox, E. P., 1929, A staining method for quantitative determination of certain rock minerals: Am. Mineralogist v. 14, p. 290-292.
- Gardiner, C. I., and Reynolds, S. H., 1931, The Loch Doon 'granite' area, Galloway: Geol. Soc. London Quart. Jour. v. 88, p. 20.
- Gates, R. M., 1953, Petrogenetic significance of perthite: Chapter 5 in Selected petrogenic relationships of plagioclase feldspar: Geol. Soc. America Memoir 52, p. 55-69.
- Gilbert, G., 1924, The relation of hardness to the sequence of ore minerals: Econ. Geology v. 19, p. 668.
- Gilbert, C. M., 1938, Welded tuff in eastern California: Geol. Soc. America Bull. v. 49, p. 1829-1862.
- Goldich, S. S., and Kinser, J. H., 1939, Perthite from Tory Hill, Ontario: Am. Mineralogist v. 22, p. 410-415.
- Goldsmith, J. R., 1950, Gallium and germanium substitutions in synthetic feldspars: Jour. Geology v. 58, p. 518-536.
- , 1952, Diffusion in plagioclase feldspars: Jour. Geology v. 60, p. 288-291.

Goodspeed, G. E., and Fuller, R. E., 1944, Replacement aplite breccia: Jour. Geology v. 52, p. 264-274.

-----, 1952, Replacement and rheomorphic dikes: Jour. Geology v. 60, p. 356-363.

Harker, Alfred, 1909, The natural history of the igneous rocks:

Higazy, R. A., 1949, Petrogenesis of perthite pegmatites in the Black Hills, South Dakota: Jour. Geology v. 57, p. 555-581.

Hintze, Carl, 1892, Handbuch der Mineralogie: p. 1414, (See Alling 1921).

Horberg, Leland, 1954, Rocky Mountain and continental Pleistocene deposits in the Waterton region, Alberta, Canada: Geol. Soc. Am. Bull. v. 65, p. 1093-1150.

Huber, O. von, 1901, Jarb. k. k. Geol. Reichsanst. vol. 50, p. 101. (See Daly 1933.)

Hulin, C. D., 1929, Structural control of ore deposition: Econ. Geology v. 24, p. 38-40.

Hunt, C. B., Averitt, Paul, and Miller, R. A., 1953, Geology and geography of the Henry Mountains region, Utah: U. S. Geol. Sur. p. 228.

Iddings, J. P., 1914, The problems of Volcanism: Yale University Press, New Haven.

Johannsen, Albert, 1916, translation of Weinschenk, Ernest, The fundamental principles of petrology: McGraw Hill Book Co., New York. Footnote p. 80 relative to Grush (sand size and grus (dust)).

-----, 1939, Petrography of the igneous rocks: v. 1, The University of Chicago Press, Chicago.

-----, 1949, A descriptive petrography of the igneous rocks: v. 2, the quartz-bearing rocks: The University of Chicago Press, Chicago.

Johns, W. M., 1952, Geology of the Third Term mine, Elliston mining district, Powell County, Montana: B. S. Thesis, Montana School of Mines.

Keith, N. L., 1939, Selective staining to facilitate Rosiwal analysis: Am. Mineralogist v. 24, p. 561-565.



- Kesler, T. L., and Olson, J. C., 1942, Muscovite in the Spruce Pine district, North Carolina: U. S. Geol. Sur. Bull. 936A.
- Klepper, M. R., 1951, Geology of the southern Elkhorn Mountains, Jefferson and Broadwater Counties, Montana: U. S. Geol. Sur. Open File Report, Montana Open File No. 19, Spokane.
- , and Robertson, Forbes, 1956, Late magmatic phenomena in the northern part of the Boulder batholith, Montana: Econ. Geology v. 51, p. 117.
- Knopf, Adolph, 1913, Ore deposits of the Helena mining region, Montana: U. S. Geol. Sur. Bull. 527.
- , 1913a, The tourmalinic silver-lead type of ore deposit: Econ. Geology v. 8, p. 105-119.
- , 1914, Is the Boulder batholith a laccolith? Discussion of paper by A. C. Lawson: Econ. Geology v. 9, p. 396-402.
- , 1955, Batholiths in time: in the crust of the earth: Geol. Soc. America Special Paper 62, p. 685-702.
- Köhler, Alexander, 1948, Zur optik des adulars: Neues. Jahrb. sect. A, No. 5-8, p. 49-55.
- 1949, Recent results of investigation on the feldspars: Jour. Geol. v. 57, p. 592-599.
- Lacroix, A., 1940, La Montagne Pelée et ses éruptions: Paris.
- Larsen, E. S., Jr., 1938, Some new variation diagrams for groups of igneous rock: Jour. Geology v. 46, p. 506-520.
- Larsen, E. S., and Bure, B. F., 1938, Potash analcime and pseudoleucite from the Highwood Mountains of Montana: Am. Mineralogist v. 23, p. 837-849.
- Larsen, E. S., 1940, Petrogenic province of Central Montana: Geol. Soc. America Bull. v. 51, p. 887-946.
- , 1948, Batholith and associated rocks of Corena, Elsinore, and San Luis Rey quadrangles, Southern California: Geol. Soc. America Memoir 29.
- , 1938, Some new variation diagrams for groups of igneous rocks: Jour. Geology v. 46, p. 506-520.
- Laskey, S. G., 1947, Geology and ore deposits of the Little Hatchet Mountains, Hidalgo and Grant Counties, New Mexico: U. S. Geol. Sur. p. p. 208.

- Laves, F., and Chaisson, U., 1950, An X-ray investigation of the high-low albite relations- Jour. Geology v. 58, p. 584-592.
- Laves, Fritz, 1950, The lattice and twinning of microcline and other potash feldspars: Jour. Geology v. 58, p. 548-571.
- , 1951, Artificial preparation of microcline: Jour. Geology v. 59, p. 511-521.
- , 1952, Phase relations of the Alkali feldspars, I Introductory remarks; II The stable and pseudo-stable phase relations in the alkali feldspar system: Jour. Geology v. 60, p. 436-450, 549-574.
- , and Goldsmith, J. R., 1954, On the use of calcic plagioclase in geologic thermometry: Jour. Geology v. 62, p. 405-408.
- , 1954, The coexistence of two plagioclases in the oligoclase composition range: Jour. Geology v. 62, p. 409-411.
- Lawson, A. C., 1914, Is the Boulder batholith a laccolith?: California Univ., Dept. Geol. Dept. Sci., Bull. v. 8, p. 1-15.
- Lindgren, Waldemar, 1933, Differentiation and ore deposition, Cordilleran region of the United States: Chapt. 4 in Ore Deposits of the Western States: Am. Inst. Min. Met. Eng. p. 152-180.
- , 1937, Succession of minerals and temperatures of formation in ore deposits of magmatic affiliation: Am. Inst. Min. Met. Eng. Trans. v. 162, p. 356-376.
- Loughlin, G. F., 1912, The gabbros and associated rocks at Preston, Conn: U. S. Geol. Soc. Bull. 492. (Alaskite p. 123-124.)
- Loughlin, G. F., and Koschmann, A. H., 1942, Geology and ore deposits of the Magdalena mining district, New Mexico: U. S. Geol. Sur. p. 200.
- Lovering, T. S., and Goddard, E. N., 1950, Geology and ore deposits of the Front Range, Colorado: U. S. Geol. Sur. pp 223.
- Lovering, T. S., and Tweto, Ogden, 1953, Geology and ore deposits of the Boulder County tungsten district, Colorado: U. S. Geol. Sur. Prof. Paper 245.
- Luczizky, W., 1905, Tsch. Min. Pet. Min. v. 24, p. 347, (isomicrocline).
- Lyons, J. B., 1944, Igneous rocks of the northern Big Belt range, Montana: Geol. Soc. America Bull. v. 55, p. 445-472.



- Mackenzie, W. S., and Smith, J. V., 1955, Alkali feldspars I, orthoclase-micropertthites: *Am. Mineralogist* v. 40, p. 707-732.
- Mackin, J. H., Cooke, E. F., and Threet, R. L., 1954, Stratigraphy of early Tertiary volcanic rocks in southwestern Utah: *Geol. Soc. America Bull.* v. 65, p. 1280 (abs.).
- Mackin, J. H., 1947, Some structural features of the intrusion in the Iron Springs district (Utah): *Utah Geol. Sur. Guidebook No. 2.*
- Mackin, J. H., and Nelson, W. H., 1950, Early tertiary welded tuffs in the Iron Springs district, Utah: *Geol. Soc. America Bull.* v. 61, p. 1528 (abs.).
- Mäkinen, Eero, 1913, Die Pegmatite von Tammela in Finland und ihre Minerale: *Comm. Geol. Finlande Bull.* v. 35.
- Mansfield, G. R., and Ross, C. S., 1935, Welded rhyolitic tuffs in southeastern Idaho: *Tran. Am. Geophysical Union* v. 16, p. 308-321.
- Marshall, P., 1935, Acid rocks of the Taupo-Rotorua volcanic district, (New Zealand): *Royal Soc. New Zealand Trans.* v. 64, p. 44.
- Mead, W. J., 1920, Notes on the mechanics of geologic structures: *Jour. Geology* v. 28, p. 505-523.
- , 1925, The geologic role of dilatancy: *Jour. Geology* v. 33, p. 685-698.
- Miller, W. J., 1918, *Geology of the Lake Placid quadrangle*: *New York State Mus. Bull.* v. 211, 212.
- Moody, J. D., and Hill, M. J., 1955, *Wrench faults tectonics*: Unpublished manuscript.
- Morey, G. W. and Ignerson, Earle, 1937, The pneumatolitic and hydrothermal alteration and synthesis of silicates: *Econ. Geol.* v. 32, p. 607-761.
- Murchison, R., 1946, Kritische Temperatur des Wassers and der wasserigen Lösungen des kiesel-sauren Natriums: *Zentralb. Min. Abt. A.*, p. 46.
- Murray-Hughes, Robert, with petrographic notes by Fitch, A. A., 1929, The geology of part of North-Western Rhodesia: *Geol. Soc. London Quart. Jour.* v. 85, p. 109-162.
- Newhouse, W. H., 1938, The microscopic criteria of replacement in opaque ore minerals: In *Labratory Investigation of Ores by Fairbanks.*

- Newhouse, W. H., 1928, The time sequence of hypogene ore mineral deposition: *Econ. Geol.* v. 23, p. 647.
- , 1940, Openings due to movement along a curved or irregular fault plane: *Econ. Geol.* v. 35, p. 445-464.
- , 1942, Ore deposits as related to structural features: Princeton University Press, New Jersey.
- Niggli, Paul, and Morey, G. W., 1913, Die hydrothermale Silikatbildung: *Zeitschr. f. anorg. Chem.*, v. 83, p. 369.
- Nockolds, S. P., 1940, The Garabal Hill-Glen Fyne igneous complex: *Geol. Soc. London Quart. Jour.* v. 96, p. 451-508.
- Nolan, T. B., 1935, The Gold Hill mining district, Utah: *U. S. Geol. Sur. prof. paper* 177.
- Oftedahl, C., 1948, Studies in the igneous rock complex of the Oslo region IX The Feldspars: *Norske vidensk. Akad. Oslo Skr. I, Math. naturu. Kl. 3*, p. 1-71.
- Osborn, E. F., 1949, Cellular structures in glass as related to structures in lava: *Jour. Geology* v. 57, p. 73-82.
- Osman, C. W., 1947, The granites of the Scilly Isles and their relation to the Dartmoor granites: *Geol. Soc. London Quart. Jour.* v. 84, p. 259.
- Osterwald, F. W., 1955, Petrology of the pre-Cambrian granites in the northern Bighorn Mountains, Wyoming: *Jour. Geology* v. 65, p. 310-327.
- Page, L. R. et. al., 1954, Pegmatite investigations 1942-1945, Black Hills, South Dakota: *U. S. Geol. Sur. Prof. Paper* 247.
- Paine, W. R., 1952, Stratigraphy of the Phosphoria formation in Montana: *M. S. Thesis*, Montana School of Mines.
- Pardee, J. T., The garrison and Philipsburg phosphate fields, Montana: *U. S. Geol. Sur. Bull.* 640.
- , and Schrader, F. C., 1933, Metalliferous deposits of the greater Helena mining region, Montana: *U. S. Geol. Sur. Bull.* 842.
- Pardee, J. T., 1950, Late Cenozoic block faulting in western Montana: *Geol. Soc. Am. Bull.* v. 61, p. 359-406.
- Park, C. F., Jr., and Carmon, R. S., Jr., 1943, Geology and ore deposits of the Metaline quadrangle, Washington: *U. S. Geol. Sur. Prof. Paper* 202.



- Parkinson, John, 1907, The geology of the Oban Hills (So. Nigeria): Geol. Soc. London Quart. Jour. v. 63, p. 313.
- Parsons, W. H., 1942, Origin and structure of the Livingston igneous rocks, Montana: Geol. Soc. America Bull. v. 52, p. 1175-1186.
- Perry, E. S., 1945, Distribution of sedimentary rocks in Montana: Montana Bur. Mines and Geol. Misc. Cont. 8.
- , Perry, E. S., 1953, Oil and gas in Montana: Mont. Bur. Geol. and Mines Mem. 35.
- Pettijohn, F. J., Sedimentary Rocks: Harper and Bros., New York.
- Pemister, J., 1926, The geology of Strath Oykell and lower Lock Shin (Explanation of Sheet 102): Scotland Geol. Sur. Memoir.
- Potgieter, C. T., 1950, The structure and petrology of the George granite plutons and the invaded pre-Cape sedimentary rocks: Univ. Stellenbosch Annals v. 26, p. 381.
- Ramberg, Hans, 1952, Chemical bonds and distribution of cations in silicates: Jour. Geology v. 60, p. 331-335.
- Regnier, Jerome, 1951, Mineralogy and paragenesis of the eastern part of the Elliston district, Montana: MS Thesis, Montana School of Mines.
- Reynolds, D. L., 1937, The eastern end of the Newry igneous complex: Geol. Soc. London Quart. Jour. v. 90, p. 585-636.
- Reynolds, D. L., 1955, Fluidization as a geological process and its bearing on the problem of intrusive granites: Am. Jour. Sci. v. 252, p. 577-614.
- Richey, J. E., 1927, The structural relations of the Mourne granites (Northern Ireland): Geol. Soc. London Quart. Jour. v. 83, p. 653-688.
- Ridge, John, 1949, Replacement and the equating of volume and weight: Jour. Geology v. 57, p. 522-550.
- Robertson, Forbes, and Regnier, Jerome, 1952, Mineralogy and paragenesis of the ore deposits in the Elliston mining district, Montana: Northwest Sci. v. 26, p. 37-38 (abs.).
- , and Vandever, P. L., 1952, A new diagrammatic scheme for paragenetic relations of the ore minerals: Econ. Geology v. 47, p. 101-105.

- Robertson, Forbes, 1953, Geology and mineral deposits of the Zosell (Emery) mining district, Powell County, Montana: Montana Bur. Mines and Geol. Mem. 34.
- , 1953a, Notes on the chemical petrology of the Boulder batholith and related rocks: Montana Acad. Sci. Proc. v. 13, p. 67-76.
- , 1954, Deuteric differentiation of gabbro at Mt. Sheridan, Wichita Mountains, Oklahoma: Geol. Soc. America Bull. v. 65, p. 1349 (abs.).
- , Pinckney, D. M., and Klepper, M. R., 1956, Notes on sequence of vein formation in the northern part of the Boulder batholith, Montana: Econ. Geology v. 51, p. 124 (abs.).
- Rogers, A. F., and Kerr, R. F., Optical Mineralogy: McGraw-Hill Book Co., New York.
- Rosenblum, S., 1955, Improved technique for staining potash feldspars: Unpublished manuscript U. S. Geol. Surv. Spokane office.
- Rosenbusch, H., 1900, Microscopical physiography of the rock-making minerals; Trans. by Iddings, J. P., John Wiley & Sons., New York.
- Ross, C. P., 1938, Mesozoic and Tertiary granite rocks in Idaho: Jour. Geology v. 36, p. 673-693.
- Ross, C. S., 1935, Origin of the copper deposits of the Ducktown type in southern Appalachian region: U. S. Geol. Surv. Prof. Paper 179.
- Sales, R. H., 1914, Ore deposits at Butte, Montana: Am. Inst. Min. Met. Trans. v. 46, p. 3-106.
- , and Meyer, Charles, 1948, Wall rock alteration at Butte, Montana: Am. Inst. Min. Met. Trans. v. 178, p. 9-35.
- , 1949, Results from preliminary studies of vein formation at Butte, Montana: Econ. Geology v. 44, p. 465-484.
- , 1956, Genetic significance of quartz-molybdenite mineralization in the Butte district, Montana: Econ. Geology v. 51, p. 126.
- Sargent, H. C., 1918, On the spilitic facies of lower Carboniferous lava flows: Geo. Soc. London Quart. Jour. v. 73, p. 19.
- Schairer, J. F., 1950, The alkali-feldspar join in the system  $\text{NaAlSi}_3\text{O}_8$ - $\text{KAlSi}_3\text{O}_8$ - $\text{SiO}_2$ : Jour. Geology v. 58, p. 512-517.
- Schwertner, C., 1934, Structures and textures of synthetic replacements in "open space": Econ. Geology v. 29, p. 611.



- Schwartz, G. M., 1942, Progress in the study of exsolution in ore minerals: *Econ. Geol.* v. 37, p. 415.
- Scrivenor, J. B., 1914, The topaz-bearing rocks of Gunong Bakau (Federated Malay States): *Geol. Soc. London Quart. Jour.* v. 70, p. 363-380.
- Shand, S. J., 1942, Phase petrology on the Cortlandt Complex, New York: *Geol. Soc. America Bull.* v. 53, p. 409-428.
- , 1944, The terminology of late magmatic and post-magmatic processes: *Jour. Geology* v. 52, p. 342-350.
- Sloss, L. L. and Hamblin, R. H., 1942, stratigraphic and insoluble residues of Madison group (Mississippian) of Montana: *Am. Assoc. Pet. Geol. Bull.* 26, p. 305-335.
- , and Laird, W. M., 1947, Devonian system in central and north-west Montana: *Am. Assoc. Pet. Geol. Bull.* 31, p. 1404-1430.
- , 1950, Paleozoic sedimentation in Montana area: *Am. Assoc. Pet. Geol. Bull.* 34, p. 423-451.
- Smith, F. G., 1948, Transport and deposition of the non-sulfide vein minerals III Phase relationships at the pegmatite stage: *Econ Geol.* v. 43, p. 535-546.
- Smyth, C. H., Jr., and Budding, A. F., 1936, Geology of the Lake Bonaparte quadrangle, (New York): *New York State Mus. Bull.* 269. (alaskite.)
- Smith, J. V., and MacKenzie, W. S., 1955, The alkali feldspars II, A simple X-ray technique for the study of alkali feldspars: *Am. Mineralogist* v. 40, p. 733-747.
- Spencer, E., 1937, The potash-soda feldspars, I, Thermal stability: *Mineralog. Mag.* v. 24, p. 453-494.
- , 1938, Heat treatment and changes of optical properties of feldspar: *Mineralog. Mag.* v. 25, p. 450.
- Spurr, J. E., 1900, Classification of igneous rocks according to composition: *American Geologist* v. 25, p. 229-230.
- , 1906, Ore deposits of the Silver Peak quadrangle, Nevada: *U. S. Geol. Survey Prof. Paper* 55.
- Stillwell, F. L., and Edwards, A. B., 1946, On the occurrence of submicroscopic gold in pyrite from the Dolphen East lode, Tavua, Fiji: *Australia Inst. Min. Met. Proc.* v. 141.

- Stoiber, B. E., 1940, Minor elements in sphalerite: *Econ. Geology*, v. 35, p. 301.
- Stone, J. B., and Sandord, F., 1929, Central Montana mining summarized: *Mining Truth*, v. 14, p. 7-24.
- Stone, R. W., and Calvert, W. R., 1910, Stratigraphic relations of the Livingston formation of Montana: *Econ. Geology* v. 5, p. 551-557, 652-669, 741,-764.
- Stone, R. W., and Bonaine, C. A., 1915, The Elliston phosphate field: *U. S. Geol. Survey Bull.* 580, pp. 373-383.
- Taylor, W. H., 1933, The structure of sanidine and other feldspars: *Zeitschr. Kristallographie*, v. 85, p. 425-442.
- Turner, H. W., 1899, Granitic rocks of Sierra Nevada: *Jour. Geology* v. 9, p. 146.
- Tuttle, O. F., and Bowen, N. L., 1950, High temperature albite and contiguous feldspars: *Jour. Geology* v. 58, p. 572-583.
- Tuttle, O. F., 1952, Origin of the contrasting mineralogy of extrusive and plutonic salic rocks: *Jour. Geology* v. 60, p. 107-124.
- Tyrrell, G. W., 1916, Further notes on the Petrography of South Georgia: *Geol. Mag.* v. 3, p. 438.
- Umpleby, J. B., 1917, Geology and ore deposits of the Mackay region, Idaho: *U. S. Geol. Survey Prof. Paper* 97.
- Uspensky, N. M., 1943, On the genesis of granite pegmatites: *Am. Mineralogist*, v. 28, p. 445.
- Verhoogen, J., 1938, Thermodynamical calculations of the solubility of some important sulfides up to 400°C.: *Econ. Geol.* v. 33, p. 776.
- Whay, J. S., 1939, Some features of the Livingston formation near Nye, Montana: *Am. Geophys. Union Trans.* 20th an. mtg. pt. 3, p. 433-437.
- Vogt, J. H. L., 1921, The physical chemistry of the crystallization and magmatic differentiation of igneous rocks: *Jour. Geology* v. 29, p. 318-350, 426-443, 515-539, 627-649.
- , 1931, The physical chemistry of the magmatic differentiation of igneous rocks: *Skr. Norske. Vidensk. Akad. Oslo* No. 3, p. 51.
- Walker, F. W., and Mathias, Morna, 1946, The petrology of two granite-slate contacts at Cape Town, South Africa: *Geol. Soc. London Quart. Jour.* v. 102, p. 499-518.
- Ward, L. K., 1912, The Heemskirk massif, Its structure and relationships: *Australian Assoc. Adv. Sci.* v. 13, p. 165-175. (floored batholith)



- Washington, H. S., 1898, The petrographical province of Essex County, Massachusetts: Jour. Geology v. 6, p. 89,-94.
- , 1917, Chemical analyses of igneous rocks: U. S. Geol. Survey Prof. Paper 99, 1917.
- Weed, W. H., 1893, The Laramie and the overlying Livingston formation of Montana: U. S. Geol. Sur. Bull. 105, p. 10-41.
- , 1899, Granite rocks of Butte, Mont., and vicinity: Jour. Geology, v. 7, p. 737-750.
- , 1901, Geology and ore deposits of the Elkhorn mining district, Jefferson County, Montana: U. S. Geol. Surv. 22nd an. Rept. Part II, p. 407-549.
- , 1912, Geology and ore deposits of the Butte district, Montana: U. S. Geol. Survey Prof. Paper 74.
- Weidman, S., 1907, The geology of north central Wisconsin: Wisconsin Geol. & Nat. Hist. Survey, Sci. Ser. 4, Bull. 16.
- Wells, F. G., 1938, The origin of the iron deposits in the Bull Valley and Iron Springs districts, Utah: Econ. Geology v. 33, p. 481.
- Wells, M. K., 1954, The structure and petrology of the hypersthene gabbro intrusion, Ardnamurchan, Argyllshire: Geol. Soc. London Quart. Jour. v. 109, p. 367-394.
- Wentworth, C. K., and Williams, Howell, 1932, The classification and terminology of pyroclastic rocks: Nat. Research Council Bull. v. 89,
- Whitehouse, M., 1937, The deuteric mineral sequence in the Enogyera granite: Mineralog. Mag. v. 24, p. 538.
- Willbourn, E. S., 1933, The geology of the scheelite mine, Kramat Pulai tin Ltd., Kinta, Federated Malay States: Geol. Soc. London Quart. Jour. v. 89, p. 449-477.
- Williams, J. S., 1948, Mississippian-Pennsylvanian boundary problems in the Rocky Mountains region: Jour. Geol. v. 56, p. 327-351.
- Williamson, W. D., 1935, The composite gneiss and contaminated granodiorite of Glen Shee, Perthshire: Geol. Soc. London Quart. Jour. v. 91, p. 382-419.
- Winchell, A. N., 1925, Studies in the feldspar group: Jour. Geology v. 33, p. 714-727.
- , 1947, Elements of Optical Mineralogy Part II, Description of minerals: John Wiley and Sons, New York.

- Winchell, A. N., 1952, Elements of Optical Mineralogy Part II, Description of Minerals: John Wiley and Sons, New York.
- Worth, R. H., 1920, The geology of the Meldon valleys near Okehampton, on the northern verge of Dartmoor: Geol. Soc. London Quart. Jour. v. 75, p. 77-112.
- Wright, F. E., and Wright, C. W., 1908, The Ketchikan and Wrangell mining district: U. S. Geol. Sur. Bull. 347.
- Wright, C. W., 1915, Geology and ore deposits of the copper Mountain and Kasaan Peninsula, Alaska: U. S. Geol. Survey Prof. Paper 87.
- Ziegler, U., 1914, The differentiation of granitic magma: Jour. Geol. v. 9, p. 264.



## VITA

Forbes Smith Robertson, the son of Percival and Gladys Smith Robertson of Elsah, Illinois, was born in New Haven, Connecticut on May 24, 1915. He graduated from The Principia Upper School, St. Louis, Missouri in 1934. He received the B.A. degree from The Principia College, Elsah, Illinois in 1938 and the M.S. degree from Washington University in St. Louis, Missouri in 1940. Two years of additional post-graduate work was taken at Massachusetts Institute of Technology, Cambridge, Massachusetts.

He worked as geologist for Reynolds Mining Corporation and Reynolds Metals Company and as senior geologist for Standard Oil Company (N.J.) to 1946. He served as economic geologist for Missouri Geological Survey and Water Resources to 1948 at which time he accepted an appointment at Montana School of Mines and Montana Bureau of Mines and Geology where he served as associate professor of geology and geologist to 1955 (On leave 1953-55). In 1953 he entered the Graduate School of University of Washington and also served as lecturer.