Contributions to Economic Geology 1963

GEOLOGICAL SURVEY BULLETIN 1182

This volume was published as separate chapters A–E

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

William T. Pecora, *Director*

CONTENTS

[Letters designate the separately published chapters]

Geology of Precambrian Rocks, Idaho Springs District, Colorado

By ROBERT H. MOENCH

CONTRIBUTIONS TO ECONOMIC GEOLOGY

GEOLOGICAL SURVEY BULLETIN 1182-A

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

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1964

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

The U.S. Geological Survey Library catalog card for this publication appears after index.

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C. 20402

CONTENTS

ILLUSTRATIONS

TABLES

GEOLOGY OF PRECAMBRIAN ROCKS, IDAHO SPRINGS DISTRICT, COLORADO

By Robert H. Moench

ABSTRACT

The Idaho Springs district, an area of Tertiary gold- and silver-rich base-metal sulfide veins in the central part of the Front Range mineral belt, lies within a terrane of Precambrian bedrock. The bedrock consists dominantly of gneissic rocks and sparse granitic and pegmatitic rocks.

The Precambrian bedrock of the Idaho Springs district is composed largely of generally conformable gneissic rocks of three principal types—interlayered biotite gneisses (or biotite gneiss), granite gneiss and pegmatite (or granite gneiss), and microcline-quartz-plagioclase-biotite gneiss (or microcline gneiss). The granite gneiss and the biotite gneiss, which consist mainly of biotite-quartz-plagioclase gneiss and sillimanitic biotite-quartz gneiss, are intimately intercalated and are separated as map units on the basis of the relative abundance of each rock type. The microcline gneiss is a light-colored rock that forms two main layers—the Quartz Hill layer and the Big Five layer. These layers are separated and are overlain and underlain by thick units of biotite gneiss with intermixed granite gneiss. Contacts between units of microcline gneiss and the other major units are generally sharp. Gneissic rocks of lesser abundance include quartz gneiss, amphibolite, calcium-silicate gneiss and garnet-hornblende-orthopyroxene gneiss.

The gneissic rocks, except the granite gneiss and possibly the garnet-hornblende-orthopyroxene gneiss, are thought to represent a metamorphosed succession of clastic sedimentary rocks. If the rocks are upright, five major lithologic units that may represent an original sedimentary stratigraphic succession can be recognized. At the base of the succession is a unit of biotite gneiss that contains several layers of quartz gneiss and a garnetiferous variety of sillimanitic biotite-quartz gneiss. This unit is overlain by alternating units of microcline gneiss (the Big Five and Quartz Hill layers) and biotite gneiss.

Three varieties of granitic rocks, emplaced in the order granodiorite, quartz diorite, and biotite-muscovite granite, were intruded in Precambrian time. These rocks form small, generally concordant bodies that are discordant in detail. The granodiorite and quartz diorite have gneissic border zones. The gneissic structures in the border zones are parallel to the foliation and lineation in the gneissic country rocks—a fact which indicates that these rocks were deformed and metamorphosed subsequent to their emplacement. The biotite-muscovite granite, on the other hand, is deformed only locally and has a primary flow structure that parallels discordant contacts.

The older deformation produced a major fold system consisting mainly of open upright and asymmetric anticlines and synclines whose axes trend sinuously north-northeast and are spaced 1 to 2 miles apart. The Idaho Springs anticline is the dominating structural feature of the district, and marks the boundary between large areas of contrasting structural trends. In the northwest part of the district the fold axes are nearly horizontal or plunge at low angles, whereas in the southern part the axes plunge steeply northeastward. Abundant folds of similar trend but smaller magnitude occur on the limbs of the major folds; many of these folds are closed and overturned. Most folds regardless of scale are disharmonic. Characteristically, a conspicuous mineral alinement is parallel to the older major fold axes; small-scale folds and boudinage are present almost normal to the major fold axes.

Pegmatitic rocks other than those associated with the granite gneiss are widespread, but they rarely form bodies sufficiently large to map. One variety forms deformed dikes in the granodiorite and quartz diorite, and is probably genetically related to these rocks. Another variety forms a thin undeformed dike; its origin is unknown. A third variety contains disseminated uraninite and large books of biotite; it may be related to the biotite-muscovite granite.

The gneissic rocks of the Idaho Springs district were deformed at least twice in Precambrian time. The older and major deformation was plastic folding; it was accompanied by the emplacement and metamorphism of the granodiorite and quartz diorite and by the recrystallization of the metasedimentary rocks. It was accompanied during its final stage or was followed by the emplacement of the biotite-muscovite granite. The older folds are now outlined by the lithologic units and constitute the structural framework of the district. The younger deformation consisted of small-scale folding and granulation of the previously foliated and deformed rocks and was confined largely to a narrow zone; it was accompanied by weak local recrystallization.

The younger deformation system is an expression of intense granulation and folding that took place in the Idaho Springs-Ralston shear zone, an extensive zone of possibly major structural discontinuity that trends northeastward across the eastern half of the Front Range. Within the Idaho Springs district, this zone is about 2 miles wide; it is largely confined to the southeast half of the district. Here it is characterized by small folds in the incompetent rock masses and by intense cataclasis in the more competent units. Cataclastic products are pervasively distributed, however, through all the rocks in the zone. The younger folds are mainly terrace and chevron types and are locally tightly closed and overturned; the largest has a breadth of about 400 feet. These folds trend N. 55° E., are remarkably straight, and plunge at various angles, largely depending upon their position on the older, larger folds. They are strongly asymmetric and their northwest limbs are raised structurally. Associated with these folds are two lineations, one parallel to the fold axes and one oriented about 80° to the fold axes.

INTRODUCTION

The Idaho Springs district is in the central part of the Front Range and forms an important part of the Front Range mineral belt (fig. 1), a northeast-trending zone of lower Tertiary intrusive rocks and hydrothermal ore deposits. This belt extends entirely across the Front Range and traverses predominantly gneissic and granitic rocks of Precambrian age that form the core of the Front Range.

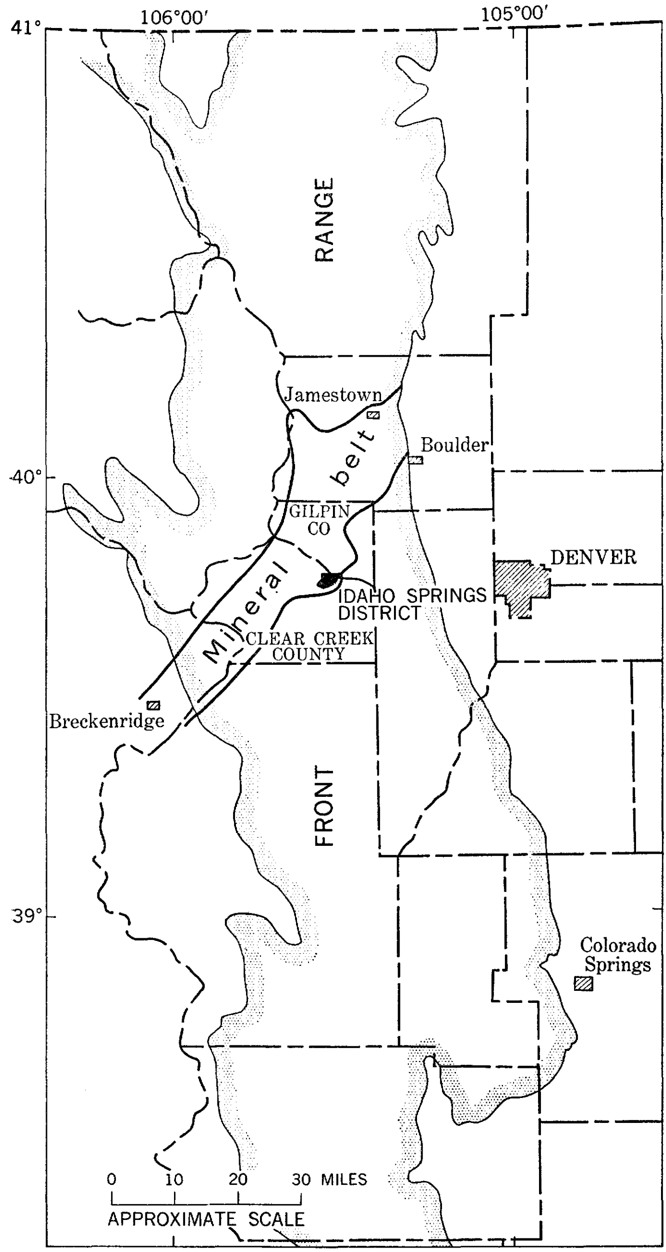

Figure 1.—Map of the Front Range, Colo., showing location of the Idaho Springs district.

The Idaho Springs district embraces about 10 square miles, lies athwart Clear Creek Canyon, and is characterized by moderately rugged topographic relief. The upland near Central City to the north is gently rolling but is deeply incised by Clear Creek Canyon and its tributaries. The maximum local relief is about 2,000 feet. The altitude of the region ranges from slightly over 7,500 feet along Clear Creek to more than 9,900 feet on Pewabic Mountain. Slopes are typically steep in the district, averaging approximately 35° in Clear Creek Canyon. Topographic forms are not markedly influenced by the bedrock.

Bedrock is generally well exposed. Outcrops are most abundant on ridges and sparsely vegetated south-facing slopes; they are least abundant in areas of low relief and on north-facing slopes where vegetation is thicker.

PURPOSE AND SCOPE OF REPORT

The Precambrian bedrock of the Idaho Springs district was mapped in detail during an investigation of the uranium and associated ore deposits of the district as part of a larger study in the central part of the Front Range mineral belt. This report describes the petrography and structure of the Precambrian rocks. The economic geology of the sulfide-bearing veins of the district will be described in a report being prepared by R. H. Moench and A. A. Drake, Jr. Because the district is centrally located in the crystalline core of the Front Range, the findings presented here should contribute to a better understanding of the structural framework of the range.

PREVIOUS AND CONCURRENT STUDIES

Spurr, Garrey, and Ball (1908) studied the southern part of the Idaho Springs district in their comprehensive investigation of the geology and ore deposits of the Georgetown quadrangle. Sydney H. Ball, who was primarily responsible for the study of the general geology of the region, published a paper (1906) in which he defined the Idaho Springs Formation as "* * * a series of interbedded, metamorphic, crystalline rocks, presumably of sedimentary origin, which are typically exposed in the hills surrounding Idaho Springs." Later, Bastin and Hill (1917) studied the northern part of the district in their investigation of the Central City quadrangle, and more recently Lovering and Goddard (1950) compiled available data from the district and added much new information in their report on the geology and mining districts of the Front Range.

Concurrent with the study of the Idaho Springs district, other field parties of the U.S. Geological Survey mapped the adjoining districts (fig. 2). The Freeland-Lamartine district and the Chicago Creek area, southwest of the Idaho Springs district, were investigated by

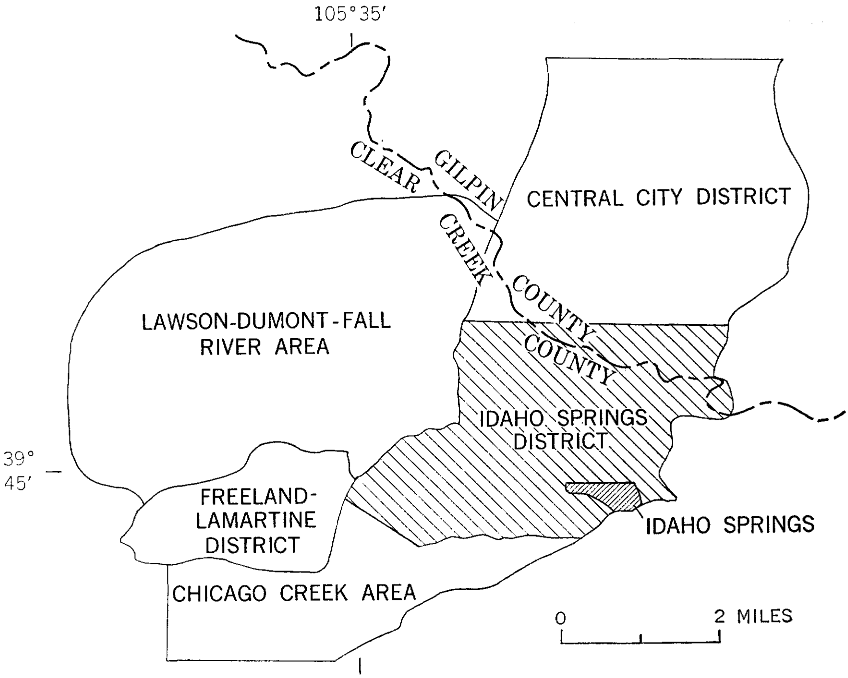

FIGURE 2.—Index map showing the relation of the Idaho Springs district to adjoining mining district.

Harrison and Wells (1956; 1959), the Lawson-Dumont-Fall River area to the west was studied by C. C. Hawley and F. B. Moore, and the Central City district to the north was examined by Sims, Drake, and Tooker (1963). In addition, Tertiary intrusive rocks of the Idaho Springs and adjoining districts were studied by Wells (1960). The uranium deposits of the region were investigated by Sims and other members of the project (Sims and others, 1963), and the altered wallrocks adjacent to the sulfide-bearing veins were studied by Tooker (1956). In addition, the joint pattern in the Precambrian rocks was investigated by Harrison and Moench (1961). A report on the Precambrian structure of the Idaho Springs and adjoining mining districts (fig. 2) was published in 1962 (Moench, Harrison, and Sims).

FIELDWORK

Fieldwork in the Idaho Springs district was done during the summers of 1953 and 1954; approximately 4 man-years of field and laboratory time was devoted to the study. The surface geology was mapped on a scale of 1:6,000 on a special topographic base map prepared by the U.S. Geological Survey from aerial photographs taken in 1951.

The accessible mines were mapped on scales of 1:480, 1:600, and 1:1,200. The writer mapped the central part of the district; other parts of the district were mapped by his colleagues, as shown on plate 1.

ACKNOWLEDGMENTS

I wish to acknowledge the help of my colleagues A. A. Drake, Jr., J. E. Harrison, C. C. Hawley, F. B. Moore, P. K. Sims, and J. D. Wells of the U.S. Geological Survey, all of whom mapped parts of the surface and some of the mines in the Idaho Springs district. In particular my understanding of the structural geology of the district was greatly clarified through many discussions with J. E. Harrison and P. K. Sims. This report represents a part of the studies of the larger Idaho Springs-Central City area by the U.S. Geological Survey on behalf of the Division of Raw Materials of the Atomic Energy Commission.

GEOLOGIC SETTING

The Front Range is composed of several types of generally conformable gneissic rocks that have been invaded by at least three varieties of intrusive rocks, all Precambrian in age. As shown by Lovering and Goddard (1950, pl. 2), the oldest rocks are dominantly biotite gneisses, called the Idaho Springs Formation, amphibolite and associated hornblendic rocks, called the Swandyke Hornblende Gneiss, and a thick quartzite unit at Coal Creek. Most of these gneissic rocks, the relative ages of which are not known, are thought to represent sedimentary rocks that have been deformed, recrystallized, and partly reconstituted at considerable depth and at high temperatures. The dominant varieties of intrusive rocks, the Boulder Creek Granite, Silver Plume Granite, and Pikes Peak Granite, form innumerable plutons, some of which are of batholithic dimensions. Except for local areas, little is known of the structure of the gneissic and intrusive rocks. The gross structural fabric of the Front Range, as delineated by Lovering and Goddard (1950), consists of areas of markedly diverse structural trends, but the relations of these diverse trends to one another is not yet known.

The recent studies of the central part of the Front Range have shown that the gneissic rocks in this area have been deformed at least twice in Precambrian time—first, plastically at high confining pressures and temperatures and, later, cataclastically at somewhat lower pressures and temperatures (Moench, Harrison, and Sims, 1962). The first deformation was the major, pervasive, gneiss-forming episode; it was accompanied by the emplacement of many bodies of intrusive rocks. The second deformation was largely restricted to a narrow northeast-trending zone, the Idaho Springs-Ralston shear zone (Tweto and Sims, 1960), that extends through the Idaho Springs district and is subparallel to the Front Range mineral belt (Sims, Moench, and Harrison, 1959; Tweto and Sims, 1960; 1963, p. 998).

Within the Idaho Springs district, the first period of deformation produced large folds whose axes trend north-northeast. The deformation was largely plastic and was accompanied by thorough metamorphism. The second deformation was confined largely to the southeast side of the district; it produced intense cataclasis and many small folds whose axes trend about N. 55° E.

TERMINOLOGY

Some terms used in this report do not connote the same meaning to all geologists and accordingly are discussed below.

Foliation.—Foliation is used as a general term for three types of planar structures in the Precambrian rocks: (1) planar mineral orientation and compositional layering, which with few exceptions are parallel, characteristic of gneissic rocks; this type of foliation is analogous to the gneissic structure of high-grade Precambrian metamorphic rocks in many parts of the world; (2) subparallel anastomosing close-spaced fractures characteristic of cataclastically deformed rocks; (3) primary flow structure expressed as a planar alinement of tabular minerals or tabular inclusions of wall rocks in the granitic intrusive rocks.

Lineation.—Lineation is used as a descriptive term for all small linear features within or on a rock (Cloos, 1946, p. 1). Five categories of lineations can be distinguished in the Idaho Springs district: axes of small folds, elongate minerals and mineral aggregates (mineral alinements), boudinage, rodding, and slickenside striae.

Massive texture.—The term "massive" is used to describe the texture of rocks that lack distinct foliation and lineation and that are nearly homogeneous in composition.

Open versus closed folds.—Folds of the district can be referred to conveniently as either open or closed. My use of the terms "open" and "closed," however, departs from the definitions of Billings (1954, p. 45), because nearly all folds of the district show some evidence of flowage of mobile layers, regardless of the tightness of folding. In this report, an open fold is one whose apical angle between opposing limbs is 90° or larger, and a closed fold is one whose apical angle is less than 90°.

Plastic versus cataclastic deformation.—The adjectives "plastic" and "cataclastic" are used to contrast the two recognized deformations in the district. "Plastic deformation" is restricted to processes that apparently did not involve rupture (Moench, Harrison, and Sims, 1962) and which characteristically produced disharmonic major folds, minor folds that typically have rounded crests and troughs, mineral alinements, and a subequidimensional granoblastic rock texture. Cataclastic deformation involved rupture of individual grains even though cohesion of the deformed mass was maintained. The characteristic products are slickenside striae, pervasive mortar and mylonitic textures, and folds that are more harmonic and tend to have sharper crests and troughs than the folds characteristic of the plastic deformation.

Coordinate system, A, B.—It is convenient to refer the linear elements of a fold system to directional coordinates. In this report, "B" refers to major fold axes and to linear elements that are about parallel to them, and "A" refers to linear elements that are nearly at right angles to the trends of the major fold. Inasmuch as two fold systems of different ages are recognized, the linear elements of the older system are called B_o and A_o, and the linear elements of the younger system are called B_y and A_y.

PETROGRAPHIC METHODS

Thin sections made from specimens collected from all Precambrian rock types were studied by standard petrographic methods. Each specimen was cut at right angles to the visible planar and linear structures, and modal analyses were made of all suitable thin sections. Wentworth stage and point-count methods were used. Because of the layered, inhomogeneous character of most of the gneissic rocks, the thin sections are not wholly representative of the outcrops or even of the specimens from which they were cut. Accordingly, only 400–600 points were counted by the point method, a technique that is sufficiently accurate for general descriptive purposes. Minerals present in amounts of less than 1 percent are reported as trace constitutents.

The composition of plagioclase feldspar was determined in many specimens by the method of maximum extinction on albite twin lamellae by use of a universal stage. In some specimens the anorthite content was estimated rapidly from the maximum extinction angles obtained without using a universal stage and from the optic sign and the approximate indices of refraction relative to balsam and quartz. Because plagioclase and potassium feldspar were readily distinguishable, the thin sections were not stained.

PRECAMBRIAN ROCKS

The Precambrian bedrock of the district is composed of gneissic, granitic, and pegmatitic rocks. The gneissic rocks include all rocks that have a distinct gneissic structure. As shown on plate 1, three varieties of gneissic rocks—interlayered biotite gneisses, granite gneiss and pegmatite, and microcline-quartz-plagioclase-biotite gneiss—are dominant and form generally conformable layers that outline the structural framework of the district. Other varieties of gneissic rocks, such as quartz gneiss, amphibolite, various calcium-silicate gneisses, and garnet-hornblende-orthopyroxene gneiss, form thin conformable layers and small lenses. The granitic rocks have a dominantly massive, equigranular granitic texture; they consist of three main varieties—granodiorite, quartz diorite, and biotite-muscovite granite. These rocks form many bodies, most of which are too small to be shown on plate 1. The pegmatitic rocks are coarse-grained seriate rocks that have the general composition of granite but which are not associated with the granite gneiss and pegmatite; they also form numerous bodies, most too small to be mapped on plate 1.

Detailed mapping of the Precambrian rocks in the Idaho Springs and adjoining districts has shown that some modifications must be made in the nomenclature of certain rock units established by Lovering and Goddard (1950, p. 19–29) and previous investigators. Accordingly, the rocks are described in this report only in terms of lithology; no formal designation is made. A comparison of the nomenclature used in this report with that of previous workers is shown in table 1. Notably, formation names, such as Idaho Springs Formation, Swandyke Hornblende Gneiss, Boulder Creek Granite, and Silver Plume Granite, are not used in this report. In view of the present state of knowledge of the stratigraphy of the gneissic rocks and correlations of the granitic rocks, such names are premature, for the gneissic rock units do not have specific stratigraphic meaning.

The nomenclature used in this report is similar to that used by Harrison and Wells (1956; 1959), but differs in three important respects: First, in the Freeland-Lamartine district and Chicago Creek area (Harrison and Wells, 1956, p. 38–41; 1959, p. 5–6), biotite-quartz-plagioclase gneiss (or biotite-quartz gneiss) and sillimanitic biotite-quartz gneiss were mapped as separate units. In the Idaho Springs district, these rocks are not separable as map units and are combined and termed interlayered biotite gneisses. Second, in the Freeland-Lamartine district, Harrison and Wells (1956, p. 49, 50) mapped a unit termed migmatite, a mixture of about equal amounts of biotite gneisses and the granite gneiss and pegmatite. In the Idaho Springs district, migmatite was not separated as a map unit. Instead, units of interlayered biotite gneisses and granite gneiss and pegmatite were mapped; these units grade by admixture from one to the other through migmatite. Third, in the Chicago Creek area (Harrison and Wells,

TABLE 1.—*Nomenclature of approximately equivalent rock units in the Idaho Springs district, as mapped by present and previous workers*

[Terms in parentheses are shortened from rock names and are generally used in text]

Ball (1906; in Spurr Garrey, and Ball, 1908)	Bastin and Hill (1917)	Lovering and Goddard (1950)	This report
			Gneissic rocks
Idaho Springs formation and Lime-silicate member.	Idaho Springs formation	Idaho Springs formation	Interlayered biotite gneisses (biotite gneiss). Quartz gneiss. Calcium-silicate gneiss.
Hornblende gneiss	Hornblende schist and lime-silicate rocks of Idaho Springs formation.	Swandyke hornblende gneiss.	Amphibolite. Calcium-silicate gneiss.
Gneissoid granite	Granite gneiss	Quartz monzonite gneiss and gneissic pegmatite.	Microcline-quartz-plagioclase-biotite gneiss (microcline gneiss).
Granite-pegmatite and associated granites and granite porphyry.	Granite pegmatite	Pegmatite	Granite gneiss and pegmatite (granite gneiss). *Pegmatitic rocks* Pegmatite.
			Granitic rocks
Quartz monzonite	Silver Plume granite	Boulder Creek granite and quartz monzonite.	Granodiorite.
Quartz-bearing diorite	Quartz diorite	Quartz diorite	Quartz diorite.
Silver Plume granite	Silver Plume granite	Silver Plume granite	Biotite-muscovite granite.

1959, p. 10), a unit termed quartz monzonite gneiss was mapped. This term was adopted from the bulk composition of the unit, determined from exposures in the Idaho Springs and Central City districts. Subsequent to the publication of the work of Harrison and Wells, however, the term quartz monzonite gneiss was changed to microcline-quartz-plagioclase-biotite gneiss (Moench, Harrison, and Sims, 1962). This change was made because the former term carries an igneous connotation, whereas the unit is probably metasedimentary in origin. (See p. A33.)

The rock names are chosen on the basis of the quantitative mineral content, the presence of megascopically visible diagnostic minerals, and the structure of the rock, as described by Harrison and Wells (1959, p. 5).

GNEISSIC ROCKS

Three major gneissic rock units—interlayered biotite gneisses, granite gneiss and pegmatite, and microcline-quartz-plagioclase-biotite gneiss—form all but a small percentage of the exposed Precambrian rocks of the district (pl. 1). For convenience, in most of the text these terms are shortened to biotite gneiss, granite gneiss, and microcline gneiss, respectively. The minor gneissic rock types are quartz gneiss, amphibolite, various calcium-silicate gneisses, and garnet-hornblende-orthopyroxene gneiss. These units are conformably interlayered, and they outline the major folds of the district.

Biotite gneiss and granite gneiss are intimately mixed in layers and lenses that range from a fraction of an inch to several hundred feet thick. Rarely is an outcrop of one rock type devoid of the other, and map units were separated only on the basis of the relative abundance of the two rock types. For this reason, mapped layers of biotite gneiss contain variable but subordinate amounts of granite gneiss, and vice versa.

The microcline gneiss, on the other hand, is readily distinguishable from two of the other main rock types. Although the microcline gneiss commonly contains small lenses of amphibolite and, less commonly, granite gneiss and biotite gneiss, these lenses are distinctly subordinate.

BIOTITE GNEISS

Interlayered biotite gneisses, or biotite gneiss, underlie more than one-third of the mapped area (pl. 1) and are equivalent to rocks previously called the Idaho Springs Formation by Ball (1906) and by Lovering and Goddard (1950, pl. 2). Two main varieties of biotite gneiss are recognized: biotite-quartz-plagioclase gneiss and sillimanitic biotite-quartz gneiss; both are locally garnetiferous. The interlayers range from about an inch to 20 feet thick and are assumed to represent original bedding. Commonly the boundaries between layers are gradational. The presence in most exposures of conformable layers and lenses of granite gneiss emphasizes the layered appearance of the unit.

BIOTITE-QUARTZ-PLAGIOCLASE GNEISS

Typical biotite-quartz-plagioclase gneiss is gray, fine grained, equigranular, and massive to faintly foliated. Some varieties have a distinct gneissic structure and a lineation shown by the alinement of groups of biotite crystals. Others are lighter gray and somewhat coarser grained than average. In places the rock has been cataclastically deformed, and mortar and mylonite textures are visible in hand specimens.

As shown by the modes given in table 2, the proportions of the major constituents—quartz, plagioclase, and biotite—are variable, and no one specimen can be considered as typical in composition. Microcline is absent in specimens from the Idaho Springs district, and this feature distinguishes the rock from some similar-appearing varieties of microcline gneiss. Garnet is present locally, and one garnetiferous specimen also contains abundant magnetite.

The composition of plagioclase is variable, ranging from sodic oligoclase to calcic andesine. The plagioclase is mostly unzoned, and within a single specimen it is remarkably uniform in composition. Albite and pericline twinning are common. The plagioclase composition does not seem to vary systematically with the quartz or biotite content of the rocks. In most specimens the plagioclase is variably altered to clay.

TABLE 2.—*Modes (volume percent) of biotite-quartz-plagioclase gneiss*

Mineral	Mode of specimen															
	MY-129b	MX-932	MX-756-2	MX-546	MX-356	MX-273	MX-217-1	MX-177-1	MX-152-2	MX-152-1	3-1-14a	MZ-103	MY-738-2	MY-244	2-2b-20	2-1-41a
Quartz	61	39	36	23	46	17	49	28	23	51	52	36	52	46	36	52
Plagioclase	26	44	49	52	39	63	29	58	54	37	35	49	28	43	11	17
Biotite	11	15	13	22	13	14	21	12	21	10	12	12	17	7	27	25
Muscovite	Trace	0	0	0	Trace	0	0	0	0	0	1	1	2	4	Trace	3
Magnetite	2	2	2	3	2	Trace	1	2	1	2	Trace	2	Trace	Trace	12	Trace
Garnet	0	0	0	0	0	0	0	0	0	0	0	0	0	Trace	13	3
Zircon	Trace	0	Trace	Trace	Trace	0	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace	0	Trace
Epidote	Trace	Trace	0	Trace	0	6	0	Trace	0	--------	0	0	0	0	0	0
Apatite	Trace	0	Trace	Trace	0	Trace	Trace	Trace	1	Trace	Trace	Trace	Trace	0	Trace	Trace
Allanite	0	Trace	0	Trace	0	Trace	0	Trace	0	0	0	0	0	0	0	0
Sillimanite	0	0	0	0	Trace	0	0	0	0	0	0	0	0	0	0	0
Sphene	--------	--------	0	0	0	Trace	0	Trace.	0	0	0	0	0	0	0	0
Plagioclase: $X'\wedge(010)$[1]	5° to 7°	23° to 26.5°	13° to 14°	9.5° to 10°	2° to 4°	17° to 18.5°	4° to 4.5°	14.5° to 16.5°	24° to 25°	25.5° to 27°	[2]	[2]	[2]	−6° to −7°	[2]	[2]
Composition	Calcic oligoclase	Calcic andesine	Sodic andesine	Calcic oligoclase	Middle oligoclase	Sodic andesine	Middle oligoclase	Sodic andesine	Calcic andesine	Calcic andesine	Andesine	Oligoclase	Oligoclase	Sodic oligoclase	Andesine	Andesine

[1] Determined on universal stage.
[2] Not determined; plagioclase composition estimated from optic sign, relative refractive indices, and maximum $X'\wedge(010)$ obtainable.

Specimen listings:

MY-129b: Sheared well-foliated variety, from north side of Clear Creek, 2,400 feet southeast of junction with Trail Creek.

MX-932: Sheared well-foliated variety, from thin lens in microcline gneiss 100 ft north of contact with unit of biotite gneiss at head of Hukill Gulch.

MX-756-2: Massive fine-grained variety, from west side of ridge 3,300 ft N. 20° W. of summit of Bellevue Mountain.

MX-546: Sheared well-foliated variety, from ridge 4,700 ft N. 54° W. of mouth of Virginia Canyon.

MX-356: Massive fine-grained variety, from Bellevue Mountain, 250 ft north of summit.

MX-273: Massive fine- to medium-grained variety, from thin lens in microcline gneiss 1,200 ft N. 55° E. of mouth of Trail Creek.

MX-217-1: Sheared well-foliated variety, from north side of Clear Creek, 2,000 ft southeast of mouth of Trail Creek.

MX-177-1: Sheared well-foliated variety, from lens in microcline gneiss, 1,500 ft N. 77° E. of mouth of Trail Creek.

MX-152-2: Well-foliated fine-grained variety, from south side of Bellevue Mountain, 2,000 ft south of summit.

MX-152-1: Massive to moderately well-foliated variety; interlayered with MX-152-2.

3-1-14a: Massive to poorly foliated variety, from about 4,500 ft upstream of mouth of Trail Creek.

MZ-103: Massive fine- to medium-grained variety, from knoll 2,600 ft N. 18° E. of margin of area on southwest end of Spring Gulch.

MY-738-2: Sheared will-foliated variety, from south side of Clear Creek, 1,050 ft S. 69° W. of portal of Big Five tunnel.

MY-244: Sheared moderately well foliated garnetiferous variety, from thin lens (not shown on pl. 1) in microcline gneiss, 1,100 feet N. 38° W. of portal of Big Five tunnel.

2-2b-20: Well-foliated garnetiferous variety, from thin lens (not shown on pl. 1) in granite gneiss about 8,500 ft upstream from mouth of Trail Creek.

2-1-41a: Massive fine-grained variety, from ridge on northwest side of Trail Creek, about 7,200 ft S. 75° W. of mouth of Trail Creek.

The composition of the garnet in the garnetiferous variety of biotite gneiss was not determined, but in the Freeland-Lamartine district, where this rock is more abundant, the garnet is almandite (Harrison and Wells, 1956, p. 40). It forms equant subhedral grains that locally poikilitically include quartz, magnetite, and biotite.

The magnetite, as viewed in a polished section, does not contain visible intergrowths of ilmenite. Hematite is present locally and forms thin veinlets adjacent to cracks in the magnetite.

Viewed microscopically, the rock is mainly fine grained, equigranular, and granoblastic in texture. Quartz and plagioclase grains average about one-half millimeter across and form a mosaic groundmass through which subparallel plates of biotite are disseminated. In the more strongly foliated varieties the major constituents are segregated into biotite-rich and biotite-poor bands, and quartz and plagioclase as well as biotite tend to be elongated parallel to the bands. Plagioclase shows albite and pericline twinning and is variably altered to clay. Biotite forms dark-brown anhedral plates and commonly contains zircon inclusions having pleochroic halos. Magnetite forms small equant grains and crystals that are disseminated through the rock. Muscovite, where present, is intergrown with biotite and locally forms plates that transect biotite plates. Textures of cataclastically deformed specimens are described on page A54.

The biotite-quartz-plagioclase gneiss contains sparse subrounded pods of calcium silicate rock. One pod, exposed on the north side of Clear Creek, about 2,400 feet southeast of the mouth of Trail Creek and about 150 feet above the level of U.S. Highway 6 and 40, is nearly spherical and about 10 inches in diameter. It is concentrically banded, but has an internal foliation produced by the subparallel orientation of pyroxene grains that cross the concentric bands. The foliation in the surrounding biotite-quartz-plagioclase gneiss wraps around the pod. Viewed in thin section, the outer part of the pod contains 62 percent quartz, 22 percent andesine, 14 percent biotite, and 2 percent combined apatite, magnetite, and zircon. It is fine grained and exhibits a granoblastic texture. The core of the pod is largely a fine-grained mosaic of quartz and bytownite and contains abundant interstitial and poikilitic epidote. It contains 42 percent quartz, 27 percent bytownite, 25 percent epidote, 5 percent augite, and 1 percent combined sphene, magnetite, and allanite. Augite grains are elongate, subparallel, sparsely disseminated, and embayed by epidote. The rounded form and concentric banding of the pods suggest that they may be metamorphosed calcareous concretions.

735-934—64——2

SILLIMANITIC BIOTITE-QUARTZ GNEISS

Sillimanitic biotite-quartz gneiss is typically dark gray and flecked with bundles of white sillimanite fibers. It contains a conspicuous foliation produced by the parallel orientation of biotite flakes and elongate quartz grains and by the segregation of these minerals into biotite-sillimanite-rich and quartz-feldspar-rich layers. Some specimens are true biotite schists. Lineation is produced by minute crinkles and by the subparallel orientation of sillimanite fibers and elongate groups of biotite flakes. In places the rock has a cataclastic texture, but this texture is commonly difficult to recognize megascopically because slippage took place largely along the previously formed foliation.

The rock is composed largely of quartz, biotite, and sillimanite in variable proportions (table 3). A light-colored variety exposed near the mouth of Spring Gulch contains only 6 percent biotite. Plagioclase (oligoclase) and microcline are abundant in many specimens, and muscovite is nearly ubiquitous.

The composition of the garnet was not determined precisely, but available data indicate that it is almandine. A quantitative spectrographic analysis (by R. G. Havens, U.S. Geological Survey) shows only 2.1 percent manganese, and semiquantitative spectrographic determinations (by the same analyst) show less than 1 percent calcium, about 3 percent magnesium, and a large amount of iron. The garnet is pink in hand specimen (colorless in thin section), anhedral, and poikilitically includes fragments of biotite.

Viewed microscopically, the sillimanitic biotite-quartz gneiss, where not cataclastically deformed, is fine to medium grained and commonly shows a layered intergrowth of elongate quartz grains interspersed with layers of biotite and sillimanite. Brown biotite forms thin, bent flakes and commonly contains zircon inclusions having pleochroic halos. Sillimanite is in bundles of minute fibers that penetrate all the important minerals of the rock, particularly biotite. Muscovite forms ragged flakes that are commonly interleaved with biotite and locally crosscut biotite plates. The muscovite flakes in places truncate sillimanite fibers and locally are penetrated by sillimanite, but the age relations between these minerals were not unequivocally determined. Quartz forms anhedral elongate grains. Magnetite typically forms small anhedral grains that are elongated parallel to the foliation, but locally it forms small subhedral crystals. Plagioclase is locally an important constituent of the rock, particularly in specimens obtained from the transition zone between layers of sillimanitic biotite-quartz gneiss and biotite-quartz-plagioclase gneiss. It forms anhedral, nearly tabular grains that are intergrown with quartz.

TABLE 3.—*Modes (volume percent) of sillimanitic biotite-quartz gneiss*

Mineral	Mode of specimen									
	3-1-14b	MY-190	MY-129c	MY-116	MX-217-2	2-2b-32c	MY-703	MY-305-1	Z-141b	MC-2
Quartz	51	26	42	63	34	41	47	41	41	33
Plagioclase [1]	10	0	0	10	0	30	21	35	20	7
Microcline	1	0	0	Trace	Trace	7	2	7	23	0
Biotite	26	40	39	24	39	17	21	10	6	36
Muscovite	8	2	6	Trace	9	4	9	5	5	1
Sillimanite	4	30	12	3	18	Trace	Trace	2	4	21
Magnetite	Trace	2	1	Trace	Trace	1	Trace	Trace	1	Trace
Garnet	0	0	0	0	0	0	0	0	0	2
Zircon	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace
Apatite	Trace	0	Trace	Trace	Trace	0	Trace	0	0	0

[1] Plagioclase grains are oligoclase; a small amount of albite is associated with microcline.

Specimen listings:

3-1-14b: Well-foliated rock interlayered with 3-1-14a (table 2).
MY-190: Sheared well-foliated rock, from west side of ridge 2,100 ft S. 25° E. of summit of Bellevue Mountain.
MY-129c: Sheared well-foliated rock, interlayered with MY-129b (table 2).
MY-116: Sheared well-foliated rock, from north side of Clear Creek, 2,800 ft southeast of mouth of Trail Creek.
MX-217-2: Well-foliated rock interlayered with MX-217-1 (table 2).
2-2b-32c: Well-foliated rock, from thin layer (not shown on pl. 1) about 6,500 ft upstream from mouth of Trail Creek.
MY-703: Sheared sillimanite-poor variety, from south side of Clear Creek, about 1,400 ft southeast of mouth of Trail Creek.
MY-305-1: Sheared well-foliated variety from north side of Clear Creek, 1,500 ft east of portal of Big Five tunnel.
Z-141-b: Unsheared moderately well foliated, light colored variety; exposed 300 ft north of mouth of Spring Gulch.
MC-2: Sheared well-foliated garnetiferous variety, from south side of hill north of Idaho Springs townsite.

Most specimens contain microcline intergrown with quartz and plagioclase, commonly in knots and layers that are segregated from the biotite-sillimanite-rich parts of the rock. In some specimens these minerals are not segregated, and microcline is in direct contact with sillimanite and commonly penetrated by sillimanite fibers. The microcline is perthitic, and both microline and plagioclase are locally rimmed by thin layers of albite.

In some specimens a cataclastic texture similar to that described on page A55 is well formed. Biotite is finely comminuted and smeared out along close-spaced anastomosing subparallel fractures that typically follow the rock layering and wrap around eye-shaped aggregates of quartz and feldspar. Quartz in the groundmass is strongly strained and shows evidence of multiple fracturing and rehealing. In the eye-shaped relicts, both quartz and feldspar are deformed. In the crests of crinkles related to the younger deformation, biotite is bent and minute sillimanite fibers are bent and broken. In places, however, larger sillimanite needles have grown in the crests of the crinkles. Such needles are undeformed and locally poikilitically include magnetite grains. In addition, small aggregates of undeformed biotite seem to have formed locally at the expense of finely comminuted biotite.

GRANITE GNEISS AND PEGMATITE

Granite gneiss and pegmatite, or granite gneiss, is found in most exposures of gneissic rocks throughout the mapped area and is most abundant in the southern half of the district (pl. 1). It is equivalent to the pegmatite of Lovering and Goddard (1950, pl. 2). Good exposures can be seen on the hill just north of Idaho Springs in bold light-colored outcrops on the northwest side of Trail Creek and on the north side of Spring Gulch.

The granite gneiss (pl. 1) grades from a leucocratic rock of granitic composition that contains only sparse wisps of dark biotitic material to one that is composed of about equal proportions of biotite gneiss and material of granitic composition. Some granite gneiss is also associated with microcline gneiss and locally forms mappable units within this rock type.

The granite gneiss forms light-colored layers and lenses that range in thickness from less than an inch to several hundreds of feet. The layers are generally conformable, and some may be traced several thousand feet. At places, however, the contacts crosscut the gneissic structure of the adjacent rocks. Such crosscutting contacts are typically ragged and gradational. Inclusions of biotite gneiss commonly have ragged ends that feather out into granite gneiss.

Typical granite gneiss, excluding inclusions, is light colored (leucocratic) due to the paucity of mafic minerals. It is fine grained to pegmatitic and generally has a well-formed gneissic structure. Foliation is produced by alternating layers and lenses of differing grain size, by the parallel elongations of quartz and feldspar, and by inclusions of biotite gneiss. Lineation is commonly seen only in the layers of biotite gneiss within the rock. The composition of the granite gneiss, excluding biotitic inclusions and rare varieties of the rock, approximates that of a leucocratic granite. The dominant minerals are quartz, microcline, and plagioclase (table 4), the proportions of which are variable (fig. 3).

Viewed in thin section, the rock has a variable grain size (seriate texture). Where not cataclastically deformed, the fine-grained variety contains quartz and feldspar that show a granoblastic texture and tend to be elongated parallel to the foliation. Microcline is strongly perthitic, and oligoclase grains are commonly rimmed by albite, particularly along contacts of the microcline and oligoclase. Muscovite forms sparse ragged flakes along the foliation and commonly replaces oligoclase. Biotite, where present, forms flakes and ragged books along the foliation, and commonly is interleaved with muscovite. Locally biotite is altered to chlorite.

Where cataclastically deformed, the granite gneiss has been fractured and locally finely granulated. Large feldspar crystals have been broken and healed with quartz, and magnetite and biotite have been smeared along shear surfaces. Mortar structures are well formed in finer grained parts.

The pegmatitic variety forms knots and lenses of various sizes in the unit throughout the district. The largest bodies are exposed on the south side of Clear Creek Canyon, between the two layers of microcline gneiss. One body is about 300 feet thick at the surface and forms at least half of the layer of granite gneiss and pegmatite exposed on the northwest side of the Big Five layer of microcline gneiss. The contacts between the pegmatitic and finer grained parts of granite gneiss are gradational and are largely conformable, though grain size changes locally along the strike of the gneissic structure. In the pegmatite, equant to elliptical microcline cleavage surfaces as much as 3 feet across were seen. The coarse microcline is white and contains graphic intergrowths of clear quartz. Biotite is sparse or absent; where present it forms ragged, bent plates and books as much as 4 inches across. In some exposures, pink microcline containing graphic quartz, masses of milky to pink quartz, and bent books of clear muscovite are intergrown and form knots and irregular layers in the white graphic microcline.

TABLE 4.—*Modes (volume percent) of granite gneiss and pegmatite*

Mineral	Mode of specimen										
	MX-441-2	MX-505	MX-639	3-1-30	3-1-59	MX-205	MX-207	H-6a	MY-310	H-2	MX-654 [1]
Quartz	37	41	48	26	38	42	36	25	19	40	50
Microcline	30	28	36	47	34	30	32	55	61	30	0
Plagioclase [2]	30	31	14	23	22	26	31	18	17	27	0
Biotite	0	0	0	0	4	Trace	Trace	0	1	Trace	Trace
Muscovite	3	Trace	2	4	2	1	1	2	2	1	0
Magnetite	Trace	0	Trace	Trace	Trace	1	Trace	Trace	0	1	50
Zircon	Trace	Trace	0	0	Trace	0	Trace	0	Trace	0	0
Allanite	Trace	0	0	0	0	0	0	0	0	0	0
Chlorite	0	0	0	2	0	0	0	0	0	1	0
Spinel	0	0	0	0	0	0	0	0	0	0	Trace

[1] Mode by visual estimate.
[2] Plagioclase grains are middle to sodic oligoclase; albite is associated with microcline, and forms rims around oligoclase grains.

Specimen listings:

MX-441-2: Sheared rock, from ridge 1,450 ft S. 30° E. of summit of Bellevue Mountain.

MX-505: Slightly granulated rock, from small body 450 ft north of mouth of Hukill Gulch.

MX-639: Sheared rock, from exposure 2,100 feet N. 60° E. of portal of Big Five tunnel.

3-1-30: Sheared rock, from ridge northwest of Trail Creek, 400 ft west northwest of mapped syncline.

3-1-59: Unsheared rock, from south side of ridge north of Trail Creek, 750 ft east-southeast of mapped syncline.

MX-205: Slightly granulated rock, from center of granite gneiss body on north side of Clear Creek, across from mouth of Trail Creek.

MX-207: Sheared rock, from southeast side of same body as MX-205.

H-6a: Slightly granulated rock, from thin layer (not shown on pl. 1) in microcline gneiss, 800 ft upstream from mouth of Trail Creek.

MY-310: Slightly granulated rock, from exposure 1,200 ft N. 55° E. of portal of Big Five tunnel.

H-2: Unsheared rock, from exposure 300 ft upstream from mouth of Trail Creek.

MX-654: Sheared quartz-magnetite gneiss, from contact zone between biotite gneiss and microcline gneiss, 2,400 ft N. 22° E. of portal of Big Five tunnel.

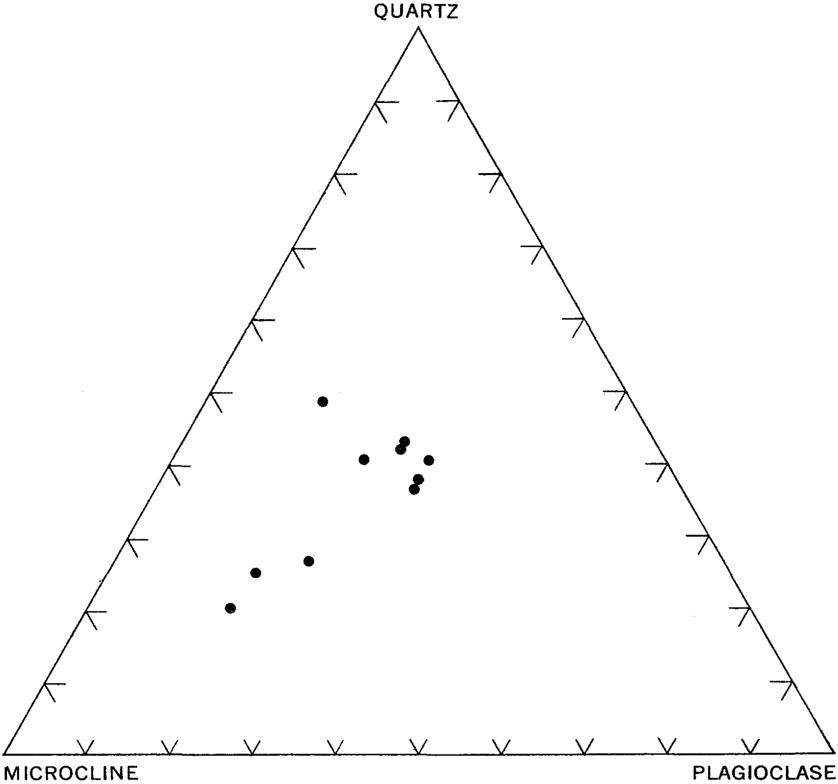

FIGURE 3.—Triangular diagram showing variation in composition (volume percent) of granite gneiss and pegmatite, excluding quartz-magnetite gneiss.

Quartz-magnetite gneiss (table 4), thought to be a variety of granite gneiss because of its spatial association with that unit, is most commonly exposed near contacts between biotite gneiss and microcline gneiss. It is also associated with bodies of granite gneiss within the biotite gneiss. The quartz-magnetite gneiss is coarse grained and contains rounded anhedral grains of magnetite embedded in anhedral quartz. The tendency of both minerals to be elongate and parallel gives a foliation to the rock. Small amounts of biotite are alined along the foliation. A thin section revealed small inclusions of green spinel in the magnetite, and a polished section revealed subhedral plates of ilmenite embedded in the magnetite. Hematite is finely intergrown with the ilmenite and locally veins the magnetite.

MICROLINE GNEISS (MICROCLINE-QUARTZ-PLAGIOCLASE-BIOTITE GNEISS)

The microcline-quartz-plagioclase-biotite gneiss, or microcline gneiss, underlies more than one-quarter of the mapped area. It is equivalent to the quartz monzonite gneiss and gneissic pegmatite of Lovering and Goddard (1950, pl. 2). Best exposures are along U.S. Highway 6 and 40 on the north side of Clear Creek.

The rock is medium gray to very light gray or tan, is fine to medium grained, and has a well-formed gneissic structure. It forms two important layers, the Quartz Hill layer and the Big Five layer. The Quartz Hill layer, the larger of the two, underlies much of the western and northern parts of the district (pl. 1) and most of the Central City district to the north. It wedges out near the southern margin of the Idaho Springs district in upper Spring Gulch (pl. 1). The Big Five layer is composed of discontinuous lenses that are intertongued or infolded with the enclosing gneisses. The Big Five and Quartz Hill layers converge to the southwest, but a junction, if it exists, is not exposed.

Both layers of microcline gneiss contain many conformable layers and lenses of amphibolite, granite gneiss, and biotite gneiss. The amphibolite forms many thin layers and lenses throughout the unit; it seems to be most abundant near the contacts of the microcline gneiss layers. Granite gneiss also forms many small bodies throughout the unit, as well as a thick layer that extends about 1,200 feet east and a mile southwest of the mouth of Trail Creek (pl. 1). Biotite gneiss forms many lenses in the Quartz Hill layer, the largest of which are in Virginia Canyon and near the eastern side of the district (pl. 1). Most of the biotite gneiss layers are composed of biotite-quartz-plagioclase gneiss or mixed biotite-quartz-plagioclase gneiss, amphibolite, and granite gneiss, but the large forked body near the eastern margin of the area contains sillimanitic biotite-quartz gneiss as well.

Most contacts between the major layers of microcline gneiss and the biotite gneiss or granite gneiss are sharp and conformable. These contacts form the most reliable structural markers and outline many small folds of both deformation systems. Inasmuch as the rocks at the southeastern contact of the Big Five layer northwest of Idaho Springs have been intensely granulated, it is difficult to distinguish between rock types.

The microcline gneiss contains variable amounts of quartz, plagioclase, microcline, and biotite (table 5). With few exceptions, microcline, plagioclase, and quartz comprise more than 90 percent of the rock, and microcline is generally subordinate to plagioclase. The composition of the rock ranges from quartz monzonite to quartz diorite (fig. 4). The rock contains a variety of accessory minerals, among which black opaque minerals, zircon, and apatite are most common. A polished surface of one specimen (MX–104–1) showed that magnetite is the most common opaque mineral, but that the magnetite

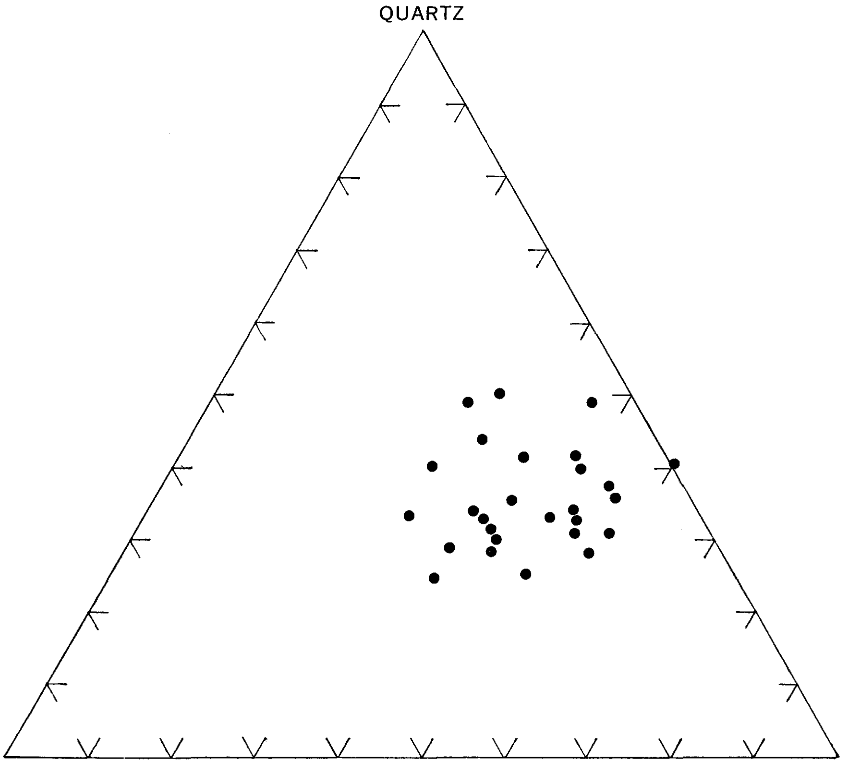

FIGURE 4.—Triangular diagram showing variation in composition (volume percent) of microcline-quartz-plagioclase-biotite gneiss.

commonly contains plates of ilmenite and is veined by hematite adjacent to fractures and grain boundaries. Pyrite was seen in two specimens.

Typically, the microcline gneiss has a well-formed foliation, but it is distinctly less well layered than biotite gneiss. Hand specimens tend to be homogeneous in composition, but within a few feet across strike the composition may grade from one extreme to the other of the other range indicated in table 5. Foliation is produced by the parallel alinement of biotite plates and by fine interlaminations of biotite-rich and biotite-poor laminae, which are typically less than a millimeter thick. Where the rock is not cataclastically deformed, biotite is in discrete, sharply defined subparallel plates or aggregates of plates that can be seen in hand specimens. Where cataclastically deformed, biotite has been smeared into elongate patches as much as several millimeters long. This cataclastic feature is characteristic of the Big Five layer and of much of the Quartz Hill layer in the Idaho Springs district.

TABLE 5.—*Modes (volume percent) of microline-quartz-plagioclase-biotite gneiss*

Mineral	Mode of specimen													
	3-1-5a	3-1-5b	3-1-7	3-1-83	H-4a	MY-354	MY-738-1	MX-178	MX-183	MX-185	MX-189	MX-190	MX-208	MX-211
Quartz	28	27	37	28	47	28	46	33	29	28	32	29	34	48
Microcline	24	25	15	26	5	15	14	35	25	16	17	17	Trace	20
Plagioclase	41	40	37	38	44	49	31	30	41	54	45	52	50	30
Biotite	7	5	8	6	4	7	7	0	Trace	1	3	0	13	2
Muscovite	0	0	0	0	0	1	2	2	0	0	0	0	0	Trace
Hornblende	0	0	0	0	0	0	0	0	0	0	Trace	0	0	0
Black opaques	Trace	1	1	Trace	Trace	Trace	0	Trace	1	1	2	2	3	Trace
Pyrite	0	0	0	0	0	0	0	0	Trace	0	0	0	0	0
Zircon	0	Trace	Trace	Trace	Trace	Trace	Trace	Trace	0	Trace	Trace	Trace	Trace	Trace
Apatite	Trace	Trace	Trace	Trace	0	Trace	0	0	Trace	Trace	Trace	0	Trace	Trace
Allanite	0	Trace	Trace	0	Trace	0	0	0	Trace	0	Trace	0	Trace	Trace
Sphene	0	Trace	Trace	Trace	0	0	0	0	Trace	0	0	0	0	0
Epidote	Trace	1	2	2	Trace	0	0	0	Trace	0	Trace	0	Trace	0
Calcite(?)	0	1	0	0	0	0	0	0	Trace	Trace	0	0	0	0
Garnet	0	0	0	0	0	0	0	0	0	0	0	0	Trace	0
Chlorite	0	0	0	0	0	0	0	0	4	0	1	0	0	0
Plagioclase:														
Core X′ ∧ (010) [1]	(2)	(2)	(2)	(2)	(2)	(3)	(3)	0° to −6°	0° to 1°	−3° to −5°	0° to −1°	−2° to −4°	10° to 11°	0° to −1°
Rim X′ ∧ (010) [1]								−11° to −16°	(2)	(2)	−16°	−17°	(2)	−14°
Core composition	Oligoclase	Oligoclase	Oligoclase	Oligoclase	Oligoclase	----------	----------	Sodic oligoclase.	Middle oligoclase.	Sodic oligoclase.	Middle oligoclase.	Middle oligoclase.	Calcic oligoclase.	Middle oligoclase.
Rim composition	Albite	Albite	----------	Albite	Albite	----------	----------	Albite	Albite	Albite	Albite	Albite	Albite	Albite

Mineral	Mode of specimen													
	MX-213	MX-314	MX-561	MX-18	M-988	MX-104-1	MX-104-2	MX-129-2	MX-26	M-992	MX-13	MX-17	MX-570a	MX-570b
Quartz	41	24	35	33	34	30	34	30	34	32	36	39	27	34
Microcline	20	23	9	15	21	24	22	27	9	14	23	11	11	9
Plagioclase	34	47	51	51	40	38	37	37	52	48	38	46	50	42
Biotite	2	5	5	0	0	5	7	4	2	6	1	0	12	15
Muscovite	3	0	0	0	0	0	Trace	Trace	Trace	0	Trace	2	Trace	Trace
Hornblende	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Black opaques	Trace	1	Trace	1	5	3	Trace	1	3	Trace	2	2	Trace	Trace
Pyrite	0	Trace	0	0	0	0	0	0	0	0	0	0	0	0
Zircon	Trace	Trace	Trace	0	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace
Apatite	0	Trace	0	Trace	0	Trace	Trace	Trace	0	Trace	Trace	0	Trace	Trace

Allanite	0	0	0	0	0	0	0	0	Trace	0	0	0	0	0
Sphene	0	Trace	0	0	0	0	0	0	0	Trace	0	0	0	0
Epidote	0	Trace	Trace	0	0	0	0	Trace	Trace	0	0	0	0	0
Calcite(?)	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Garnet	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Chlorite	0	Trace	0	0	0	Trace	0	1	0	0	0	0	0	0
Plagioclase:														
Core X′ ∧ (010) [1]	0°	3.5° to 4°	0° to −2°	−5° to −6°	−8° to −9°	−5° to −6°	−1° to −4°	−6° to −11°	3° to −3°	−6° to −7°	−4° to −6°	−8°	1° to 3°	0° to 3°
Rim X′ ∧ (010) [1]	−15.5°	−12° to −16°	(2)	(2)	−16° to −18°	−16°	−10°	−15.5° to −16°	−15°	−12°	−13°	−14° to −15°	(2)	−13°
Core composition	Middle oligoclase.	Calcic oligoclase.	Middle oligoclase.	Sodic oligoclase.	Sodic oligoclase.	Sodic oligoclase.	Middle oligoclase.	Sodic oligoclase.	Middle oligoclase.	Sodic oligoclase.	Sodic oligoclase.	Sodic oligoclase.	Middle oligoclase.	Middle oligoclase.
Rim composition	Albite	Albite	Albite	Albite	Albite	Albite	Sodic oligoclase.	Albite	Albite	Albite	Albite	Albite	Albite	Albite

[1] Determined on universal stage.
[2] Not determined; plagioclase composition estimated from optic sign, relative refractive indices, and maximum X′ ∧ (010) obtained without use of universal stage.
[3] Not estimated.

Specimen listings:

3-1-5a: Sheared rock from exposure about 2,100 ft upstream from mouth of Trail Creek.
3-1-5b: Sheared, slightly coarser grained rock from same locality as 3-1-5a.
3-1-7: Sheared rock from exposure about 2,300 ft upstream from mouth of Trail Creek.
3-1-83: Slightly granulated rock from exposure on hill 2,300 ft S. 53° E. of mouth of Trail Creek.
H-4a: Ungranulated rock from exposure about 500 ft upstream from mouth of Trail Creek.
MY-354: Sheared rock from Big Five layer 3,300 ft N. 42° E. of portal of Big Five tunnel.
MY-738-1: Slightly granulated rock from thin layer on south side of Clear Creek 900 ft S. 70° W. of portal of Big Five tunnel.
MX-178: Sheared rock from exposure on north side of Clear Creek 1,400 ft N. 68° E. of mouth of Trail Creek.
MX-183: Slightly granulated rock from exposure 4,100 ft upstream from mouth of Fall River.
MX-185: Strongly sheared rock from exposure 3,800 ft upstream from mouth of Fall River.
MX-189: Moderately granulated rock from exposure 3,200 ft upstream from mouth of Fall River.
MX-190: Moderately granulated rock from exposure 3,000 ft upstream from mouth of Fall River.
MX-208: Moderately granulated rock from north side of Clear Creek 600 ft N. 69° E. of mouth of Trail Creek.
MX-211: Slightly granulated rock from north side of Clear Creek 900 ft east of mouth of Trail Creek.
MX-213: Strongly granulated rock from north side of Clear Creek 1,100 ft S. 75° E. of mouth of Trail Creek.
MX-314: Largely ungranulated rock from west side of Bellevue Mountain 3,600 ft N. 30° W. of summit.
MX-561: Sheared rock from center of Big Five layer 1,350 ft N. 52° W. of portal of Big Five tunnel.
MX-18: Ungranulated rock from west side of Bellevue Mountain 4,150 ft N. 70° W. of summit.
M-988: Ungranulated rock from west side of Bellevue Mountain 4,350 ft N. 59° W. of summit.
MX-104-1: Strongly sheared rock from exposure near mouth of Fall River.
MX-104-2: Strongly sheared rock from exposure near mouth of Fall River.
MX-129-2: Sheared rock from gully 1,300 ft N. 80° E. of mouth of Fall River.
MX-26: Slightly granulated rock from west side of Bellevue Mountain 3,450 ft N. 62° W. of summit.
M-992: Moderately granulated rock from west side of Bellevue Mountain 4,150 ft N. 40° W. of summit.
MX-13: Ungranulated rock from exposure 4,700 ft upstream from mouth of Fall River.
MX-17: Ungranulated rock from west side of Bellevue Mountain 4,150 ft N. 72° W. of summit.
MX-570a: Sheared rock from Big Five layer 1,950 ft N. 9° E. of portal of Big Five tunnel.
MX-570b: Same outcrop as MX-570a.

Viewed in thin section, the rock where not cataclastically deformed is fine to medium grained, subequigranular, and granoblastic in texture. Quartz and, to a lesser degree, plagioclase and microcline grains tend to be elongated parallel to the foliation. Microcline is somewhat perthitic and variably twinned. At places, particularly on the western side of the district, microcline is only faintly twinned. Plagioclase is well twinned (albite and some pericline) and two varieties are present, as described in the following paragraph. Olive-brown to brown biotite forms subparallel anhedral plates having sharp edges. Equant and locally euhedral magnetite grains are disseminated through the rock, and ilmenite occurs in many specimens as plates embedded in magnetite. Hornblende is rare and occurs in small ragged grains. Garnet is likewise rare and occurs in small grains distributed along foliation planes; its composition was not determined. Zircon and apatite form small subhedral grains, and allanite and sphene form larger, but less common, anhedral equant grains. Epidote is common and forms anhedral grains that are typically associated with biotite. At places, biotite is partly chloritized; it is also locally interleaved with and truncated by ragged flakes of muscovite. Muscovite forms sheafs that embay feldspars. Small amounts of secondary interstitial calcite are in some specimens.

Two varieties of plagioclase—differing in composition and habit—may be recognized in most specimens. Equant to somewhat elongate grains form the bulk of the plagioclase in each specimen. The grains are unzoned and range in composition from sodic to calcic oligoclase from one specimen to another. The second variety is albitic in composition and forms rims around the oligoclase grains, particularly along contacts of the plagioclase and microcline. The contacts between the oligoclase grains and albite rims are typically sharp. In places, the albite is myrmekitically intergrown with quartz and the intergrowths embay microcline. Albite also is present in microcline as film-type perthite.

The microscopic features of the cataclastically deformed rocks are similar in their important respects to those described on page A55.

QUARTZ GNEISS

Quartz gneiss, equivalent to part of the Idaho Springs Formation of Ball (1906), forms several thin but extensive layers in the biotite gneiss and granite gneiss in the southeast corner of the district (pl. 1). Good exposures of quartz gneiss can be seen on the south side of the hill just north of the Idaho Springs townsite.

Quartz gneiss is very light gray or tan and fine to medium grained and has a well-formed gneissic structure and a lustrous glassy appearance due to its high quartz content. The rock forms layers that rarely exceed 15 feet in thickness, but one was traced a total distance of about 6,000 feet; others may be equally extensive. The layers are conformable, and their contacts are sharp or grade over a few inches.

As shown in table 6, quartz is the dominant constituent. Sodic plagioclase is locally abundant, and mafic minerals are sparse.

TABLE 6.—*Modes (volume percent) of quartz gneiss*

Mineral	Mode of specimen				
	MY-274	MY-304	MY-308	MY-801	MC-8
Quartz	68	73	60	67	82
Microcline	6	0	6	8	5
Plagioclase [1]	25	24	29	18	6
Biotite	0	Trace	4	2	6
Muscovite	1	3	1	4	1
Magnetite	Trace	Trace	Trace	1	Trace
Zircon	Trace	Trace	Trace	0	Trace
Apatite	0	Trace	0	0	0
Sillimanite	0	0	Trace	0	0

[1] Plagioclase is albite to oligoclase.

Specimen listings:

MY-274: From layer 4 ft thick exposed 850 ft N. 59° E. of portal of Big Five tunnel.
MY-304: From layer 10 ft thick exposed 1,650 ft N. 86° E. of portal of Big Five tunnel.
MY-308: From layer 15 ft thick exposed 1,200 ft N. 71° E. of portal of Big Five tunnel.
MY-801: From layer 1,500 ft S. 19° W. of portal of Big Five tunnel.
MC-8: From layer exposed on south side of hill north of Idaho Springs; exact location not recorded.

Quartz gneiss is exposed only in the area of intense younger deformation, and all specimens studied show evidence of intense granulation. Viewed in thin section, small elliptical eyes of deformed quartz and feldspar lie in a groundmass formed largely of sheared, intensely strain shadowed elongated quartz. Biotite, muscovite, and magnetite are finely comminuted and have been smeared out along subparallel anastomosing fractures. The sillimanite was observed as a few fibers in a relatively undeformed flake of biotite.

AMPHIBOLITE AND ASSOCIATED CALCIUM-SILICATE GNEISS

Amphibolite forms many small layers and lenses throughout the district, and many good exposures are readily accessible along U.S. Highways 6 and 40 on the north side of Clear Creek (pl. 1). Most of the amphibolite bodies are too small to show on plate 1. The largest bodies are more than 100 feet thick, and one layer has an exposed length of more than 4,000 feet. Amphibolite is most abundant along the contacts between microcline gneiss and biotite gneiss. It also forms innumerable small layers and lenses within the microcline gneiss and a few large bodies in biotite gneiss.

Amphibolite is dark gray to black and fine to medium grained; typically it has a distinct gneissic structure, though locally it is massive. Plagioclase (andesine to labradorite) and hornblende are the dominant constituents of typical amphibolite (table 7). Viewed in thin section,

TABLE 7.—*Modes (volume percent) of amphibolite and associated calcium-silicate gneiss*

Mineral	Mode of specimen													
	2-1-61a	2-2b-32a	3-1-8a	H-3	H-4d	MX-127	MX-135-1	MX-406	MX-756-1	MY-129a	MX-672	MY-311-1	MX-212-1	MX-212-2
Quartz	21	1	1	1	1	7	17	3	9	6	6	1	12	31
Plagioclase	24	25	34	26	40	43	28	46	24	29	37	29	55	40
Amphibole	45	73	55	46	57	26	34	51	61	55	57	68	30	4
Biotite	0	1	1	23	0	0	18	Trace	4	8	0	Trace	2	0
Clinopyroxene	0	0	0	0	2	8	0	0	0	0	0	0	0	0
Magnetite	3	Trace	1	Trace	Trace	1	1	Trace	1	Trace	Trace	1	0	4
Apatite	Trace	Trace	3	Trace	Trace	Trace	1	Trace	Trace	1	Trace	Trace	0	0
Sphene	4	0	3	3	Trace	1	Trace	0	0	0	Trace	0	Trace	3
Calcite	0	0	0	1	0	Trace	0	0	0	Trace	0	0	Trace	0
Epidote	3	Trace	2	0	Trace	14	1	0	1	Trace	Trace	1	1	17
Microcline	0	0	0	0	0	Trace	0	0	0	0	0	0	0	0
Allanite	0	Trace	Trace	0	0	0	Trace	0	0	1	0	0	0	0
Zircon	0	0	Trace	0	0	0	0	0	Trace	Trace	0	0	0	0
Rutile	0	0	0	0	0	0	0	0	0	Trace	0	0	0	0
White mica	0	0	0	0	0	0	0	0	0	Trace	Trace	Trace	0	0
Sulfide	0	0	0	0	0	0	0	0	0	0	0	0	0	1
Plagioclase:														
X'∧(010) [1]	(2)	(2)	(2)	(2)	(2)	15°	25° to 28°	21° to 24°	(2)	(2)	(2)	20° to 21°	15.5° to 17°	38° to 48°
Composition	Labradorite	Labradorite	Andesine	Andesine	Andesine	Sodic andesine	Calcic andesine	Middle andesine	Labradorite	Labradorite	Labradorite	Middle andesine	Sodic andesine	Bytownite

[1] Determined on universal stage.
[2] Not determined; plagioclase composition estimated from optic sign, relative refractive indices, and maximum X' ∧ (010) obtainable.

Specimen listings:

- 2-1-61a: Typical amphibolite from ridge 7,300 ft S. 73° W. of mouth of Trail Creek.
- 2-2b-32a: Typical amphibolite (not shown on pl. 1) exposed on northwest side of Trail Creek at east end of thick sill of biotite-muscovite granite.
- 3-1-8a: Typical amphibolite (not shown on pl. 1) exposed in Trail Creek at contact between microcline gneiss and granite gneiss.
- H-3: Biotite-bearing amphibolite exposed on north side of Trail Creek 450 ft upstream from mouth.
- H-4d: Pyroxene-bearing amphibolite from same locality as H-3.
- MX-127: Pyroxene-and-epidote-bearing amphibolite (not shown on pl. 1) exposed in gully 2,100 ft N. 47° E. of mouth of Trail Creek.
- MX-135-1: Biotite-bearing amphibolite (not shown on pl. 1) from north side of Clear Creek about 600 ft north of mouth of Trail Creek.
- MX-406: Typical amphibolite from center of large body exposed on ridge 1,100 ft south-southeast of summit of Bellevue Mountain.
- MX-756-1: Biotite-bearing amphibolite (not shown on pl. 1) from west side of ridge 2,900 ft north-northwest of summit of Bellevue Mountain.
- MY-129a: Biotite-bearing amphibolite (not shown on pl. 1) from same locality as MY-129b (table 2).
- MX-672: Typical amphibolite from center of thick lens 1,000 ft northeast of portal of Big Five tunnel.
- MY-311-1: Typical amphibolite from center of thick lens 1,100 ft northeast of portal of Big Five tunnel.
- MX-212-1: Typical amphibolite from north side of Clear Creek 1,000 ft S. 80° E. of mouth of Trail Creek.
- MX-212-2: Calcium silicate gneiss associated with MX-212-1.

amphibolite typically has an equigranular granoblastic matrix of fine-grained equant plagioclase and subordinate quartz and somewhat larger subparallel blades of anhedral to euhedral hornblende. Commonly these minerals are segregated into hornblende-rich and hornblende-poor layers; this segregation produces the foliation. Lineation is produced by subparallel alinement of hornblende. Plagioclase is mostly unzoned, ranges in composition from sodic andesine to labradorite, and is twinned according to the albite and pericline laws. The hornblende blades are locally poikilitic and contain inclusions of plagioclase and quartz. The optical properties are those of hastingsite, a sodium-bearing variety of green hornblende. Equant subhedral to euhedral grains of magnetite and apatite, and less commonly, sphene and allanite are disseminated through the rock. A bladed opaque mineral, probably ilmenite, was seen also.

In pyroxene-bearing varieties of amphibolite, equant grains of pyroxene are disseminated through the rock or are concentrated in layers. The pyroxene is anhedral and commonly ragged in outline, and appears to have been replaced by hornblende along fractures and cleavages. In one specimen (H–4d, table 7), the pyroxene has the optical properties of diopside; in another (MX–127, table 7) it has the optical properties of augite.

In biotite-bearing varieties of amphibolite, plates of biotite are typically disseminated through the rock. The plates are thin, anhedral on all sides, and straight and undeformed. They appear to be randomly oriented and transect hornblende blades and plagioclase grains. Locally biotite flakes have grown along hornblende cleavages. Commonly the larger biotite plates poikilitically enclose plagioclase and quartz. The biotite ranges from tan to dark brown; locally, it is altered to green chlorite.

Epidote, calcite, microline, and white mica seem to have formed at the expense of hornblende and plagioclase. The epidote is interstitial and markedly embays plagioclase and hornblende along grain boundaries. Commonly it is myrmekitically intergrown with a clear unidentified substance, possibly quartz, albite, or both, all of which are attributable to the alteration of plagioclase. The epidote is commonly most abundant in or near shear surfaces that transect or parallel the foliation. Microcline is interstitial and was seen only in a specimen that contains abundant epidote. Calcite is also interstitial, and it embays plagioclase. Fine-grained muscovite occurs as ragged flakes in plagioclase.

Calcium-silicate gneiss that is associated with amphibolite is lighter colored, commonly greenish, and more coarsely crystalline than the amphibolite. It forms conformable layers and crosscutting veinlike bodies and irregular masses in the amphibolite. The crosscutting contacts are generally gradational and ragged, and in places the gneissic structure of the amphibolite can be traced through the calcium-silicate gneiss. Harrison and Wells (1959, p. 7) have noted similar crosscutting relations between amphibolite and calcium-silicate gneiss.

Specimens of calcium-silicate gneiss may contain, as their dominant constituents, quartz and epidote, or quartz and garnet, or all three minerals.

Variable amounts of amphibole, clinopyroxene, biotite, and plagioclase may also be present. Specimen MX–212–2, in table 7, for example, contains abundant quartz, epidote, bytownite, a small amount of amphibole, and accessory minerals. The amphibole in this rock is tremolite-actinolite. A layer associated with 2–1–61a of table 7 contains the same assemblage with the addition of clinopyroxene. A rock mass associated with 3–1–8a contains garnet, quartz, epidote, hornblende, and labradorite. A small amphibolite body about 1,400 feet north-northwest of the mouth of Virginia Canyon and the large body exposed about 3,000 feet upstream from the mouth of Spring Gulch contain layers and crosscutting masses that are rich in garnet, quartz, clinopyroxene, plagioclase, hornblende, and some biotite; epidote is sparse or absent.

Epidote and most of the quartz appear to have been the last minerals to form. Quartz (and possibly some albite) and epidote are myrmekitically intergrown and embay hornblende, clinopyroxene, and plagioclase, as previously described. Epidote also surrounds and embays garnet. Garnet is typically granular, and its age relations to other minerals are difficult to determine. Locally, garnet poikilitically encloses pyroxene and hornblende, a feature that suggests that it is younger than these minerals.

The composition of the garnet was not determined, but data obtained by Harrison and Wells (1959, p. 8) on garnet from similar calcium-silicate rocks in the Chicago Creek area suggest that it is in the andradite-grossularite series.

Field and petrographic observations indicate that these varieties of calcium-silicate gneiss formed at the expense of amphibolite. The calcium-silicate gneiss is spatially associated with amphibolite and forms masses whose contacts crosscut the gneissic structure of amphibolite. These masses exhibit replacement textures against amphibolite, and microscopically the calcium-silicate minerals—particularly epidote—embay the constituent minerals of typical amphibolite.

CALCIC BIOTITE-QUARTZ-PLAGIOCLASE GNEISS

Calcic biotite-quartz-plagioclase gneiss forms two thin layers on the north side of Clear Creek between the Big Five and Quartz Hill layers of microcline gneiss (pl. 1).

The rock is dark gray to black, fine grained, and has a conspicuous gneissic structure. The layers it forms are conformable and generally less than 10 feet thick. One layer was traced a distance of about 1,700 feet from U.S. Highway 6 and 40 eastward to the Idaho Springs fault. It was not found on the northeast side of the fault.

Calcic biotite-quartz-plagioclase gneiss, composed largely of quartz, biotite, and plagioclase, is similar to biotite-quartz-plagioclase gneiss, but the plagioclase is more calcic, having the composition of bytownite. (Maximum $X' \wedge (010)$ determined using universal stage is 54°.) Modal analyses of two specimens show, respectively, 29 and 34 percent quartz, 28 and 21 percent plagioclase, 32 and 30 percent biotite, 5 and 7 percent muscovite, and 6 and 5 percent accessory minerals. One specimen contains 3 percent hornblende, the optics of which are similar to hastingsite. The accessory minerals are magnetite, apatite, epidote, allanite, and zircon.

Viewed in thin section, the rock is fine grained and exhibits a granoblastic texture that shows evidence of granulation. Plagioclase and strain-shadowed quartz form a mosaic of equant to slightly elongate grains that are traversed by many subparalled fractures. Biotite is in ragged subparallel plates that are strongly bent and locally smeared out along the fractures. In one specimen, hornblende is in large deformed ellipitical grains or eyes. Flecks of muscovite are abundantly disseminated in the plagioclase and are an alteration product of plagioclase. Epidote is anhedral and interstitial; it locally surrounds equant grains of allanite and is commonly associated with biotite.

CALCIUM-SILICATE GNEISS

One variety of calcium-silicate gneiss seems to be unrelated to amphibolite. It forms four bodies on the north and east sides of Pewabic Mountain (pl. 1).

The rock is mottled dark to light, crudely banded, fine to coarse grained, and has a poorly defined gneissic structure. Because of poor exposure in that area, the true shapes of these bodies are not known; the contacts only approximately outline the areas in which calcium-silicate gneiss is abundant in surface float. Possibly the outlined bodies form a single conformable layer.

The rock is composed of various proportions of diopside, epidote, quartz, scapolite, oligoclase, microcline, and arfvedsonite (a sodium-rich amphibole). A carbonate mineral is locally abundant, and sphene, apatite, and magnetite are common accessory minerals. Viewed microscopically, the rock is fine to coarse grained and commonly crudely banded. Diopside is granular and intergrown with epidote, which apparently embays the diopside. The habits of epidote

735-934—64——3

throughout the rock suggest that it was one of the last minerals to form. Arfvedsonite (pleochroic from light yellow green to deep blue, very small 2V) commonly forms ragged flakes along cleavages in diopside, and in places it is in coarse anhedral blades that form dark bands in the rock. Scapolite, quartz, oligoclase, and microcline are granular in texture and commonly form fine- to coarse-grained bands containing smaller amounts of diopside and epidote. The scapolite has a fairly high birefringence and indices of refraction that are distinctly greater than those of associated quartz and feldspar; these properties indicate that the scapolite is calcic in composition. A single grain of pale-orange garnet was seen.

GARNET-HORNBLENDE-ORTHOPYROXENE GNEISS

Exposures of garnet-hornblende-orthopyroxene gneiss are restricted to several small bodies on the prominent ridge about 1,300 feet south-southeast of the summit of Bellevue Mountain (pl. 1).

The rock is dark red brown, fine grained, and massive to weakly layered. It forms small layers and lenses that are embedded in a thick mass of granite gneiss and pegmatite. The layers are generally conformable to the gneissic structure of the enclosing rocks, and their contacts are sharp.

The rock is composed mainly of garnet, and contains two kinds of amphibole as well as orthopyroxene, quartz, magnetite, apatite, and allanite. The garnet is fine grained and equant, and gives the rock a dominantly granular texture. It is almandite-spessartite in composition, for it contains about 12 percent manganese (quantitative spectrographic analysis by R. G. Havens), larger amounts of iron, aluminum, and silicon, less than 5 percent calcium, and less than 1 percent magnesium (semiquantitative spectrographic analysis by R. G. Havens). The lattice dimension ($a_0=11.614\pm0.002$ angstroms; analysis by A. J. Gude, 3d, U.S. Geol. Survey) is close to that of pure spessartite (Skinner, 1956, p. 429.) Megascopically the garnet is red, but in thin section it is nearly colorless. Magnetite and quartz are both anhedral and fill interstices between garnet grains. Apatite and allanite form equant grains that are disseminated through the rock.

Amphibole and orthopyroxene occur in irregular patches and layers. Two kinds of amphibole are present: (1) green hornblende having the optical properties of hastingsite and (2) a colorless amphibole having the optical properties of cummingtonite. The orthopyroxene is colorless to very pale green, shows schiller structure, and has the optical properties of hypersthene. The hastingsite is in large anhedral blades having ragged boundaries, and it poikilitically includes fine-grained garnet. The cummingtonite and hypersthene are interbladed with the hastingsite and appear to have replaced it along cleavages. In turn the cummingtonite appears to have replaced hypersthene along cleavages.

ORIGIN OF THE GNEISSIC ROCKS

The gneissic rocks, with the possible exceptions of granite gneiss and garnet-hornblende-orthopyroxene gneiss are believed to represent a thick succession of metamorphosed sedimentary rocks. The biotite gneiss and associated quartz gneiss and some types of calcium-silicate gneiss have been included in the Idaho Springs Formation and have long been recognized to be of metasedimentary origin (Ball, 1906; Bastin and Hill, 1917; and Lovering and Goddard, 1950). This conclusion is drawn from the compositions and the well-layered, apparently bedded character of the rocks.

Three interpretations of the sequence of major layers are possible: (1) the main layers of gneissic rocks are upright, the oldest sediments formed the biotite gneiss near Idaho Springs and the youngest sediments formed the rocks that cap Bellevue and Pewabic Mountains. This interpretation of the stratigraphic succession is reinforced by the presence of quartz gneiss and garnetiferous sillimanitic biotite-quartz gneiss in the unit near Idaho Springs and their lack of repetition in other units. As discussed later, however, this evidence is not conclusive. (2) The apparent succession may represent two limbs of a large, nearly isoclinal overturned anticline that plunges gently southwest. If so, the oldest sediments formed the unit of biotite gneiss between the Quartz Hill and Big Five layers of microcline gneiss, and the youngest sediments formed the biotite gneiss units near Idaho Springs and on Bellevue and Pewabic Mountains. (3) It is further possible that the Quartz Hill layer was sheared off where it pinches out on the southern side of the district. This shearing could have cut across an upright succession at a low angle and produced a repetition by faulting, or it could have been subparallel to the axial plane of a large overturned fold.

The original sediments were metamorphosed at high temperatures and pressures and now have mineral assemblages that are consistent with the upper range of the almandine-amphibolite facies, as defined by Fyfe, Turner, and Verhoogen (1958, p. 231), and probably with the sillimanite-almandine-orthoclase subfacies of Turner and Verhoogen (1960, p. 549). The common association of sillimanite and microcline suggests that the high-temperature conditions under which muscovite and quartz react to form sillimanite, microcline, and water were reached at one time. However, the fact that muscovite is present in nearly all specimens of sillimanitic biotite-quartz gneiss suggest either that equilibrium was not reached at that high temperature or that the reaction has since reversed. The petrographic evidence is not conclusive, but it is unlikely that sillimanite formed as a result of any other reaction, and it is most likely that in pervasive regional metamorphism equilibrium was attained, that is, that all the muscovite formed during prograde metamorphism was converted to sillimanite and microcline at the highest grade. The age relations between muscovite and sillimanite are not obvious, but the textural differences of muscovite, biotite, and feldspars suggest that muscovite formed later. These relations suggest that the muscovite is largely a product of retrograde metamorphism. Other evidence for retrogression is seen possibly in the sparse relict grains of hornblende in the microcline-quartz-plagioclase-biotite gneiss and possibly in the replacement of amphibolite by calcium-silicate gneiss.

The biotite gneiss probably is metamorphosed interbedded sandstone and mudstone. The biotite-quartz-plagioclase gneiss layers probably were sandstone beds and, to judge from their present mineral composition, these beds were variable in composition and not particularly quartzose. The sillimanitic biotite-quartz gneiss layers, on the other hand, probably were mudstone beds. The presence of sillimanite indicates that they were aluminum-rich, as is characteristic of clay-rich mudstone. As described previously, contacts between layers of biotite-quartz-plagioclase gneiss and sillimanitic biotite-quartz gneiss commonly are gradational, and some contacts are suggestive of graded bedding.

As indicated by Harrison and Wells (1959, p. 9), an increase in the iron content in the original sediments would favor the formation of almandine garnet and a deficiency of iron would favor potassium feldspar. This relation may explain the garnetiferous varieties of biotite-quartz-plagioclase gneiss and sillimanitic biotite-quartz gneiss, and it may explain the biotite-poor and microcline-rich variety of sillimanitic biotite-quartz gneiss (table 3, specimen Z–141–b).

The quartz gneiss layers probably represent sandstone beds that were much more quartzose than those that formed biotite-quartz-plagioclase gneiss. Such thin and extensive layers are difficult to explain otherwise. The local association of quartz gneiss with garnetiferous sillimanitic biotite-quartz gneiss is particularly interesting. This association may mean that the original sediments near Idaho Springs were better sorted into quartz-rich and iron-and-clay-rich layers than those that formed the biotite gneiss elsewhere.

The microcline gneiss is less well layered, lighter in color, and more homogeneous than the biotite gneiss, and its composition is similar to quartz monzonite and granodiorite. For these reasons, Lovering and Goddard (1950, p. 23) considered this rock, which they called quartz monzonite gneiss, to be igneous in origin. Available evidence does not preclude this possibility but does favor a metasedimentary origin. Though layering is less pronounced than in the biotite gneiss, a gneissic structure produced by fine laminations is conspicuous in the microcline gneiss. The composition of the unit is variable within short distances—a feature that gives a crudely layered appearance to the rock. Further, the two main layers have conformable contacts, and they contain many conformable layers and lenses of amphibolite and interlayered biotite gneiss. These features are more reasonably attributed to sedimentary than to igneous processes. Unless the rock was greatly modified in composition by metasomatic processes, it may have formed by the recrystallization of arkose.

Because the amphibolite has the composition of andesite or basalt, Lovering and Goddard (1950, p. 20) postulated that it largely represents metamorphosed flows or sills of basic composition. The lack of discordant contacts makes an intrusive origin unlikely. The amphibolite may represent metamorphosed impure dolomitic sediments or sediments rich in mafic volcanic detritus, as postulated by Harrison and Wells (1959, p. 10).

The calcic biotite-quartz-plagioclase gneiss is similar to biotite-quartz-plagioclase gneiss. The chief differences are that the calcic biotite-quartz-plagioclase gneiss contains hornblende and a much more calcic plagioclase; this gneiss may have been derived from a calcareous sandstone.

The origin of calcium-silicate gneiss is poorly understood. Some is an alteration product of amphibolite, as indicated by abundant field and petrographic evidence. During alteration, calcium garnet, epidote, and quartz appear to have formed at the expense of hornblende and plagioclase. The abundance of epidote along or near shear surfaces in specimens of amphibolite indicates that this alteration may have taken place during the younger Precambrian deformation. The calcium-silicate gneiss on Pewabic Mountain contains diopside, calcium-rich scapolite, epidote, arfvedsonite, quartz, oligoclase, and microcline. It may represent metamorphosed impure limestone. Conceivably it is an alteration product of amphibolite but, if so, no amphibolite remains.

The granite gneiss may have formed by many processes. This rock is intimately mixed with other rock types, particularly biotite gneiss, and it increases in abundance southwestward across the district, apparently at the expense of the biotite gneiss. These features, combined with observations at outcrops, suggest that the granite gneiss formed largely by replacement of biotite gneiss. However, the evidence does not preclude the possibilities that the southwestward increase in abundance of granite gneiss represents an original sedimentary facies change or that the rock was introduced from a magmatic source. The granite gneiss formed partly before or during the older Precambrian deformation, for it is cut by granodiorite and has been folded. Some formed later or was remobilized, for it locally forms phacoliths in the crests of younger Precambrian folds.

Some of the thin leucocratic microcline-rich layers in the sillimanitic biotite-quartz gneiss may have formed by the reaction of quartz and muscovite to sillimanite and microcline during high-grade metamorphism. This reaction would not tend to form material of granitic composition because it uses quartz, but inasmuch as water is evolved in the reaction, an aqueous solution may have formed that was capable of redistributing quartz and alkali feldspars. Accordingly, the microcline formed by the reaction, and some of the feldspars and quartz originally in the rock may have segregated into discrete layers along foliation planes.

Too little of the garnet-hornblende-orthopyroxene gneiss is exposed in the district to permit speculation on its origin.

GRANITIC ROCKS

Three varieties of granitic rocks—granodiorite, quartz diorite, and biotite-muscovite granite—are exposed in the Idaho Springs district. They form small bodies that are generally concordant but in detail locally cut across the gneissic structure and layering of the enclosing rocks. All three varieties are more abundantly exposed in the Chicago Creek area than in the Idaho Springs district, and comprehensive descriptions of them are given by Harrison and Wells (1959, p. 12–20).

GRANODIORITE AND QUARTZ DIORITE

A few small bodies of granodiorite, probably equivalent to the Boulder Creek Granite of Lovering and Goddard (1950, pl. 2), are exposed in the southeastern part of the area. The most accessible exposure is on the end of the ridge just south of the mouth of Spring Gulch. Quartz diorite forms somewhat larger bodies in the same general area. Good exposures of quartz diorite can be seen in Spring Gulch and on the north side of Clear Creek about 3,300 feet east of the portal of the Big Five tunnel. These two rocks are described together because their structural relations to the country rocks are similar, and they are closely related in space and time.

Granodiorite and quartz diorite are massive to well foliated, medium to coarse grained, and medium to dark gray. In the field, granodiorite is distinguished from quartz diorite by its distinctly lighter color; the massive parts of the quartz diorite have rough, knobby fracture surfaces due to the presence of clots of blocky pyroxene and amphibole crystals. Otherwise the two rocks are similar and form small, generally lenticular and concordant bodies, the central parts of which are massive and the borders of which are well foliated and lineated. In the Chicago Creek area, dikes of quartz diorite cut bodies of granodiorite (Harrison and Wells, 1959, p. 15).

Foliation in the borders of granodiorite and quartz diorite bodies is produced by a faint compositional layering and by subparallel alinement of biotite. It is generally conformable to the gneissic structure in the wallrocks and locally crosses discordant contacts, as shown by Harrison and Wells (1959, fig. 4). Lineation is produced by elongation of biotite aggregates and quartz-feldspar aggregates. It plunges moderately north-northeast parallel to the folds of the older Precambrian deformation in the enclosing rocks. (See p. A39–A90.) In addition one body of quartz diorite contains small folded dikes of light-colored coarse-grained quartz diorite. The folds also plunge moderately north-northeast. Harrison and Wells (1959, p. 12, 13) clearly demonstrated the relation of lineation in the deformed border phase of the granodiorite to that in the enclosing rocks.

Both granodiorite and quartz diorite contain dikes of light-colored coarse-grained to pegmatitic rock. These dikes have the composition of quartz monzonite to quartz diorite and contain randomly oriented books of biotite. The dikes are thought to be genetically related to the granodiorite and quartz diorite, for they are more abundant within these rocks than in the country rocks, and they have been affected by folds having the same bearing as the lineations in the granodiorite and quartz diorite.

Granodiorite is gray, fine grained, well foliated to medium grained and massive. In the Chicago Creek area it ranges in composition from quartz monzonite to granodiorite, but has the average composition of granodiorite (Harrison and Wells, 1959, p. 12). The two specimens from the Idaho Springs district that were studied (table 8, specimens MY–919, and MY–997) contain only small amounts of quartz, slightly more plagioclase (dominantly andesine) than microcline, and about 25 percent biotite. Sphene is distinctly more abundant than in other rock types of the district and is one of the characteristic accessory minerals of granodiorite in the Chicago Creek-Idaho Springs-Central City area. Magnetite and apatite are also abundant, and trace amounts of zircon were seen. Epidote and white mica are alteration products.

Viewed in thin section the texture of one specimen (MY–997, table 8) is fine grained and granular. Quartz and feldspar form a granoblastic groundmass, and biotite is in randomly oriented books that tend to form aggregates. Plagioclase is weakly zoned and is twinned according to the albite and pericline laws. Magnetite, sphene, and apatite form disseminated equant subhedral to anhedral grains; locally sphene forms rims around magnetite. The other specimen (MY–919) shows a cataclastic texture that is characteristic of the zone of intense younger Precambrian deformation. (See p. A54–A55.)

Quartz diorite is dark gray, fine to coarse grained, and well foliated to massive; the finer grained specimens show the best foliation. Some specimens of massive rock have a mottled appearance due to clots of blocky hornblende and pyroxene and interstitial plagioclase and quartz. Though microcline is locally more abundant than plagioclase, it is sparse or absent in most specimens (table 8; Harrison and Wells, 1959, table 8). Quartz forms 5 to nearly 20 percent of the massive rocks, plagioclase 20 to 30 percent, and mafic minerals (augite, hornblende, biotite, and magnetite) more than 45 percent. Only the mas-

TABLE 8.—*Modes (volume percent) of granodiorite and quartz diorite*

Mineral	Mode of specimen							
	Granodiorite		Quartz diorite					
	MY–919	MY–997	MY–521–4[1]	MY–521–3[1]	MY–521–1	MY–917–2[1]	MY–917–1[1]	MX–849
Quartz	14	10	5	10	33	5	5	19
Plagioclase	29	34	20	42	53	20	10	30
Microcline	26	22	15	0	0	0	15	0
Augite	0	0	Trace.	0	0	25	0	8
Hornblende	0	0	40	15	0	25	30	38
Biotite	24	26	15	21	14	20	30	0
Sphene	4	3	3	1	Trace	Trace	5	1
Magnetite	2	3	0	10	Trace	0	0	Trace
Pyrite	0	0	1	0	Trace	2	1	0
Apatite	1	2	1	1	Trace	3	3	Trace
Zircon	Trace	Trace	Trace	Trace	Trace	Trace	0	0
Epidote	Trace	Trace	0	0	Trace	0	0	4
White mica	Trace	Trace	0	Trace	0	0	0	0
Allanite	0	0	0	Trace	0	Trace	1	0
Plagioclase composition[2]	Andesine	Oligoclase to andesine.	Andesine	Andesine	Oligoclase.	Andesine	Oligoclase.	Labradorite.

[1] Approximate modes by visual estimation.
[2] Approximate plagioclase compositions estimated from optic sign, relative refractive indices and maximum $X' \wedge (010)$ obtained without use of universal stage.

Specimen listings:

MY–919: Small granodiorite body exposed in stream bed 1,300 ft upstream from mouth of Spring Gulch; specimen is granulated, well foliated.
MY–997: Small granodiorite body exposed on end of ridge at mouth of Spring Gulch; specimen is fine grained, faintly foliated, but not granulated.
MY–521–4: Lenticular body of quartz diorite on north side of Clear Creek 3,300 ft east of portal of Big Five tunnel; specimen is massive medium-grained rock from center of body.
MY–521–3: Same body as MY–521–4; specimen is fine-to medium-grained rock from border of body.
MY–521–1: Same body as MY–521–4; specimen is from small coarse-grained folded dike, fold plunges steeply north-northeast.
MY–917–2: Lenticular body of quartz diorite 900 ft upstream from mouth of Spring Gulch; specimen is massive medium-grained rock from center of body.
MY–917–1: Same body as MY–917–2; specimen is foliated fine-grained rock from border of body.
MX–849: Phacolithic body of massive medium-grained quartz diorite 3 ft. wide, 2 ft thick, exposed 1,500 ft N. 31° W. of portal of Big Five tunnel; fold plunges moderately north-northeast.

sive varieties contain augite. Well-foliated rocks are more variable in composition and generally contain more biotite than the massive rocks. The plagioclase of massive varieties is typically andesine, but one specimen (MY–849, table 8), which contains labradorite, has a composition of quartz gabbro. In foliated varieties the plagioclase is andesine or oligoclase, and in the pegmatitic light-colored dikes the plagioclase is oligoclase. Sphene, apatite, magnetite, pyrite, allanite, and zircon are common accessory minerals. Magnetite and pyrite generally do not occur together. Epidote and white mica are alteration products.

Viewed in thin section, the massive quartz diorite has a variable grain size (seriate texture) and consists of blocky augite and hornblende containing interstitial quartz and plagioclase. Hornblende embays augite along cleavages, and the augite is present only in relict grains having ragged boundaries. Biotite is in books having ragged borders and may have formed at the expense of hornblende. Plagioclase is weakly zoned to unzoned and shows complex twinning according to the albite and pericline laws. In foliated rocks biotite flakes tend to be subparallel and the light and dark minerals tend to be segregated into layers.

BIOTITE-MUSCOVITE GRANITE

Biotite-muscovite granite, equivalent to the Silver Plume Granite of Ball (1906), is in many small bodies throughout the district, but it forms only a small percentage of the bedrock. The largest bodies are on the western side of the district, and the most accessible exposure of a large body is on the north side of Trail Creek in the extreme southwest corner of the district.

The largest bodies of biotite-muscovite granite are thin lenticular concordant sheets; many smaller sheets and phacoliths and a few thin dikes are exposed also. One sheet on the southwest side of Bellevue Mountain has an exposed length of about 1,500 feet and a maximum thickness of about 50 feet. Another, in the extreme southwest corner of the district, has an exposed length of about 3,100 feet and a maximum thickness of more than 100 feet. The phacoliths are much smaller and occupy the crests of folds that plunge moderately to steeply north-northeast. The contacts of the sheets and phacoliths are generally concordant, but in detail they are locally discordant. The contacts are generally sharp, but they are commonly lined with a thin layer of leucocratic pegmatite.

The granite is typically light gray to pink, fine to medium grained, homogeneous, and massive to faintly foliated. Thin tabular crystals of feldspar about a centimeter long can be seen in most exposures.

Commonly, particularly near contacts, the rock has a foliation produced by the subparallel alinement of the tabular feldspar crystals and small biotite plates. This foliation is nearly parallel to the contacts of the biotite-muscovite granite, even where the contacts are discordant, and is interpreted to be a primary flow structure.

The biotite-muscovite granite is composed principally of quartz, microcline, plagioclase, and smaller amounts of biotite and muscovite (table 9). As shown on figure 5, the proportions of quartz, microcline, and plagioclase are remarkably uniform; this diagram closely resembles one of Harrison and Wells (1959, p. 18). With the exception of a small amount of albite associated with perthitic microcline, the plagioclase is largely oligoclase. Zircon, monazite, and magnetite are common accessory minerals.

Viewed in thin section, the biotite-muscovite granite is anhedral to subhedral, granular and equigranular to seriate. Microcline commonly forms anhedral or subhedral tabular crystals that exhibit carls-

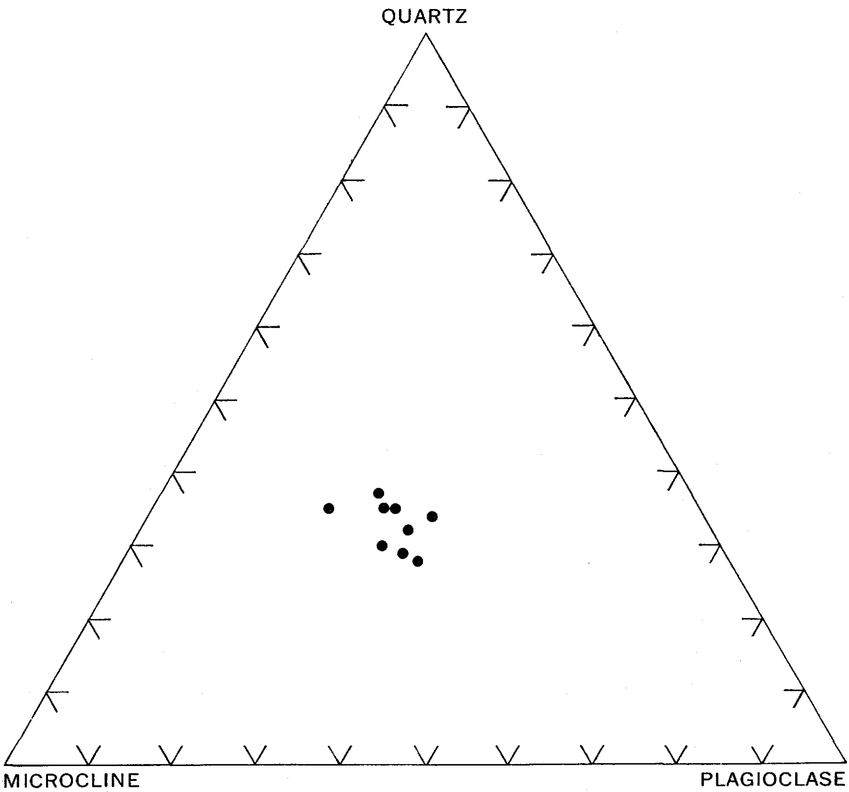

FIGURE 5.—Triangular diagram showing variation in composition (volume percent) of the biotite-muscovite granite.

TABLE 9.—*Modes (volume percent) of biotite-muscovite granite*

Mineral	Mode of specimen								
	MX-990	MX-800-1	MX-391	MX-267	MX-89-2	Z-129	L-X-7	MX-930-1	3-1-74
Quartz	32	30	34	32	28	26	32	32	26
Microcline	42	33	34	34	38	34	29	34	34
Plagioclase [1]	20	30	23	25	28	32	32	26	30
Biotite	1	3	4	5	3	Trace	2	0	7
Muscovite	5	2	4	4	3	7	5	8	3
Zircon	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace	Trace
Monazite	Trace	Trace	Trace	0	Trace	Trace	Trace	Trace	0
Magnetite	Trace	2	1	Trace	Trace	1	Trace	Trace	Trace
Apatite	0	0	0	0	0	Trace	Trace	0	Trace
Rutile	0	0	0	0	0	0	Trace	0	0
Allanite	0	0	0	0	0	0	0	Trace	Trace
Chlorite	Trace	0	0	0	0	0	0	0	Trace

[1] Plagioclase largely oligoclase, excepting albite associated with microcline.

Specimen listings:

MX-990: Slightly granulated rock from phacolith on south side of knoll 3,800 ft N. 32° E. of portal of Big Five tunnel.
MX-800-1: Ungranulated rock from small lens on west side of gully 1,400 ft S. 72° W. of summit of Bellevue Mountain.
MX-391: Slightly granulated rock from lens 1,000 ft due east of summit of Bellevue Mountain.
MX-267: Slightly granulated rock from lens 2,000 ft due east of mouth of Trail Creek.
MX-89-2: Ungranulated rock from south end of large lens 3,000 ft S. 60° W. of summit of Bellevue Mountain.
Z-129: Intensely granulated rock from thin lens on ridge 1,500 ft S. 34° W. of portal of Big Five tunnel.
L-X-7: Slightly granulated rock from thin dike exposed in crosscut connecting workings on Alma Lincoln and Elliot and Barber veins (south side of Clear Creek Canyon, about 1,000 ft east of mouth of Trail Creek).
MX-930-1: Granulated rock from small body 3,700 ft S. 23° E. of summit of Bellevue Mountain.
3-1-74: Granulated rock from thin lens 1,300 ft west of mouth of Trail Creek.

bad twinning, as well as the usual twinning network. The tabular crystals, as much as a centimeter long, are embedded in a groundmass of quartz, plagioclase, microcline, biotite, and muscovite. The microcline contains thin films of perthite and rounded blebs of quartz. Plagioclase is weakly zoned to unzoned oligoclase; it forms equant grains that are commonly coated with thin, sharply defined rims of albite, particularly along plagioclase-microcline boundaries. Quartz also forms equant grains, but it tends to be interstitial to plagioclase and microcline. Biotite, which is locally partly chloritized, forms sharply defined books and is commonly interleaved with muscovite. Muscovite also forms irregular plates that poikilitically include quartz and altered plagioclase. Zircon, monazite, and magnetite—the most commonly observed accessory minerals—occur in small disseminated euhedral to subhedral grains. In many specimens the rock texture is disrupted by granulation and mortar structures that are characteristic of the area affected by intense younger Precambrian deformation (p. A54–A55).

ORIGIN OF THE GRANITIC ROCKS

It is generally agreed that the granodiorite, quartz diorite, and biotite-muscovite granite are of magmatic origin (Lovering and Goddard, 1950, p. 25–29; Harrison and Wells, 1959, p. 12–20). The rocks are fairly homogeneous, having sharp and locally discordant contacts, and the granodiorite and biotite-muscovite granite have a primary flow structure that is parallel to discordant contacts. These rocks are believed to have been emplaced in the order listed above during the older period of Precambrian deformation.

The granodiorite and quartz diorite were intruded sufficiently early in the first period of deformation to have been deformed and metamorphosed. Both rocks form generally conformable bodies that are characteristic of syntectonic intrusive rocks, and near their borders both have a gneissic structure. As shown by Harrison and Wells (1959, p. 12, 13), this foliation crosses discordant contacts of the granodiorite and is continuous with the gneissic structure in the country rocks. In addition, the dominant lineations in the granodiorite are parallel to the lineations in the country rocks that formed during the older Precambrian deformation. Dikes of quartz diorite cross the granodiorite (Harrison and Wells, 1959, p. 15), but they contain a similar gneissic structure. The largest quartz diorite bodies in the Idaho Springs district contain thin pegmatitic dikes that were folded along north-northeast-trending axes of folds that are related to the older period of Precambrian deformation.

The biotite-muscovite granite was probably emplaced near the end of the older deformation, or possibly later. Unlike the granodiorite and quartz diorite, the biotite-muscovite granite does not have a metamorphic foliation or gneissic structure, and it forms dikes that cross the earlier igneous rocks as well as folds of the older deformation (Harrison and Wells, 1959, p. 17–20). These relations suggest that the biotite-muscovite granite is post-tectonic. On the other hand, it forms phacoliths in the crests of north-northeast-trending folds; this relationship suggests that stresses of the older deformation were still active when the magma was emplaced.

All the granitic rocks are locally cataclastically deformed—a fact that supports the interpretation that they were emplaced earlier than the younger period of Precambrian deformation.

PEGMATITIC ROCKS

A few small bodies of pegmatite are shown on plate 1. These rocks were not studied in detail and little can be said of their origin. In general they are composed largely of quartz and feldspar; some varieties contain abundant biotite.

The pegmatite dikes in and near the bodies of granodiorite and quartz diorite are lighter in color than these rocks and are thought to be late-stage products of them.

One pegmatite dike, exposed about 1,500 feet southwest of the portal of the Big Five tunnel, crosses the layering and gneissic structure of the country rocks at a low angle. It is coarse grained and light colored, and shows no signs of having been deformed. Inasmuch as it is in an area of intense younger Precambrian deformation, it probably was emplaced later than that deformation.

One variety of pegmatite is distinguishable by the presence of randomly oriented books of biotite and of small disseminated octahedral crystals of uraninite. This variety is widespread, but it forms bodies that are too small to show on plate 1. An unsuccessful attempt, described by Sims and others (1963, p. 10–12), has been made to mine this rock for its uranium content at the Hudson tunnel in Virginia Canyon.

The biotite-uraninite-bearing pegmatite may be related to the biotite-muscovite granite, for the structural relations of these rocks are similar. The absence of a gneissic structure in the pegmatite suggests that it formed during or later than the last stage of the older Precambrian deformation, but it is locally cataclastically deformed and contains lineations that are related to the younger Precambrian deformation. Both the pegmatite and the biotite-muscovite granite have high radioactivities (Harrison and Wells, 1959, p. 19).

TERTIARY AND QUATERNARY ROCKS

TERTIARY INTRUSIVE ROCKS

The Idaho Springs district has been intruded by a network of igneous dikes of Tertiary age and by several plutons of irregular outline (pl. 1). The dikes follow many preexisting planes of weakness, such as joints, faults, and foliation in the Precambrian rocks, and other features. The plutons are concentrated in the northeast corner of the district, and local irregular patterns of dikes suggest that other plutons may exist at depth. Some of the plutons may be the eroded throats of volcanoes.

The rocks were emplaced in a sequence that exhibits changing habits of the intrusive bodies, as well as changing composition (Wells, 1961). With few exceptions, the older rocks of the sequence form plutons, thick dikes, and some thick lenticular concordant masses, whereas the younger rock types form thin but long dikes.

Nine varieties of Tertiary intrusive rocks have been identified in the Idaho Springs district including, from oldest to youngest, leucocratic granodiorite porphyry, albite granodiorite porphyry, alkali syenite porphyry, quartz monzonite porphyry, granite porphyry, bostonite porphyry, quartz bostonite porphyry, trachytic granite porphyry, and biotite-quartz latite. They are not classified on plate 1. These rocks were emplaced during the Laramide orogeny in early Tertiary time (Lovering and Goddard, 1950, p. 47). All except the biotite-quartz latite were emplaced prior to the formation of the metalliferous veins of the district, which took place 57 to 70 million years ago, according to lead-uranium determinations (Holmes, 1946; Phair, 1952; Faul and others, 1954, p. 263). There is no reason to suspect that the biotite-quartz latite was emplaced much later than the formation of the veins or the earlier intrusive rocks. The sequence of emplacement was determined by study of crosscutting relations in the Idaho Springs and adjoining areas.

Detailed descriptions of the Tertiary intrusive rocks are given by Wells (1960) and are not repeated here.

VEINS

The veins of the district typically fill fault fissures and are similar in structure and mineralogy to those classed as mesothermal by Lindgren (1933, p. 529). Most of the veins strike northeast to east and dip north to northwest, at medium to high angles, and a few strike northwest and dip northeast at medium angles. Displacements along the faults filled with the vein minerals are largely strike-slip; they are typically only a foot or two and uncommonly as much as 80 feet. The Idaho Springs fault, expressed on the surface as a long northwest-trending vein (pl. 1), shows an exceptionally large apparent strike-slip displacement of about 600 feet.

The principal vein minerals are pyrite, sphalerite, galena, chalcopyrite, tennantite, quartz, and local carbonate minerals. These minerals are intergrown and solidly fill most of the fissures. In places the vein minerals fill a single fissure and form symmetrically banded structures characterized by a band of intergrown base-metal minerals bordered on either side by bands of intergrown pyrite and quartz. Such veins are rarely more than a foot wide. More commonly the veins are characterized by a zone of many close-spaced anastomosing, interconnecting fractures that have been cemented with ore and gangue minerals. In such veins, fractured, altered, and pyritized rock is traversed by innumerable veinlets of quartz and pyrite, which are cut locally by pods and veinlets of base metal minerals. Such veins are as much as 30 feet wide but are typically less than 5 feet wide.

Since the discovery of placer gold in Idaho Springs and lode gold in Central City in 1859, ores valued at about $50 million have been mined from the veins of the Idaho Springs district. The principal ore metal is gold, but in some areas silver predominates, and in most areas the ore contains subsidiary copper, lead, and zinc.

Detailed descriptions of the veins will be given in a report being prepared by R. H. Moench and A. A. Drake, Jr., and are not repeated here.

QUATERNARY DEPOSITS

The Quaternary deposits are alluvium and colluvial creep debris and talus. Creep debris is widespread but was mapped only where it completely covers broad areas. The debris sheets rarely exceed 10 feet in thickness, but they effectively cover large areas of bedrock. They are composed of a heterogeneous mixture of angular rock fragments and fine-grained material, some of which has moved downhill a considerable distance. Ridges of creep debris as high as 20 feet are common in many gullies that are flanked by debris sheets and are probably the result of pressures developed from the persistent downhill creep on either side. Talus is common on steep slopes below cliffs.

Quaternary alluvium covers the floor of Clear Creek Canyon and parts of the valleys of Trail Creek and Spring Gulch; it is locally present on terraces that are well above Clear Creek. The alluvium at the present drainage levels consists of fine to coarse gravel, some of which is locally derived and some of which is derived from several miles upstream. Spurr, Garrey, and Ball (1908, p. 83, 84) noted three sets of terraces near Idaho Springs. The terraces are cut in bedrock at about 160 feet, 55 feet, and 25 feet, respectively, above Clear Creek; the two higher ones are capped by about 20 feet of gravel, and the lower is capped by about 5 feet of gravel. The terrace gravel is fine to coarse and contains well-rounded boulders and cobbles.

Some of the gravel on the terraces or on the present valley floors may have been deposited in Pleistocene time by melt water from valley glaciers, which are known to have existed in the headwaters of Clear Creek and some of its tributaries. Aside from the gravel deposits, no evidence of glaciation has been recognized in the Idaho Springs district.

PRECAMBRIAN STRUCTURE

Two episodes of Precambrian deformation are recognized in the Idaho Springs district, an older deep-seated plastic deformation that took place under powerful confining pressures and a younger cataclastic deformation at shallower depths.

The older deformation was pervasive and resulted in generally broad open folds that define the structural framework of the district. Concurrent with this deformation, the rocks were thoroughly recrystallized and, with the possible exception of the biotite-muscovite granite, the succession of Precambrian intrusive rocks was emplaced. Lineations were formed in the metamorphic rocks, either parallel to the major fold axes (B_0) or nearly at right angles to them (A_0).

The younger deformation, which formed as part of the extensive

Idaho Springs-Ralston shear zone (Tweto and Sims, 1960), was mainly restricted to a zone about 2 miles wide along the southeast side of the district. It resulted in cataclasis and some small-scale crossfolds, which are superposed on the older folds, and in two new sets of lineations, one (B_y) parallel to the axes of the younger folds and the other (A_y) inclined at a large angle to them. At places a new foliation consisting of closely spaced shear planes was formed, but most shearing related to the deformation took place along original foliation surfaces.

Some features in the Idaho Springs district could be interpreted to indicate the existence of a fold system still older than the older of the two recognized systems. As shown on plate 1, the Big Five layer of microcline gneiss on the west limb of the Idaho Springs anticline converges southwestward with the Quartz Hill layer of the same rock type. This convergence is interpreted tentatively as being the result of stratigraphic pinchouts of the two layers as well as of the intervening unit of biotite gneiss. Alternatively, the convergence of the two layers of microcline gneiss could reflect an overturned anticline that plunges gently southwestward. A southwest plunge would be mostly likely because the Big Five layer strikes generally more to the east and dips more steeply than the Quartz Hill layer, and their line of intersection would plunge gently southwestward. With this interpretation, the Big Five layer would be an overturned equivalent of the Quartz Hill layer, and the intervening biotite gneiss would be in the core of the anticline and, hence, stratigraphically lower than the biotite gneiss exposed near Idaho Springs and on Bellevue Mountain. If such a fold exists, the linear elements related to it were completely obliterated by the plastic deformation that created the older of the two recognized fold systems. The unit of biotite gneiss between the Quartz Hill and Big Five layers extends eastward into the Ralston Buttes quadrangle where D. M. Sheridan (written communication, 1961) interprets the unit as the core of a large fold that plunges gently southwestward; he infers that this fold represents the oldest of three deformations. Wells and others (1961) also postulated that the rocks in the same general area were deformed three times in the Precambrian.

It is further possible that the southwestward pinchout and convergence of the Big Five and Quartz Hill layers of microcline gneiss are related to shearing along the Idaho Springs-Ralston shear zone. As recognized by Tweto and Sims (1963, p. 998), this shear zone is a major structural discontinuity in the Front Range. It may have had its inception during or even before the older Precambrian deformation, and the recognized cataclasis and folding of the younger deformation may only express the latest of several recurrent movements in the shear zone. If so, the pinchout and convergence of major rock units may have resulted from early movements along the zone.

Neither three Precambrian deformations nor the earlier inception of the Idaho Springs-Ralston shear zone can be proved in the Idaho Springs district, and further discussion will be confined to the recognized older and younger Precambrian deformations.

OLDER DEFORMATION

The most conspicuous structural features of the district are folds that were formed during the older deformation. These folds, which trend north-northeast, are outlined by the major gneissic rock units and define the structural framework of the district. Small folds and lineations of many types are abundant. Most of these are almost parallel to the major fold axes (B_o), but some are oriented nearly at right angles to the fold axes (A_o).

FOLDS

The Idaho Springs anticline (pl. 1) is the dominating structure in the southeast part of the district and appears to mark the boundary between contrasting structural trends in a large part of the Front Range (See Lovering and Goddard, 1950, pl. 2.) For several miles east of the anticline, the layered rocks strike dominantly northwest, whereas west of the anticline the rocks strike dominantly northeast. The Idaho Springs anticline, as it appears on plate 2 of Lovering and Goddard (1950), is alined with an elongate mass of Boulder Creek Granite that extends north-eastward from the batholith of Boulder Creek Granite that is exposed to the south of the district. Within the district, several bodies of granodiorite and related quartz diorite are exposed near the crest and on the east limb of the anticline—a fact that suggests that the elongate mass may extend beneath the district along the anticline.

Within the district, the Idaho Springs anticline is open and upright, and its axis trends sinuously northeast. The fold is strongly asymmetric. (See pl. 2, section *C–C′*, particularly.) Its northwest limb strikes northeast and dips steeply northwest (about 50°–90°), its southeast limb strikes west to northwest and dips moderately (about 30°–50°) north and northeast. The axial plane of the fold is steep, and the axis plunges moderately about N. 30° E. Near Idaho Springs the anticline is well defined and has sharply opposing limbs, but northeastward it loses its definition. In the Quartz Hill layer of microcline gneiss the fold is shown only by the broadly arcuate southern contact of that unit. The thickness of the rocks seems to be greater

735–934—64——4

on the east limb than on the west limb. This relation can be seen in the eastward thickening of the Big Five and Quartz Hill layers of microcline gneiss and in the numerous conformable bodies of quartz diorite and granodiorite near the crest and on the east limb of the fold (pl. 1).

With local exceptions, dips decrease gradually northwestward from the crest of the Idaho Springs anticline to the Pewabic Mountain and Trail Creek synclines—distances of about 7,000 and 10,000 feet, respectively (pl. 2, sections *C–C′*, *A–A′*). These synclines are largely broad, open structures of low relief and sinuous north-northeast trend. They mark the boundary between generally steeply dipping rocks to the east and comparatively flat-lying rocks to the west. Several north-northeast-trending folds of low amplitude are exposed northwest of the Pewabic Mountain syncline. Most important of these is the Central City anticline, which expands greatly northward and becomes the dominating anticline in the Central City district. (See Moench, Harrison, and Sims, 1962, pls. 1 and 2.)

Several folds mapped in the Bellevue Mountain area are strongly disharmonic and become tighter and overturned along the trace of their axial planes. Such changes are thought to result largely from differences in relative competency between major lithologic units. The Pewabic Mountain syncline, for example, is an open fold of low amplitude where the Quartz Hill layer of microcline gneiss is involved. This fold extends upward into the thick unit of relatively incompetent biotite gneiss and becomes closed and overturned to the southeast. Many small-scale examples of similar disharmonic folds have been seen in outcrops and hand specimens. Some small disharmonic folds show far more extreme changes in amplitude than are know to occur on larger structures and illustrate the contrasting behavior of different materials during the older deformation.

A large closed fold near the east side of the district is shown by the map pattern of a body of biotite gneiss in the Quartz Hill layer of microcline gneiss (pl. 1). This fold is isoclinal, for both limbs strike nearly east and dip about 65° N. If the southern limb is upright, the fold is an overturned syncline. The trace of its axial plane trends generally east, but small drag folds and lineations indicate that the axis of the fold plunges about N. 10°–20° E. at about 60°. The folded layer of biotite gneiss is not sufficiently extensive to determine the dimensions of the fold, but the map pattern shows that it is a large feature.

This fold is exposed near the nose of the Idaho Springs anticline, and although the two folds probably plunge in the same direction, their respective axial planes trend at wide angles to one another (pl. 1). The axial plane of the isoclinal syncline, in fact, probably overlaps both limbs of the Idaho Springs anticline. This relation suggests that the isoclinal fold originated before the Idaho Springs anticline achieved its present form and that it was warped by the Idaho Springs anticline.

Another overturned, nearly isoclinal syncline is shown by the map pattern of a body of granite gneiss and bordering amphibolite in the middle of the southwestern end of the Quartz Hill layer about 2,500 feet northeast of the map boundary (pl. 1). Both limbs strike northeast and dip steeply northwest. Asymmetric folds shown by the map pattern have the expected asymmetry for such a syncline and indicate that it plunges almost due north 30°–55°.

Several large asymmetric folds are shown by the northwest contact of the Quartz Hill layer near the southwest side of the district between 1,000 and 3,500 feet northeast of the map boundary. These folds have long, presumably overturned southeast limbs and short northwest limbs, and their axes plunge about 35° N. 10° E. Geological Survey mapping of the Rockford tunnel, which crosses the Trail Creek syncline, indicates that these folds continue in depth and suggests that the Quartz Hill layer pinches out in depth along an axis that plunges north at 30°–55°.

Large closed folds of the older deformation were seen only in the Quartz Hill layer, but undoubtedly some are in the Big Five layer as well and in the intervening unit of biotite gneiss. On the other hand, large closed folds evidently are not in the unit of biotite gneiss southeast of the Big Five layer. Here, several thin layers of quartz gneiss (pl. 1) were traced to detect any closed isoclinal folds that might be present. Although the quartz gneiss layers outline the Idaho Springs anticline and many small folds, they do not converge or double back on a large scale.

Abundant and varied minor folds occur on the limbs of the principal folds and are generally parallel to the major fold axes. Many of these structures are small-scale replicas of the large folds in that they are open and have gently dipping limbs, but many are upright or overturned closed folds and a few are recumbent. Where asymmetric, most of the minor folds have the relation of drag folds to the next larger fold; the axial planes typically converge upward on the flanks of the major anticline. Several closed, nearly isoclinal folds having amplitudes of several tens of feet are exposed on the northwest limb of the Idaho Springs anticline in a narrow zone near the mouth of Trail Creek. Recumbent folds having axial planes subparallel to the prevailing foliation are locally common in areas of gentle dip, especially in the crests of anticlines or the troughs of synclines. Several recumbent folds can be seen in the map pattern on the Trail Creek syncline (pl. 1).

Drag folds locally indicate relative movements reversed to that produced in simple flexural slip folding. Such drag folds range in size from less than an inch to a few hundred feet across; they are sufficiently abundant in places to be useless for determining larger structure. Processes by which reversed drag folds might have formed are discussed by Moench, Harrison, and Sims (1962, p. 42).

Many small folds (A_0) were formed during the older deformation nearly at right angles to trends of the dominant folds of this deformation. These folds are mainly open, symmetrical warps of low amplitude, although tight asymmetric folds are present. The open folds are characteristic of the relatively competent microcline gneiss, whereas the tight folds are characteristic of the less competent biotite gneiss.

ASSOCIATED FOLIATION AND TEXTURES

The rocks that were not affected by the younger deformation have a foliation that is analogous to the gneissic structure that characterizes high-grade metamorphic rocks of the Precambrian shield areas of various parts of the world. The typical rock texture shows oriented plates of biotite set in an equigranular granoblastic groundmass of quartz and feldspar. Moench, Harrison, and Sims (1962, pl. 4, fig. 1) show the texture of typical microline gneiss from an area unaffected by the younger deformation. Biotite is in distinct plates that show some preferred orientation, and quartz and feldspars are nearly equigranular. Other rock types show similar textures, as described in this report.

LINEATIONS IN B_0

Lineations that parallel the major fold axes of the older fold system are dominantly axes of small folds and mineral alinements.

The folds increase in size from minute crinkles or warps and include many types, most of which are drag folds. The small folds range from open warps to tightly closed asymmetric overturned folds that are commonly strongly disharmonic; some are recumbent. Ptygmatic folds are common adjacent to pegmatite bodies.

The mineral alinements consist of the parallel alinement of elongate minerals, mostly hornblende and sillimanite, and of elongate aggregates of quartz, feldspar, and (or) biotite. These lineations are ubiquitous throughout the district and are particularly conspicuous where the rocks were not deformed by the younger fold system. Probably more than 90 percent of the mineral lineations in the region are in the B_0 direction.

All lineations measured in surface exposures were plotted on the lower hemisphere of Schmidt equal-area nets to summarize the data on lineation from the district. The resulting three diagrams, representing 3,435 readings, are shown on figure 6. Each diagram is in the

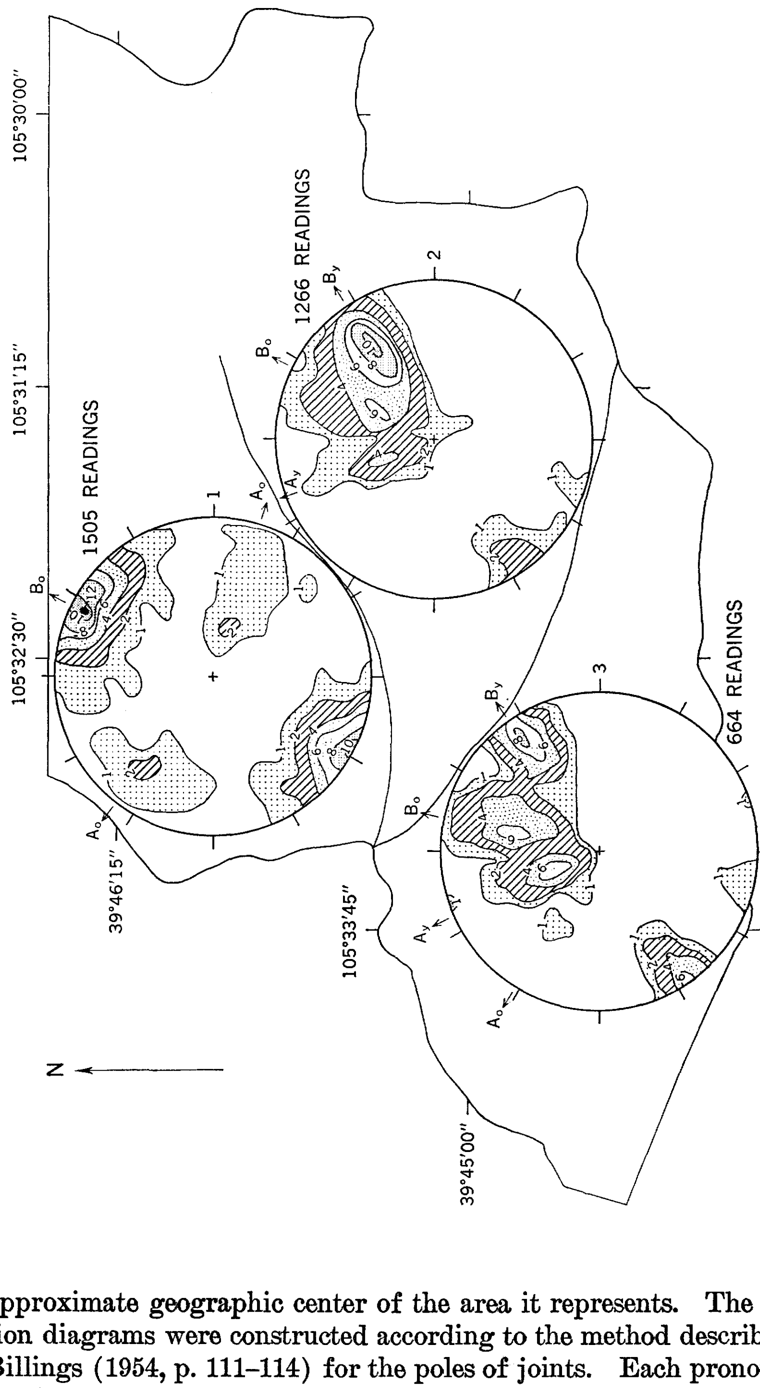

FIGURE 6.—Lineation diagrams for the Idaho Springs district.

approximate geographic center of the area it represents. The lineation diagrams were constructed according to the method described by Billings (1954, p. 111–114) for the poles of joints. Each pronounced maximum, therefore, represents the approximate bearing and plunge of many readings. Because of inherent inaccuracies due to the method of contouring (see Harrison and Moench, 1961, p. 5), the maxima may depart a few degrees from the true maxima, especially in those areas where two lineation directions only a few degrees apart are present, as in diagrams 2 and 3 of figure 6.

The lineations in B_0 are represented by strong maxima (6 and 12 percent) that range in bearing from about N. 10° E. to N. 25° E. Actually the range in bearing of individual observations in the field is much greater, from about north to east-northeast, and partly explains the broad spread of the contours. In the area of diagram 1, the average plunge of lineations in B_0 is near horizontal. In the areas of diagrams 2 and 3, the plunge of these lineations is steep to the northeast. This apparent southward steepening of plunge is consistent with the changes in plunge of the large folds (pl. 1).

LINEATIONS IN A_0

Linear elements in the A_0 direction—approximately normal to the B_0 direction—are dominantly axes of small folds but include mineral elongation and boudinage. The folds occur throughout the district, mainly as symmetrical, open warps of low amplitude and as a few asymmetric chevron folds.

Boudinage in A_0 is common. It is expressed by relatively competent layers that have been broken or necked under tension, and the surrounding less competent material has flowed inward to fill the spaces. In many boudins, pegmatite also partly fills the spaces between broken edges. Most broken or necked layers are a foot thick or less, and they have been stretched a few inches at the most. The axes of boudins, or lines along which the competent layers have broken or necked, plunge parallel to other linear elements in A_0.

Mineral alinements in the A_0 direction are rare and were seen only where A_0 folds are abundant in biotite gneiss.

Lineations in A_0 form distinct maxima only on diagram 1 of figure 6. In the area represented by diagram 1, which lies astride several open folds, lineations in A_0 exhibit two broad maxima, each of which reflects opposing limbs of the anticlines. Diagram 3 shows a 1 percent area in the northwest quadrant that represents lineations in A_0.

The folds and boudinage in A_0 probably formed during a late stage of the older deformation. The fact that they are superposed on more complex folds in B_0 indicates their relatively late origin, yet the fact that the folds in B_0 and A_0 are similar in their plastic character and are statistically about normal to one another suggests that they are genetically related. Small folds in A are common in complex metamorphic terranes (Cloos, 1946, p. 26–29), and there is no reason to suppose that the folds in A_0 in the Idaho Springs district represent a separate period of deformation.

CHARACTER OF DEFORMATION

The nature of the deformation that produced the older fold system can be deduced from the types of folds and the structure and texture of the metamorphosed rocks. The major folds trend north-northeast, and the principal stresses that formed them presumably were oriented west-northwest but were probably deflected locally by inhomogeneities in the deformed materials; marked variations in the trends and plunges of fold axes were thus produced. The small folds and boudinage in A_0 reflect alternate shortening and lengthening in the direction of the major fold axes, apparently at a late stage in the deformation. The disharmonic folds formed as a result of contrasting competency of different rock units—especially the weaker biotite gneiss and stronger microcline gneiss—and local differences in competency within units, but the simplicity of the structural framework attests the ability of the rocks to transmit stress. With the possible exception of rare ptygmatic folds in the migmatitic rocks, the characteristics of near fluidity that have been described from some other terranes (Bain, 1931) are absent. The boudins, however, attest the ductility under tension of the relatively competent microcline gneiss.

The gneissic structure of all the metasedimentary rocks, their granoblastic texture, and the ubiquitous mineral lineation were achieved during the older deformation. These features, combined with the high-grade mineral assemblages of all the rocks indicate that high temperatures and pressures prevailed.

Possibly the large syntectonic body of granodiorite whose northern margin is exposed in the southern part of the Chicago Creek area (Harrison and Wells, 1959, pl. 1) affected folding during the older deformation. The northward plunge of the older Precambrian folds steepens markedly towards this contact, from nearly horizontal in the Bellevue Mountain area to an average of more than 60° in the Chicago Creek area (Harrison and Wells, 1959, fig. 9). In addition, as described previously, there may be some relation between the Idaho Springs anticline and the "peninsula" of granodiorite that projects northward from the main batholith in the area of Mount Evans (Lovering and Goddard, 1950, pl. 2). Possibly this batholith acted as a competent buffer in the deformation, or it may have produced some of the major structures in the country rocks as it was emplaced.

YOUNGER DEFORMATION

The younger deformation recognized in this district is part of the Idaho Springs-Ralston shear zone (Tweto and Sims, 1960, p. B8–B10) which extends northeastward to the edge of the Front Range, a distance of more than 20 miles, and southwestward through the Chicago Creek area (Sims, Moench, and Harrison, 1959). As recognized by Tweto and Sims (1960; 1963, p. 998), this shear zone is a zone of possibly major structural discontinuity, the inception of which may greatly predate the recognized younger deformation. The recognized younger folds and cataclasis, in fact, may express only late Precambrian movements along the zone.

The younger deformation, manifested by cataclasis, minor folding, and by weak local crystallization, was intense in a northeast-trending 2-mile-wide zone in the southeast part of the district. Within this zone the relatively incompetent rock units are folded and sheared parallel to the fold limbs, whereas the relatively competent rock units are not folded, but are pervasively sheared. The approximate northwest margin of this zone trends northeastward through the Quartz Hill layer of microcline gneiss (pl. 1). The southeast margin is less well defined, but it probably is nearly on the southeast corner of the district and extends northeastward approximately from the mouth of Spring Gulch. The folds are small and distinctly asymmetric, the northwest sides having been raised relative to the southeast sides. They trend persistently N. 55° E., parallel to the length of the deformed zone, and plunge at various angles, depending upon their position on the older, larger folds. Associated with these folds are two lineations, one (B_y) parallel to the axes and one (A_y) oriented at a large angle to them. Northwest of the zone of intense younger deformation, effects of the younger deformation are limited to narrow zones characterized mainly by a cataclastic rock texture and by lineations of the same type and bearing as those in the main zone.

A theoretical treatment of superposed folds has been presented by Weiss (1959, p. 91–106), and its application in the Idaho Springs-Central City area is discussed in a separate paper (Moench, Harrison, and Sims, 1962). McBirney and Best (1961, p. 495–498) have experimentally produced superposed folds that are remarkably similar to the younger Precambrian folds described here.

FOLDS

Younger folds range from less than a foot to as much as 400 feet in breadth and are abundant and generally closely spaced in the zone of intense deformation. The axes of the folds trend consistently N. 55° E., and the axial planes are subparallel and dip steeply southeast to vertical. Although the folds are small, the axial planes are remarkably straight and persistent and can be traced for several thousand feet along strike. The folds tend to be sharp crested and have distinctive forms. They range from slight bends in otherwise uniformly dipping layers to complex structural terraces, closed and overturned chevron folds, and southeast-facing monoclines. All the folds are strongly asymmetric and all except the monoclines have steep, generally long northwest limbs and short crumpled southeast limbs. The younger folds are superposed on the Idaho Springs anticline, and their form depends on the geometry of superposition, combined with the type of movement that produced them. These subjects are treated in detail by Moench, Harrison, and Sims (1962) and are discussed in a later section. In general, all fold types except the monoclines occur on the northwest limb of the Idaho Springs anticline, and southeast-facing monoclines occur on the gently-dipping southeast limb.

The traces of the axial planes of several younger folds are shown on plate 1. Because the folds are narrow, only the anticlinal bends are shown. Approximate locations of the intervening synclinal bends are shown by deflections of rock layers and by the foliation symbols.

Younger folds deflect the southern contact of the Quartz Hill layer of microcline gneiss at many places (pl. 1). These folds were traced long distances through biotite gneiss and granite gneiss, but they were not traceable into the Quartz Hill layer. This unit behaved more competently than the biotite gneiss; it yielded largely by fracturing rather than folding. For example, large younger folds were mapped in the Big Five tunnel (section *D–D′*, pl. 2), and they correlate well with folds mapped at the surface. Two closely spaced folds deflect the southern contact of the Quartz Hill layer about 1,500 feet west of the tunnel, and they combine in depth to form a single large fold at the tunnel level (pl. 1). At the surface directly above this fold, however, the Quartz Hill layer, though intensely sheared, is not disoriented from a northwesterly dip. A similar prominent deflection of the same contact can be seen about 3,000 feet east of the tunnel.

On the northwest limb of the older Idaho Springs anticline the younger folds have steep, long, and fairly uniform northwest limbs that do not seem to have been rotated much from their previous attitudes. The southeast limbs, however, are extremely crumpled and are characterized by closed chevron-type crenulations. The axial planes of the mapped folds and many of the crenulations on their limbs are steep and subparallel, and can be traced across many rock layers. The general attitude of the southeast limb of a younger fold may be gentle and give the whole structure the form of a terrace, or it may dip steeply southeast and give the structure the form of an open or closed chevron fold.

On the southeast limb of the Idaho Springs anticline the younger folds commonly have the form of a southeast-facing monocline. Here the northwest limbs of the younger folds commonly strike west to northwest and dip moderately northward, as governed by the previous attitude, whereas the southeast limbs are extremely crumpled and dip generally southeastward. Examples of monoclines in the Chicago Creek area are described by Harrison and Wells (1959, p. 30, 31).

Tightly closed and locally overturned younger folds are exposed in Virginia Canyon in a zone about 1,500 feet wide on the northwest limb of the Idaho Springs anticline. (See section *C–C′*, pl. 2.) These folds have long northwest limbs and short southeast limbs, and both limbs commonly dip steeply southeast, differing in dip by as little as 10°. Their axial planes dip steeply southeast, parallel to those of more open younger folds. Such folds are the reason for the sharp serrations of the Big Five layer of microcline gneiss about 4,000 feet northeast of the portal of the Big Five tunnel and are exposed in the biotite gneiss to the northeast. Some of these folds are so tightly closed that they were first recognized only by a sharp change in the orientation of the older mineral alinements in B_0 from one limb to another. On the northwest limbs, the lineations in B_0 plunge steeply northeast, whereas on the southeast limbs, they are nearly horizontal or plunge southwest in accord with the expectable deflection of older lineations by younger folds.

In places the younger folds are disharmonic, and in general they obey the law of competency (Knopf and Ingerson, 1938, p. 161); that is, the more competent layers are more broadly arched than the incompetent layers, which are more intensely crinkled. Disharmonic folding is most pronounced where 5- or 10-foot-thick layers of massive biotite-quartz-plagioclase gneiss or quartz gneiss are interspersed in thick layers of schistose sillimanitic biotite-quartz gneiss. The competent massive layers are broadly arched, whereas the less competent schistose layers are intricately crumpled and obviously thickened in the fold crests.

Minor folds are abundant on the limbs of the larger young folds. They are generally sharp crested, and their axial planes parallel those of the larger folds. The axial planes of many of the minor folds can be traced in section across rock layers of different competencies; such folds are not true drag folds. True drag folds that are confined to a relatively incompetent layer are common, however, and indicate that successively higher layers moved towards the anticlines. Examples of typical minor younger folds are shown by Moench, Harrison, and Sims (1962).

CATACLASTIC TEXTURES

In the zone of intense younger deformation the rocks are deformed cataclastically to varying degrees. Excepting some pegmatites, all examined specimens of Precambrian rock from the zone show some evidence of cataclasis. Northwest of the boundary of the zone of nearly pervasive cataclasis (pl. 1), the rocks have been cataclastically deformed only locally.

Cataclastic deformation produced a foliation that is characterized by close-spaced subparallel anastomosing shears. Except locally, this foliation is subparallel to the previously formed gneissic structure and is nowhere parallel to the axial planes of the younger folds. Biotite typically is smeared into minute flakes along the shear planes; mortar structure and at places mylonite are formed. The cataclastically deformed rocks are thoroughly cemented by quartz, which apparently was redistributed and recrystallized during cataclasis. Moench, Harrison, and Sims (1962, pls. 4, 6) illustrate typical cataclastic textures in microcline gneiss and in biotite-muscovite granite.

The cataclastic type of foliation is superposed on the older gneissic structure of the metamorphic rocks, and it largely follows the older foliation. In the incompetent biotite gneiss, for example, the anastomosing fractures produced by cataclasis are parallel to the original layering of the rock on both limbs of younger folds. These folds do not have a fracture cleavage or cataclastic foliation that cuts across the gneissic structure. Micas are bent and sillimanite needles are broken at the crests of crinkles and minor folds in B_y; but they are not cut by microfaults. In the relatively competent microcline gneiss, the cataclastic foliation generally is parallel to the older foliation, but locally, particularly in the Quartz Hill layer directly above the folds in the biotite gneiss, the new cataclastic foliation unquestionably cuts the older foliation. Instead of folding, the competent gneiss apparently yielded by shearing along fractures that dip uniformly northwest.

The northwest boundary of the zone of cataclasis is gradational and is placed only approximately (pl. 1). Southeast of the boundary, cataclasis is almost pervasive, whereas to the northwest cataclasis progressively diminishes and is confined to narrow zones. It has been recognized, however, as far as a mile north of the mouth of Fall River. Because the cataclasis is subparallel to the gneissic structure, the boundary is shown to dip rather gently northwest. Southeastward from the boundary the cataclastic foliation generally steepens, and northwestward it flattens, conforming with the gneissic structure. The southeast boundary of the zone probably extends northeastward from the vicinity of the mouth of Chicago Creek, but data are not adequate to show it on plate 1.

LINEATIONS IN B_y

Lineations in the B_y coordinate are dominantly small folds and crinkles, which are similar in form to the larger structures of the younger fold system. Mineral alinements and boudinage are common, but they are subordinate to small folds and crinkles.

A mineral alinement in the B_y direction is manifested by elongate crystals and aggregates of biotite, sillimanite, and muscovite. These minerals are typically coarser than they are in rocks that are not deformed by the younger deformation and are more numerous in the crests of minor folds.

Boudinage is present locally in the B_y direction; it reflects thinning of fold limbs and movement of material parallel to the limbs. Many of the boudins in B_y are broken along their axes and are healed by pegmatite.

Small lenticular bodies of pegmatite locally occupy the crestal areas of some younger folds. Some of the pegmatite has been sheared subsequent to its crystallization.

Lineations in B_y consistently bear N. 55° E. but plunge at various angles, depending mainly on the attitude of the original foliation surfaces. Diagrams 2 and 3 of figure 6, which represent areas within the zone of intense younger deformation, each have a lineation maximum that bears about N. 55° E. and plunges about 10° to 30° NE. Lineations in B_y are uncommon in the area of diagram 1, but those present may partly explain the east-northeast or west-southwest spread of the contours that are roughly concentric to the B_0 maximum.

LINEATIONS IN A_y

Slickenside striae and rodding are diagnostic lineations of the A_y coordinate, but folds are also present. Slickenside striae grade into a streaking marked by smears of finely comminuted biotite. The rodding is expressed by stretched mineral aggregates and grades into mineral alinement. Not uncommonly, small cross fractures segment the elongate minerals or mineral aggregates. Slickenside striae and rodding are common in the cataclastically deformed parts of the microcline gneiss north of Idaho Springs, and they are characteristic of the folded biotite gneiss, where they occur along foliation planes on both limbs of younger folds.

Folds in A_y are generally small and uncommon. Some are well exposed along Trail Creek (pl. 1) and in the Freeland-Lamartine district (Harrison and Wells, 1956, p. 63) on the northwest side of the intensely deformed zone. At the time of their mapping, Harrison and Wells (1956, p. 63) called these folds b_2, a minor fold direction younger than the major folds. The folds are low-amplitude 3- to 6-inch warps in biotite gneiss.

The interpretation that the lineations in A_y are genetically related to the younger deformation is based on several relationships observed in the field. Though widespread in occurrence, lineations in A_y occur most commonly in the zone of intense younger deformation, and the dominance of slickenside striae and rodding is consistent with the cataclastic character of the younger deformation. The fact that these lineations are abundant on both limbs of younger folds and bear consistently about 80° to the trend of the fold axes relates the lineations in A_y to slippage parallel to fold limbs during folding. The mechanism of slippage probably did not differ greatly from drag folding and probably was a differential movement of successively higher layers towards the anticlinal crests. The slickenside striae in the competent rock units, such as microcline gneiss and biotite-muscovite granite, are not directly related to folding, because these rock units did not yield by bending. The character and orientation of these lineations in A_y, however, are consistent with those in the rocks that were folded; further, there is abundant indication that folds in the incompetent units become shear zones in the competent units.

Lineations in A_y plunge N. 10°–40° W. on the northwest limbs of younger folds, or S. 10°–40° E. on the southeast limbs. In diagrams 2 and 3 of figure 6, they form strong and markedly elongate maxima that bear about N. 25° W. The contours do not extend much into the southeast quadrant of either diagram, because lineations in A_y were difficult to measure in the extremely crumpled southeast limbs of the younger folds. Slickenside striae that were measured on the southeast limbs, however, plunge consistently south-southeast.

GEOMETRY OF SUPERPOSITION

The younger folds are superposed on the Idaho Springs anticline, an asymetric major fold of the older deformation that is characterized by a steeply dipping northwest limb and a gently dipping southeast limb. The form and plunge of the younger folds depend on the geometry of superposition on this fold combined with the type of movement that produced them. The plunge of a younger fold is controlled by the line of intersection between the axial plane of the younger fold and the original foliation surface, and the shape of a fold is a function of the type of displacement shown by the fold. From the shapes of the younger folds and the orientation of lineations in A_o, it is inferred that the northwestern limb of each younger fold was raised relative to its southeastern limb.

The apparent distance parallel to the axial plane that the anticlinal bend of a fold has been raised relative to the synclinal bend is termed "lift" by Moench, Harrison, and Sims (1962) and provides a basis for comparing different cross sections of a fold. This concept is illustrated diagrammatically in figure 7.

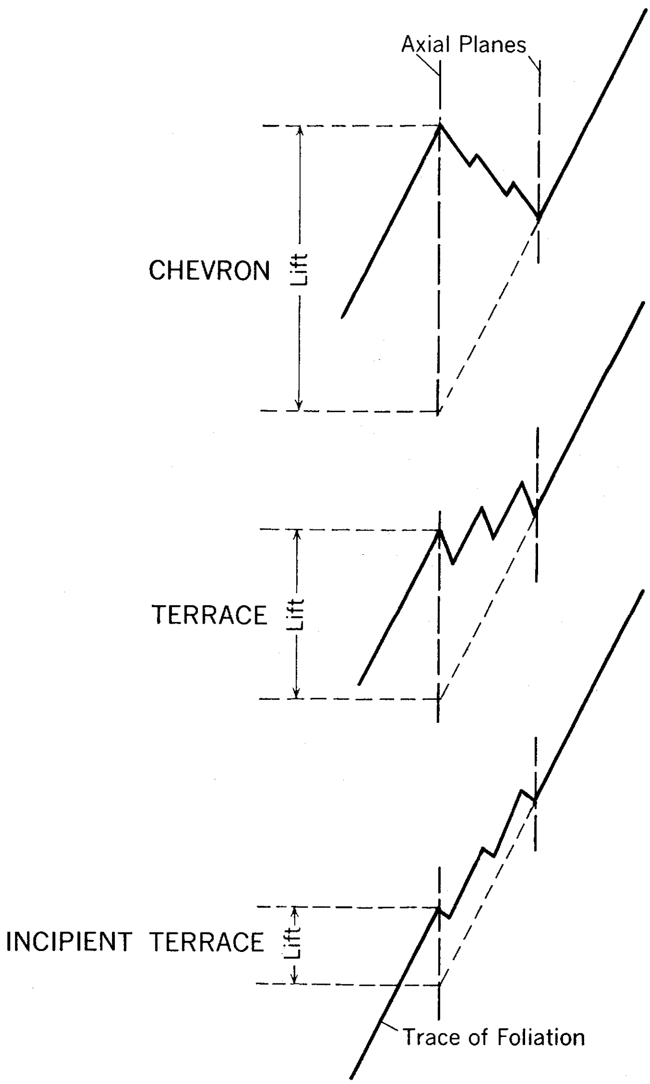

Figure 7.—Diagrams illustrating lift as related to younger folds. After Moench, Harrison, and Sims (1962, fig. 4).

The relation of younger folds to the Idaho Springs anticline, which is strongly asymmetric and plunges north-northeast, is illustrated diagrammatically on figure 8. All younger folds on the northwest limb of the Idaho Springs anticline are asymmetric and have long northwest limbs and short southeast limbs, as shown by the chevron in figure 8. As such folds are traced over the crest and down the southeast limb of the Idaho Springs anticline, they steepen in plunge to the northeast and change in shape to southeast-facing monoclines. This form results from the same type of movement—northwest side up relative to the southeast side—as that which formed younger folds on the northwest limb of the anticline; the northeastward plunge of the younger monoclines is controlled by the northeastward dip of the southeast limb of the older anticline. Similar relations of younger and older folds are shown on the detailed geologic map of the Chicago Creek area (Harrison and Wells, 1959, pl. 1).

Some folds are known to change along their traces from small warps in the foliation planes to crumpled structural terraces to chevron folds and ultimately to tightly closed folds, as illustrated diagrammatically in figure 9. These changes in form are due mainly to

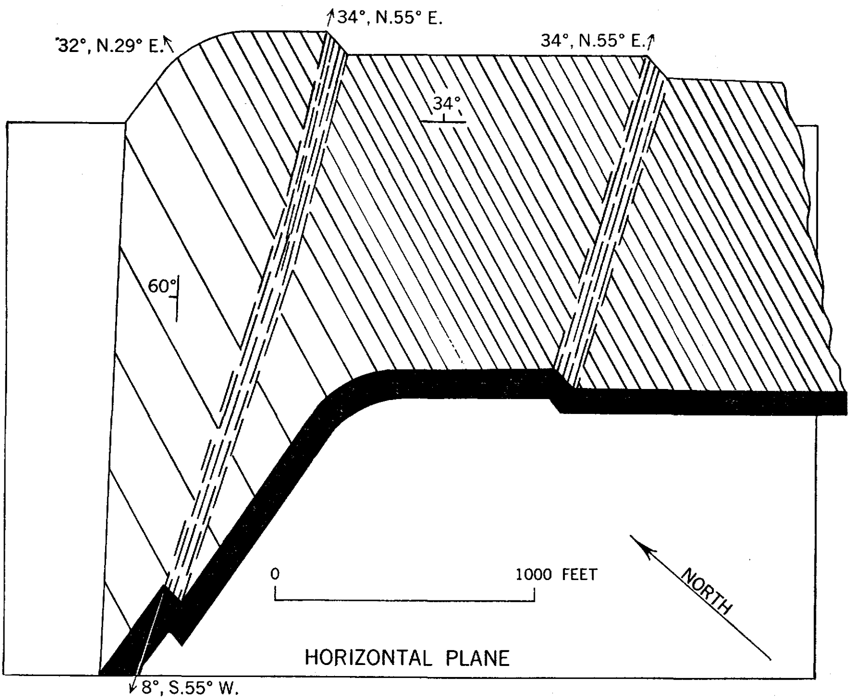

FIGURE 8.—Generalized isometric diagram showing superposition of younger folds on the Idaho Springs anticline. After Moench, Harrison, and Sims (1962, fig. 5).

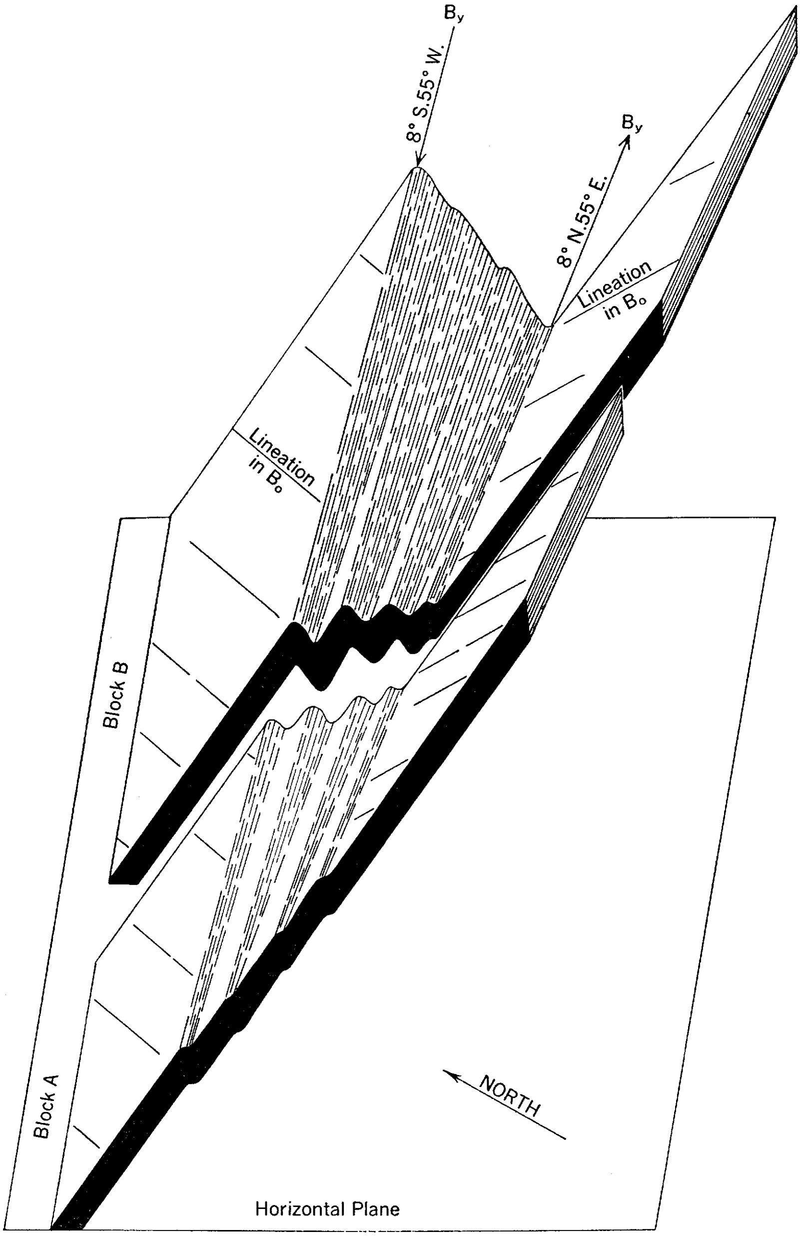

FIGURE 9.—Generalized isometric diagram showing progressive changes along a younger (B_y) fold as produced by increasing lift. After Moench, Harrison, and Sims (1962, fig. 6).

changes in the amount of lift along the trace of the folds and not to changes in the angular relations of the younger fold to the older foliation. Because the younger folds do not extend indefinitely along strike, variations in lift are unavoidable and produce different cross sections of a fold.

As expected, lineations related to the older fold system are deflected in plunge and bearing where intersected by a younger fold. In most exposures, however, the amount of deflection in bearing is small, amounting to 20° or less. Deflections of older lineations by a younger fold are shown diagrammatically on figure 10. The younger fold in B_y plunges moderately (about 35°) N. 55° E., and thus it conforms to a commonly observed field relation. Lineations in B_o (mineral alinement) are deflected by the fold. On the northwest limb of the fold (right side of diagram), the mineral alinement plunges steeply

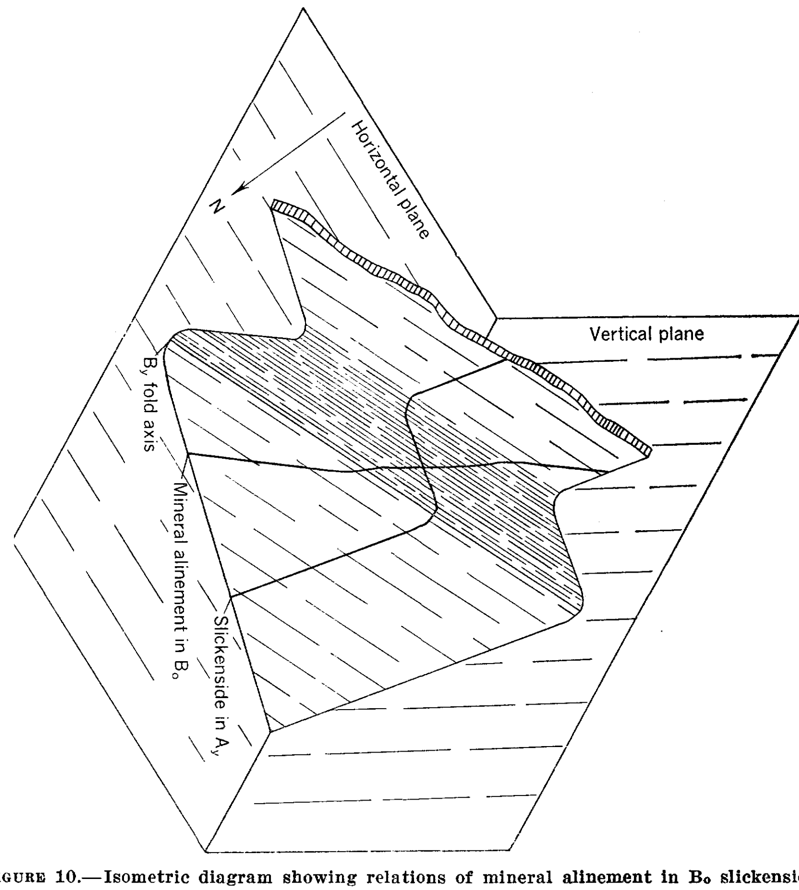

FIGURE 10.—Isometric diagram showing relations of mineral alinement in B_o slickensides in A_y to a fold in B_y. After Moench, Harrison, and Sims (1962, fig. 7).

735-934—64——5

about N. 25° E.; as this lineation is traced over the fold crest it deflects to a more northerly bearing, but as it is traced down the southeast limb it returns to a north-northeast bearing (or south-southwest if the plunge reverses). Similarly, the lineation in B_0 is deflected where it crosses the synclinal bend. In contrast a slickenside-striae lineation in A_y that bears N. 35° W. and S. 35° E., at right angles to the fold axis, is not deflected in bearing as it is traced across the folds.

The generalization that deflections of lineations in B_0 are small applies only to the most common angular relations between older and younger structures; locally the deflection is extreme. The amount of deflection depends upon the original angle between the older lineation and the younger fold axis and the amount of rotation on the younger fold axis; the wider the angle and the greater the rotation, or bending, the greater the deflection. This simple relation is complicated in the Idaho Springs district by the fact that the younger fold axes and axes of internal rotation are rarely parallel (Moench, Harrison, and Sims, 1962).

CHARACTER, RELATIVE AGE, AND ORIGIN OF DEFORMATION

The cataclasis and minor folding that characterize the younger deformation are interpreted to have resulted from regional shearing, which possibly was a deep manifestation of faulting. These features are expressions of movements along the narrow but extensive Idaho Springs-Ralston shear zone (Tweto and Sims, 1960; 1963, p. 998)—a zone of possibly major structural discontinuity and possibly of great antiquity. The recognized effects of the younger deformation, in fact, may express only late movements in the shear zone.

The folds and cataclasis of younger deformation post-date the biotite-muscovite granite, which evidently was emplaced at the end of the older deformation, because this rock is locally intensely cataclastically deformed. The younger deformation is probably Precambrian in age, however, because it is older than faults that are thought to have originated in the Precambrian (Tweto and Sims, 1963, p. 1001) and because the Idaho Springs-Ralston shear zone does not extend into the overlying sedimentary rocks of Pennsylvanian age on the east flank of the Front Range. Data obtained by Tweto and Pearson (1958, p. 1748) on a similar shear zone in the Sawatch Range to the southwest tend to corroborate a Precambrian age, for most of the shearing on this zone "preceded the emplacement of Precambrian dike rocks of several types." Tweto and Pearson point out, however, that some movement resulting in gouge-filled fractures took place later than the Cambrian.

Movement patterns in the Idaho Springs-Ralston shear zone are poorly understood and further studies are needed. Most likely, the movements differed from time to time, and the structural data obtained in a small area may indicate only the latest movements in that area. To judge from the consistent asymmetry of the younger folds, the northwest side of the zone must have moved upward relative to the southeast side, and the cumulative displacement across the zone was probably several thousands of feet. The inference that the movements were dominantly vertical and not parallel to the strike of the zone is fortified by the consistent north-northwest or south-southeast bearing of the A_y slickenside striae and rodding. Evidence for some shortening of the crustal segment across the zone is given, on the other hand, by the tightly closed and locally overturned folds and by the observation that shear planes in the competent Quartz Hill layer of microcline gneiss dip northwest. Movements along these shear surfaces must have been up-dip because folds in the underlying incompetent biotite gneiss formed by coupled movements—northwest side up relative to the southeast side. Also, subhorizontal zones of shearing extend northwestward beneath the Bellevue Mountain area parallel to the gneissic structure of the rock—a fact that suggests that subhorizontal thrust movements took place locally.

These inferred movement patterns are shown diagrammatically in figure 11. Block *A* is a schematic replica of a typical younger fold in the contact zone between incompetent biotite gneiss and the competent Quartz Hill layer of microcline gneiss. Folds having axial planes that dip steeply and uniformly southeast have been traced long distances through the incompetent unit (pl. 1). Where the folds

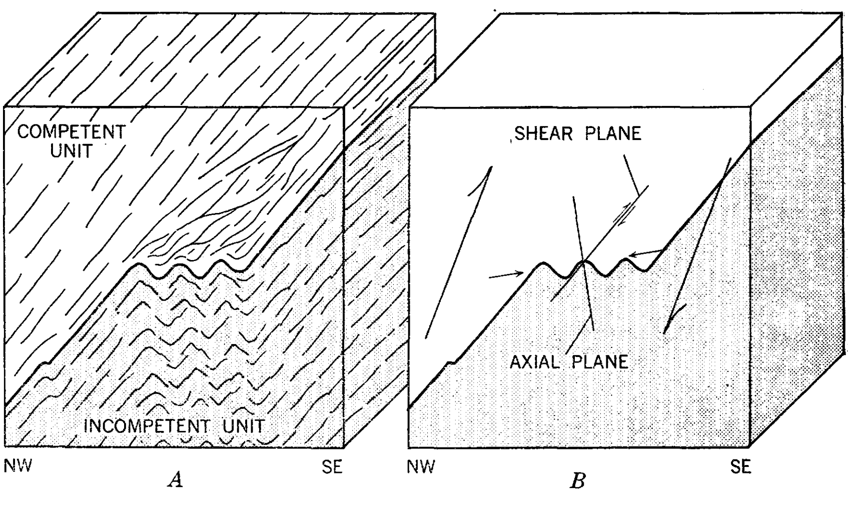

FIGURE 11.—Schematic block diagrams showing relation between folds in incompetent unit and shear planes in competent unit (*A*), and inferred rotational stress pattern (*B*).

enter the competent unit they deform the contact, but the competent unit is sheared along northwest-dipping fractures and is not noticeably folded. The competent rock directly above the fold is extremely sheared, and arcuate fractures extend around elliptical deformed relicts of the rock. Features similar to those shown in block *A* have been seen in small folds about a foot in width. As stated, the folds and deflection of the contact must have formed by upward movements of the northwest side relative to the southeast side and by some shortening of the crustal segment. If, as shown in Block *B*, the shear planes had the movement indicated and the axial planes formed normal to the component of shortening, then the rotational movement pattern shown may be inferred.

McBirney and Best (1961, p. 497) concluded that the linear features produced in their experiments "lie within the plane normal to the maximum shortening, but their plunge is independent of the position of intermediate and minimum strain axes." Unlike their experimentally produced superposed folds, the younger folds of the Idaho Springs district probably formed under coupled (and not opposed) stress, and the axial planes, which contain the B_y linear elements, may or may not be normal to the direction of maximum shortening. In figure 11, however, the axial plane is shown to be normal to the component of compression in the couple.

The younger monoclines on the southeast limb of the Idaho Springs anticline may have formed by a similar pattern of movement. However, one would expect to find crosscutting shear planes that dip northwest in the competent layers, and such shear planes have not been recognized on the monoclines.

The younger deformation probably took place in an environment of high pressure-temperature, although not so high as that of the older deformation. The absence of gouge and breccia, characteristic of near-surface faulting, and the fact that shearing was pervasive through a wide zone are suggestive of high confining pressures. On the other hand, the dominantly cataclastic character of the younger deformation suggests that it took place at somewhat shallower depths than the older deformation. The prevailing temperatures of the younger deformation were fairly high, because biotite, although smeared along shear surfaces, is rarely chloritized, and both biotite and sillimanite have recrystallized locally to form coarse mineral alinements in B_y. In addition, some pegmatite was formed or redistributed along the crests of younger folds. On the other hand, the local chloritization of biotite, and possibly the widespread distribution of muscovite in rocks that contain sillimanite and microcline, and the local recrystallization of amphibolite to form calcium-silicate gneiss are suggestive of some metamorphic retrogression that may have taken place during the younger deformation. It is possible that the observed effects of the younger deformation were produced over a long period of time at progressively shallower depths—an hypothesis that might explain the apparently conflicting evidence for recrystallization at high temperatures and local metamorphic retrogression.

SUMMARY

The bedrock of the Idaho Springs district is composed largely of a conformable succession of gneissic rocks that have been deformed at least twice in Precambrian time and invaded by three varieties of granitic rocks. Most of the gneissic rocks are probably metamorphosed sedimentary rocks. Nothing can be said of their age or of the sedimentary conditions that produced them because these rocks have been completely metamorphosed, intensely deformed, and their composition partly changed.

Five major lithologic units are recognized that may represent an original sedimentary stratigraphic succession. Lowest in this succession is a unit of biotite gneiss that contains several layers of quartz gneiss, a garnetiferous variety of sillimanitic biotite-quartz gneiss, and layers of granite gneiss. This unit is overlain successively by the Big Five layer of microcline gneiss, a thick unit of biotite gneiss, the Quartz Hill layer of microcline gneiss, and, at the top, a thick unit of biotite gneiss. The three central units in this succession pinch out near the southwest side of the district. These pinchouts and the convergence of the Big Five and Quartz Hill layers may be related to a large overturned anticline that plunges gently southwest or possibly to shearing along the Idaho Springs-Ralston shear zone. For lack of conclusive evidence, the simplest interpretation—that the pinchouts are stratigraphic, and that the succession is upright and not a folded repitition—is favored in this report.

With the exception of the granite gneiss, the gneissic rocks are thought to represent a succession of clastic sedimentary rocks. The biotite gneiss may represent interbedded sandstone and shale, whereas the microcline gneiss may represent more massive, and possibly cleaner, sandstone. Most likely the quartz gneiss is metamorphosed quartzose sandstone. The amphibolite and associated calcium-silicate gneiss may represent metamorphosed impure dolomitic sediments or possibly basalt; the calcium-silicate gneiss not associated with amphibolite may represent sediments that were still more calcareous or dolomitic.

The granite gneiss is intimately mixed with the biotite gneiss and may have formed by replacement of this rock. The possibilities are not excluded, however, that the granite gneiss was introduced, or that the southward increase in abundance of the rock reflects a sedimentary facies change.

Two main tectonic events mark the Precambrian structural history in the Idaho Springs district. The older deformation produced the main structural framework and was a plutonic, gneiss-forming chapter of history. It evidently coincided with the peak intensity of metamorphism, which was equivalent to the upper range of the almandine amphibolite facies as defined by Fyfe, Turner, and Verhoogen (1958, p. 231) and by Turner and Verhoogen (1960, p. 549). The younger deformation was superposed on the first, and was more limited in distribution and more cataclastic in character. It took place under slightly lower pressure-temperature conditions, and locally it may have effected slight metamorphic retrogression.

The older deformation produced a major fold system having axes that trend sinuously north-northeast. In the Idaho Springs district, this fold system is dominated by the Idaho Springs anticline. The deformation was sufficiently plastic to produce markedly disharmonic folds and a great abundance and variety of minor structures that formed by plastic flowage, but it was not so plastic as to produce large flowage folds of the type described by Bain (1931). It was accompanied by complete recrystallization that produced the gneissic structure of the rocks. Granodiorite bodies were emplaced contemporaneously with this deformation, but sufficiently early to be folded and to assume an internal metamorphic structure that is continuous with the structure of the wallrocks. Small bodies of quartz diorite were emplaced soon after the intrusion of the granodiorite, and biotite-muscovite granite was probably emplaced during the waning stages of the deformation or possibly later.

The younger deformation resulted from movements along the northeast-trending Idaho Springs-Ralston shear zone (Tweto and Sims, 1960)—a zone of possibly major structural discontinuity in the Front Range. Within the district, movements along the zone produced intense cataclasis and many small folds. Though individually small, the folds are consistently asymmetric and suggest a cumulative known displacement across the zone—northwest side up relative to the southeast side—of several thousands of feet. The fold axes bear N. 55° E. with remarkable persistency but, because the folds are superposed on the older structural framework, they plunge at angles that are controlled by the intersections of their axial planes with the preexisting attitude of the foliation. Although the younger deformation produced some features of plastic folding and was accompanied by some recrystallization, it was characteristically cataclastic. Some large masses of competent rocks, as microcline gneiss, were intensely sheared but little folded, whereas the less competent biotite gneiss was both sheared and folded. In the folded rocks, cataclastic products formed by slippage on preexisting foliation planes for much the same reason that drag folds form on the limb of a larger fold; but in the competent rocks the cataclastic products formed on closely spaced, subparallel, anastomosing shear surfaces that mostly dip northwest and in places transect the original foliation. Because the younger folds are superposed on the older ones, the shapes assumed by the younger folds are controlled by the geometric relation between them and the preexisting attitude of the foliation and by the coupled movements that produced the folds. From all indications, these movements were uniform in sense—northwest side up, and no recognized strike-slip component. These factors combined to produce folds in B_y that range from complex structural terraces to tightly closed chevron folds on the northwest limb and monoclines on the southeast limb of the older (B_o) Idaho Springs anticline.

During the Laramide revolution, a succession of porphyritic intrusive rocks was emplaced, and associated metalliferous veins of the Front Range mineral belt were formed. Very likely these features represent a reactivation of the Idaho Springs-Ralston shear zone (Tweto and Sims, 1960; Sims, Moench, and Harrison, 1959), for the mineral belt and this zone are subparallel and their southeast boundaries nearly coincide.

REFERENCES CITED

Bain, G. W., 1931, Flowage folding: Am. Jour. Sci., 5th ser., v. 22, p. 503–530.

Ball, S. H., 1906, Pre-Cambrian rocks of the Georgetown quadrangle, Colorado: Am. Jour. Sci., 4th ser., v. 21, p. 371–389.

Bastin, E. S., and Hill, J. M., 1917, Economic geology of Gilpin County and adjacent parts of Clear Creek and Boulder Counties, Colorado: U.S. Geol. Survey Prof. Paper 94, 379 p.

Billings, M. P., 1954, Structural geology: 2d ed., New York, Prentice-Hall, Inc., 514 p.

Cloos, Ernst, 1946, Lineation, a critical review and annotated bibliography: Geol. Soc. America Mem. 18, 122 p.

Faul, Henry, ed., and others, 1954, Nuclear geology—a symposium on nuclear phenomena in the earth sciences: New York, John Wiley & Sons, Inc., 414 p.

Fyfe, W. S., Turner, F. J., and Verhoogen, John, 1958, Metamorphic reactions and metamorphic facies: Geol. Soc. America Mem. 73, 259 p.

Harrison, J. E., and Moench, R. H., 1961, Joints in Precambrian rocks, Central City-Idaho Springs area, Colorado: U.S. Geol. Survey Prof. Paper 374–B, p. B1–B14.

Harrison, J. E., and Wells, J. D., 1956, Geology and ore deposits of the Freeland-Lamartine district, Clear Creek County, Colorado: U.S. Geol. Survey Bull. 1032–B, p. 33–127.

——— 1959, Geology and ore deposits of the Chicago Creek area, Clear Creek County, Colorado: U.S. Geol. Survey Prof. Paper 319, 92 p.

Knopf, E. B., and Ingerson, F. E., 1938, Structural petrology: Geol. Soc. America Mem. 6, 270 p.

Lindgren, Waldemar, 1933, Mineral deposits: 4th ed., New York and London, McGraw-Hill Book Co., Inc., 930 p.

Lovering, T. S., and Goddard, E. N., 1950, Geology and ore deposits of the Front Range, Colorado: U.S. Geol. Survey Prof. Paper 223, 319 p. [1951].

McBirney, A. R., and Best, M. G., 1961, Experimental deformation of viscous layers in oblique stress fields: Geol. Soc. America Bull., v. 72, no. 3, p. 495–498.

Moench, R. H., Harrison, J. E., and Sims, P. K., 1962, Precambrian folding in the Idaho Springs-Central City area, Front Range, Colorado: Geol. Soc. America Bull., v. 73, no. 1, p. 35–58.

Sims, P. K., and others, 1963, Geology of uranium and associated ore deposits, central part of the Front Range Mineral Belt, Colorado: U.S. Geol. Survey Prof. Paper 371, 119 p.

Sims, P. K., Drake, A. A., Jr., and Tooker, E. W., 1963, Economic geology of the Central City district: U.S. Geol. Survey Prof. Paper 359, 231 p.

Sims, P. K., Moench, R. H., and Harrison, J. E., 1959, Relation of Front Range mineral belt to ancient Precambrian structures [abs.]: Geol. Soc. America Bull., v. 70, no. 12, pt. 2, p. 1749.

Skinner, B. J., 1956, Physical properties of end-members of the garnet group: Am. Mineralogist, v. 41, p. 428–436

Spurr, J. E., Garrey, G. H., and Ball, S. H., 1908, Economic geology of the Georgetown quadrangle, Colorado: U.S. Geol. Survey Prof. Paper 63, 422 p.

Tooker, E. W., 1956, Altered wall rocks along vein deposits in the Central City-Idaho Springs region, Colorado, *in* Swineford, Ada, ed., Clays and clay minerals: Natl. Research Council Pub. 456, p. 348–361.

Turner, F. J., and Verhoogen, John, 1960, Igneous and metamorphic petrology: 2d ed., New York, McGraw-Hill Book Co., Inc., 694 p.

Tweto, Ogden, and Pearson, R. C., 1958, Great Precambrian shear zone, Sawatch Range, Colorado [abs.]: Geol. Soc. America Bull., v. 69, no. 12, pt. 2, p. 1748.

Tweto, Ogden, and Sims, P. K., 1960, Relation of the Colorado mineral belt to Precambrian structure, *in* Short papers in the geological sciences: U.S. Geol. Survey Prof. Paper 400–B, p. B8–B10.

——— 1963, Precambrian ancestry of the Colorado Mineral Belt: Geol. Soc. America Bull., v. 74, no. 8, p. 991–1014.

Weiss, L. E., 1959, Geometry of superposed folding: Geol. Soc. America Bull., v. 70, no. 1, p. 91–106.

Wells, J. D., 1960, Petrography of radioactive Tertiary igneous rocks, Front Range mineral belt, Colorado: U.S. Geol. Survey Bull. 1032–E, p. 223–272 [1961].

Wells, J. D., Sheridan, D. M., and Albee, A. L., 1961, Metamorphism and structural history of the Coal Creek area, Front Range, Colorado, *in* Short papers in the geologic and hydrologic sciences: U.S. Geol. Survey Prof. Paper 424–C, p. C127–C131.

INDEX

[Italic numbers indicate major references]

The U.S. Geological Survey Library has cataloged this publication as follows:

Moench, Robert Hadley, 1926–
Geology of Precambrian rocks, Idaho Springs district, Colorado. Washington, U.S. Govt. Print. Off., 1964.

iv, 70 p. maps (1 col.) diagrs. (1 col.) tables. 24 cm. (U.S. Geological Survey. Bulletin 1182–A)
Contributions to economic geology.
Part of illustrative matter fold. in pocket.
Prepared on behalf of the U.S. Atomic Energy Commission.
Bibliography: p. 67–68.
(Continued on next card)

Moench, Robert Hadley, 1926– Geology of Precambrian rocks, Idaho Springs district, Colorado. 1964. (Card 2)

1. Geology—Colorado—Idaho Springs region. 2. Geology, Stratigraphic—Pre-Cambrian. I. Title. (Series)