Altered Wallrocks in the Central Part of the Front Range Mineral Belt Gilpin and Clear Creek Counties, Colorado

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GEOLOGICAL SURVEY PROFESSIONAL PAPER 439

Prepared on behalf of the United States
Atomic Energy Commission

U.S. DEPARTMENT OF THE INTERIOR
March 3, 1849

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ALTERED WALLROCKS IN THE CENTRAL PART OF THE FRONT RANGE MINERAL BELT, GILPIN AND CLEAR CREEK COUNTIES, COLORADO

By E. W. Tooker

ABSTRACT

The wallrocks of fissure veins in the mining districts of the central part of the Front Range mineral belt are altered to successive zones of sericitized and argillized rock. In general, the altered wallrocks are similar throughout the region in that sericite is adjacent to the veins and clay minerals are farther out, but in detail the degree of alteration, the width of the altered zones, and to some extent the type of alteration varies because of differences in lithology and structure of the host rocks and differences in the intensity of the hydrothermal environment.

The wallrocks are predominantly of Precambrian age, but some are of Tertiary age. The Precambrian wallrocks include a variety of metamorphic and igneous rocks that can be grouped into two general types: (1) those containing moderately abundant K-feldspar, and (2) those deficient in K-feldspar but containing abundant iron-bearing minerals. The first group includes biotite-muscovite granite, granodiorite, microcline-quartz-plagioclase-biotite gneiss, and biotite-quartz-plagioclase and related gneisses; the second group includes quartz diorite, amphibolite, garnet-quartz gneiss, and biotite schist. The Tertiary igneous rocks include abundant, small, irregular intrusive dikes and plutons of alkalic and calc-alkalic rocks such as porphyritic bostonite and quartz monzonite.

The ore deposits consist of a simple suite of base-metal sulfide minerals in a dominantly quartz gangue and are valuable chiefly for their gold and silver content. The ore minerals were deposited in three closely related stages, which were, from oldest to youngest, a uranium stage, a pyrite stage, and a base-metal stage. Alteration largely preceded ore deposition. Precipitation of the ore minerals of the pyrite and base-metal stages yielded a concentric zonal pattern that is particularly well defined in the Central City district.

The width of the altered rock zones that bound the veins ranges from a few inches up to a few tens of feet. In general, veins in the central zone, as at Central City, are bounded by wider altered rock zones than veins in the intermediate or peripheral zones; however, little or no relation between width of the ore zone and width of altered rock is apparent in other areas, such as at the Freeland-Lamartine district. Supergene alteration effects are superposed on the hydrothermal altered rock and can be distinguished by the presence in the veins of secondary minerals such as limonite and chalcocite. Clay minerals formed by supergene processes have not been distinguished in this study from the hypogene clay minerals.

Most of the fresh and altered rocks are characterized by similar, repetitive mineralogical-textural zones: (1) fresh rock, (2) weakly argillized rock, (3) strongly argillized rock, and (4) sericitized rock. Fresh rock of zone 1 grades in to rock of zone 2 in which plagioclase and hornblende are altered incipiently to clay minerals. A gradual softening of the rock occurs as clay minerals completely replace plagioclase, hornblende, and some of the biotite in intensely argillized rock (zone 3). Further alteration results in the remainder of the biotite and the clay minerals being converted completely into sericite (zone 4), and in the recrystallization of quartz into aggregates of small strain-free crystals; K-feldspar is not visibly altered in zone 4. The boundary between the soft argillized rock of zone 3 and the hard, bleached, sericitized, and at places pyritized rock of zone 4 commonly is sharp. Along some veins, the alteration assemblages differ from those in the normal altered rock zones in that the sericite zone (4) extends into the argillic zone (2) and the clay zone (3) is virtually missing. Another exception to the normal zone pattern, found where the altered rock next to the vein was originally composed almost wholly of minerals unstable in the hydrothermal environment, is a complete absence of sericitized rock; instead the whole width of the zone is argillized. The clay mineral components in zones 2 and 3 consist of mixed assemblages of montmorillonite, illite, random mixed-layer montmorillonite-illite, chlorite, kaolinite, and halloysite. The iron needed for the pyrite in zone 4 was derived in part from altered magnetite (and hematite) and mafic minerals such as biotite. Carbonate minerals and quartz occur at many places in zone 4 as well as in the veins.

The distribution of elements in altered wallrocks is considered to be largely the result of movements of ions in and out of the interstices of a silicon-oxygen or aluminum-oxygen framework in which the oxygen-ion positions are the most abundant and largest. Calculations from chemical analyses of samples of representative rocks show that, in the alteration of equal volumes of rock, small amounts of K^+, Fe^{+2}, C^{+4}, H^+, S^{-2}, and at places Al^{+3} were added, and Si^{+4}, Na^+, Ca^{+2}, Fe^{+3}, and Mg^{+2} were removed. In the chemical sense, this is a hydrolysis process in which there is a replacement of chemical equivalents of metal cations by H^+ ions which enter the structure and produce polarized hydroxyl groups. In the argillized zone hydration water is added also but without attendant replacements. The distribution of trace elements concealed in rock-forming minerals shows that most tend to be removed or at least displaced veinward. Relatively large interstitial ions with low ionization potential generally are removed, while smaller ions with larger ionization potential and higher valences are concentrated in rocks near the veins.

The powder pH of fresh and altered rocks mostly gives a measure of the relative hydrolysis of the mineral components. In general, clay minerals and sericite do not hydrolyze significantly, giving lower values than feldspar or mica. In the peripheral zone, fresh rock of granitic composition has an average powder pH of 8, whereas altered rock of the same composition has an average pH of 6. In the cental zone, in areas not greatly affected by supergene oxidation, the dominant rock—microcline-quartz-plagioclase-biotite gneiss—has an average powder pH of about 6 where fresh and about 4.5 where altered adjacent to veins. These differences between the peripheral and central ore-zone rocks are considered to indicate greater supergene action in the central zone.

The hydrothermal solutions that altered the rocks, as inferred from observable surface and underground waters and the mineralogical and chemical changes in the host wallrocks, were relatively dilute, slightly acid, and initially carried K and subsequently CO_2 and S ions or complexes. As the rocks were altered, these solutions were modified by chemicals replaced in the wallrocks and by vein-forming ore metals. Possibly the hot springs in the area, such as those at Idaho Springs, represent spent hydrothermal solutions.

Altered wallrocks are not infallible guides to ore, for alteration was largely accomplished before the deposition of the valuable vein minerals, and further, there is no close correlation between width of the altered rock envelope and width of the vein filling. Aside from uranium, most of the ore constituents were deposited in vein openings following the wallrock alteration. Primary uranium minerals are believed to have been deposited earlier in veins that were open during late stages of the rock alteration.

INTRODUCTION

The altered wallrocks bordering the hydrothermal vein deposits in the mining districts of the central part of the Front Range mineral belt, Colorado, were studied as a part of a detailed examination of this region by the U.S. Geological Survey on behalf of the U.S. Atomic Energy Commission. A primary aim of the study was to determine the usefulness of altered wallrocks as a prospecting guide in the search for uranium ore. In addition, the study complements concurrent geologic studies of the ore deposits in that determination of the character and extent of the altered metamorphic and igneous wallrocks has yielded data bearing on the origin of the ore deposits themselves.

LOCATION

The area investigated during this study is in Gilpin and Clear Creek Counties, Colo., about 35 miles west of Denver (fig. 1); it lies between long 105°29′ and 105°38′ W. and lat 39°50′ and 39°43′ N. and includes the mining districts of Central City, Idaho Springs, Freeland-Lamartine and Chicago Creek, and Lawson-Dumont-Fall River (pl. 1).

The region is in the center of the Front Range segment of the Colorado mineral belt, which extends northeastward across the north-trending Front Range. It is east of the Continental Divide and is dissected by the drainage system of Clear Creek; altitudes range from 7,500 feet at Idaho Springs and 8,500 feet at Central City to 10,700 feet in the Freeland-Lamartine district.

PREVIOUS INVESTIGATIONS OF ALTERED WALLROCKS

Previously published reports, on parts of the Idaho Springs, Freeland-Lamartine, Lawson, Dumont, and Fall River[1] districts by Spurr, Garrey, and Ball (1908) and on the Central City and parts of the Idaho Springs districts by Bastin and Hill (1917), present some generalized petrographic descriptions of the altered wallrocks. Lovering and Goddard (1950) summarize observations on altered wallrocks associated with ore deposits in this and other parts of the Front Range mineral belt, noting (p. 77) that clay minerals become more and more prominent members of the alteration suite northeastward along the mineral belt.

Several reports published during the past few years describe related altered wallrocks in other Front Range mining districts. Lovering (1941, 1950) and Lovering and Tweto (1953) described the hydrothermally altered wallrocks associated with tungsten veins at Nederland; Gonzalez-Bonorino (1956, 1959) investigated these same altered rocks. Wright (1954) described the altered rocks associated with silver-bearing base metal-uranium veins in the Caribou district. The term "argillic alteration" was first used by Lovering (1941, p. 236) for the prominently developed clay-mineral assemblage found in altered Boulder Creek granite bordering tungsten veins near Nederland. According to Lovering and Tweto (1953, p. 58), a broad zone in which sericite, allophane, beidellite, hydrous mica, dickite, and halloysite occur successively veinward is followed by a narrow zone of sericitized rock adjacent to the vein.

Gonzalez-Bonorino (1959) recognized six distinctive alteration patterns, based on chemical, X-ray, and petrographic data, in the altered rocks in different parts of the mineral belt. The most complex of these found in the Nederland tungsten—Gold Hill pyritic gold districts consists of four main zones, characterized by hydromica (innermost zone), orthoclase, kaolinite, and montmorillonite. Two zones were found in the Caribou lead-silver-zinc mine characterized respectively by kaolinite (inner) and montmorillonite.

In an earlier study of the hydrothermal alteration of monzonite at the Caribou mine, Wright (1954) described complex relationships in more detail than did Gonzalez-Bonorino. Petrographic, X-ray diffraction, chemical, and differential thermal analysis data reveal (p. 148) an outer propylitic, a succeeding argillic, and an inner sericitic zone:

> In Stage 1 (comprising "comparatively unaltered" rock) the ferromagnesian minerals, biotite and hornblende, were in part altered to chlorite, magnetite, and pyrite, while the felsic minerals remained essentially fresh. In Stage 2, chlorite replacement of biotite was important, but the chlorite disappeared in the more intensive Stages 3 and 4 (as it did also in Sales and Meyer's * * * argillized and sericitized zones). Minor sericite replace-

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[1] The Lawson, Dumont, and Fall River districts of Spurr, Garrey, and Ball (1908) have been combined into the Lawson-Dumont-Fall River district in this report.

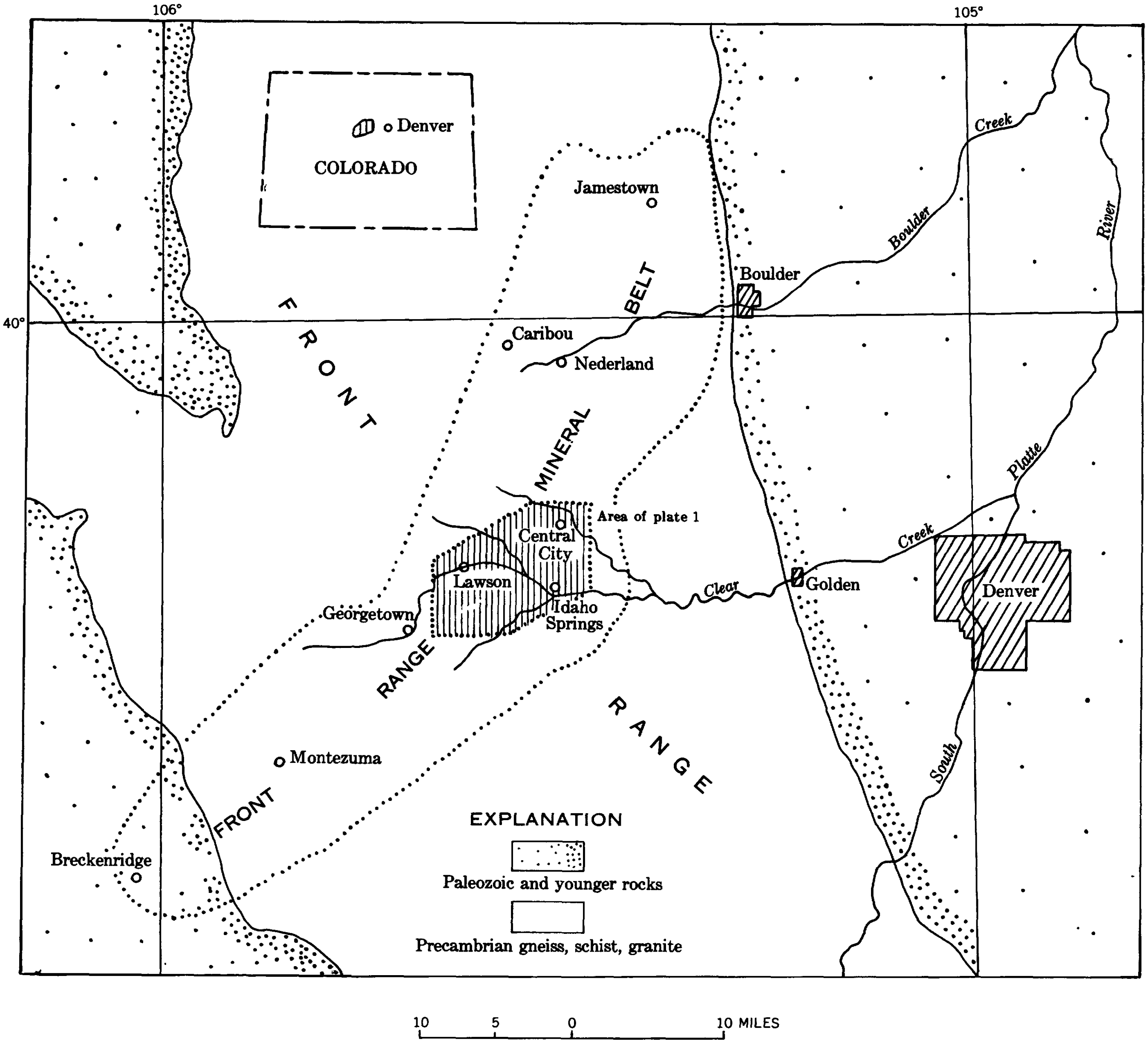


FIGURE 1.—Index map of the central part of the Front Range mineral belt area, Colorado, showing location of area investigated on plate 1

ment of biotite, and kaolinite development in feldspars, occurred in the more intensively altered portion of Stage 2. Stage 3 is characterized by the development of clays, especially hydromica (replacing biotite), and kaolinite and montmorillonite (chiefly in feldspars), together with some sericite. The ratio kaolinite: montmorillonite increases with the intensity of alteration approaching Stage 4. Some kaolinite remains in Stage 4 (the most intense), but is subordinate to sericite and fine-grained quartz, the most abundant alteration products.

Beyond the confines of the Front Range, classic investigations by Lovering and others (1949) and by Sales and Meyer (1948, 1949) have greatly influenced subsequent studies of altered wallrocks. A summary of field and laboratory data on hydrothermal alteration by Kerr (1955) and the discussion of altered rocks as guides to ore by Schwartz (1955) provide extensive review bibliographies on these subjects.

BACKGROUND GEOLOGIC STUDIES

Several reports that describe the geology and ore deposits of the mining districts in the central part of the Front Range mineral belt have been published as a result of the current investigations by the Geological Survey, and others, now being prepared, will be published later. One of these (Sims and others, 1963) describes the uranium deposits and summarizes the data on the ore deposits from all of the districts included in the Survey's current investigations in this region. Detailed reports that describe the geology of the Freeland–Lamartine district (Harrison and Wells, 1956), the Chicago Creek area (Harrison and Wells, 1959), and the Central City district (Sims, Osterwald, and Tooker, 1955; Sims, Drake, and Tooker, 1963) have been published; the reader should refer to these for a full description of the rock units and ore deposits of these districts. For a description of the Tertiary intrusive rocks of the region, the reader is referred to a paper by Wells (1960). Several papers (Harrison and Leonard, 1952; Moore and Butler, 1952; Sims and Tooker, 1955; Wells, 1955; Hawley and Moore, 1955; Sims, 1956; Sims and Tooker, 1956; Tooker, 1956; and Drake, 1957) describe some details of features of the uranium deposits.

ACKNOWLEDGMENTS

The present study was undertaken by the U.S. Geological Survey on behalf of the Division of Raw Materials of the U.S. Atomic Energy Commission. Fieldwork and sampling were carried out from 1953 to 1955; laboratory investigations were done intermittently from 1954 to 1957. I am indebted to P. K. Sims, who directed the geologic investigation in the region, and to A. A. Drake, Jr., J. E. Harrison, C. C. Hawley, R. H. Moench, F. B. Moore, and J. D. Wells for assistance in sampling and for advice on problems of the geology of the region and of individual mines. These and other associates in the Geological Survey have been helpful in discussions during all phases of the study. The writer also wishes to acknowledge the assistance of Peter Buseck, A. R. Dearth, and J. R. McDonald of the Geological Survey in the field and laboratory. Many local mining people provided information of great value to this study, for which I am deeply grateful.

METHODS OF INVESTIGATION

The methods of field collection and laboratory examination of samples are briefly summarized here to provide a basis for interpretation or future reinterpretation of these data. Over periods of time, methods of study change and are improved, thus it is desirable to establish for the record the means by which these data were derived.

FIELD STUDIES

The wallrocks of veins throughout the region were systematically examined and sampled in conjunction with mapping of the mines. The zones of altered rock bordering the veins are generally narrow, and it was not practical to map them. Instead, samples of different kinds of altered rocks that could be distinguished megascopically were collected and sketches were made to show the locations of the samples with respect to the veins and other structural features.

TABLE 1.—*Samples examined by chemical analysis and semiquantitative spectrographic analysis*

[Chemical analyses are in table 18; semiquantitative analyses are in figures 36 to 62. V, wallrock caught up in vein; G, rock in the gouge zone]

Sample	Altered-rock zone	Rock type	Location
1[1]	1	Biotite-muscovite granite	Nabob and Jo Reynolds mine.
2[1]	2–3	do	Do.
3[1]	3	do	Do.
4[2]	1	Granodiorite	M. and M.-Dixie mine.
5[1]	3	do	Do.
6[1]	3	do	Do.
7[1]	3–4	do	Do.
8	4	do	Do.
9	2	Microcline-quartz-plagioclase-biotite gneiss.	E. Calhoun mine.
10	3	do	Do.
11	3–4	do	Do.
12	3–4	do	Do.
13a	4	do	Do.
13	V	do	Do.
14[1]	1	Granodiorite	Surface, Central City district.
15[1]	1	do	Do.
16	1–2	Quartz diorite	Nabob and Jo Reynolds mine.
17	3	do	Do.
18[1]	4	do	Do.
19	1	Biotite-quartz-plagioclase gneiss	Essex mine.
20	2	do	Do.
21	3	do	Do.
22	4	do	Do.
23	V	do	Do.
23a	V	do	Do.
24	1–2	do	R. H. D. McKay shaft.
25	3	do	Do.
26	3	do	Do.
27	1–2	do	Almadin mine.
28	3–4	do	Do.
29	1–2	Biotite-amphibole gneiss	Do.
30	3–4	do	Do.
31	1	Amphibolite	Kitty Clyde mine (31) and R. H. D.-McKay mine.
32	2	do	Do.
33	3	do	Do.
33a	3	do	Do.
34	3–4	do	Do.
35	1–2	Garnet-quartz gneiss	Golconda mine.
36	1–2	do	Do.
37	2–3	do	Marys mine.
38	2–3	do	Do.
39	3–4	do	Almadin mine.
40	3–4	do	Do.
41	1	Granodiorite	Hayes and Wheeler Tunnel.
42	2	do	Do.
43	3	do	Do.
44	3–4	do	Do.
45	4	do	Do.
46	1	Microcline-quartz-plagioclase-biotite gneiss.	Essex mine.
47	2	do	Do.
48	3	do	Do.
49	4	do	Do.
50	G	do	Do.
51	V	do	Do.
52	V	do	Do.
53	V	do	Do.
54	G	do	Do.
55	V	do	Do.
56	4	do	Do.
57	4	do	Do.
58	3	do	Do.
59	3–4	do	Do.
60	4	do	Do.
61	4	do	Do.
62	G	do	Do.
63	1	do	Do.
64	2	do	Do.
65	3	do	Do.
66	4	do	Do.
67	G	do	Do.
68	V	do	Do.
69	1	Quartz diorite	Jo Reynolds mine.
70	2	do	Do.
71	3	do	Do.
72	3	do	Do.
73	3	do	Do.
74	4	do	Do.
75	4	do	Do.
76	2–3	do	Nabob mine.
77	3	do	Do.
78	3–4	do	Do.
79	G	do	Do.
80	1	do	Jo Reynolds mine.
81	3	do	Do.

See footnotes at end of table.

TABLE 1.—*Samples examined by chemical analysis and semiquantitative spectographic analysis*—Continued

Sample	Altered-rock zone	Rock type	Location
82	1–2	Biotite gneiss	Widow Woman mine.
83	3	do	Do.
84	4	do	Do.
85	V	do	Do.
86	4	do	Cherokee mine.
87	4	do	Do.
88	V	do	Do.
89	4	do	Do.
90	4–V	do	Do.
91	3	do	Do.
92	3–2	do	Do.
93	1	do	Banta Hill mine.
94	2	do	Do.
95	3	do	Do.
96	4	do	Do.
97	V	do	Do.
98	1–2	Biotite-quartz gneiss	Cherokee mine.
99	3	do	Do.
100	4	do	Do.
101	V	do	Do.
102	1–2	do	Do.
103	3	do	Do.
104	4	do	Do.
105	4–V	do	Do.
106	1–2	do	Do.
107	3	do	Do.
108	4	do	Do.
109	G	do	Do.
110	V	do	Do.
111	1	Quartz-monzonite porphyry	Banta Hill mine.
112	2	do	Do.
113	3	do	Do.
114	4	do	Do.
115	1–2	Bostonite	Phoenix mine.
116	3–4	do	Do.
117	G	do	Do.
118	V	do	Do.

[1] Composite sample.
[2] Dump sample.

More than 500 samples were collected and studied during the investigation. These samples represent a diverse group of altered rocks related to a variety of geologic structures within an area of about 40 square miles. Analytical data are presented here for the representative samples listed in table 1.

Because preservation of natural moisture is desirable in samples containing a high proportion of clay minerals, the samples collected were placed in watertight cardboard containers and sealed by tape until ready for laboratory examination. These containers were not sufficiently airtight to prevent dehydration of the moist underground samples for more than 3 months; polyethylene plastic bags were found to be effective liners for the cardboard containers and retained natural moisture for longer periods of time.

LABORATORY STUDIES

The samples were studied in the laboratory with the petrographic microscope and X-ray diffractometer, and were analyzed spectrographically and chemically. Sample preparation and study for each different type of altered or fresh rock varied slightly.

Both thin section and oil immersion methods were used in optical studies. Friable rocks which were to be thin sectioned, including those containing clay minerals, were impregnated with Canada balsam before being cut by a diamond saw in a kerosene oil bath. A useful, undisturbed, and unfractured specimen of friable clay-mineral rock for thin sectioning can be obtained readily when natural moisture is preserved; when soft and moist, many of the altered rocks can be sliced easily with a knife. Modal analysis of the rocks was by means of a point-count microscope stage and tabulator. The area studied commonly was small because the feature observed was small or the slide was difficult to make; therefore the number of counts are recorded on the tables that show modal analyses. The sodium cobaltinitrite stain for the identification of potassium-bearing minerals in thin sections (Chayes, 1952) and in rock-slab specimens, also a very useful technique, was used in the study of these rocks.

X-ray diffraction was used most commonly to study the clay minerals and micas. Powder samples of clay minerals were prepared for use on a Phillips Norelco diffractometer with recording strip chart. The sample was scanned at the rate of 2° per minute between the angles 2° and 40° 2θ (a range of $d=44$ to 2.25A) at 35 KVP, 20 m.a., ¼° and 0.006-inch slits, but with variable full-scale deflections of 120 to 800 counts per second. A Phillips powder camera was used in a study of altered mica and K-feldspar.

For the most part bulk samples of the clay minerals in rock units formed from altered rock minerals were studied in preference to selected microsamples of clay minerals resulting from the alteration of particular minerals in the rock. A limited number of microsamples were examined also. While microsamples reveal important information about details of mineral alteration and are worthy of future study, this investigation was concerned primarily with the chemistry and mineralogy of altered wallrock units.

The procedure for the separation of clay-mineral assemblages from altered rock samples and their preparation for X-ray examination follows. The softened rock was crushed by hand in a mullite mortar; when a jaw crusher was used on less altered samples the fine fraction was checked for tramp iron. The crushed sample was dry ground to a powder in a rotating pebble mill using flint pebbles. The resulting powder was dispersed in distilled water. No wetting or dispersing agent normally was used; occasionally a small amount of dilute NH_4OH was needed. Clay-mineral separation was effected by gravity according to Stokes Law, and the -2μ clay mineral size fraction was pipetted off. The ground rock was redispersed several times (5–6 average) and the sedimentation technique repeated to obtain a representative sample. The accumulating clay mineral was kept from drying, the resulting total sample was dispersed as a thickened slurry, and a small sample was pipetted onto a glass microscope slide. When dry a clay-mineral aggregate consisting of subparallel flakes was available directly for X-ray study. The clay slides were reversed (Schultz, 1958, p. 367), using double-coated pressure-sensitive tape and X-rayed again where differential settling was noted. These glass-slide samples were amenable for direct ethylene glycol treatment using the vapor pressure method (Brunton, 1955), and for moderate heat treatment of the sample.

The following notes concern the selection and size of representative fresh and altered rock samples used for chemical and spectrographic analyses. The selection of a sample representative of each fresh and altered rock type was attempted; the individual clay-mineral subzones could not be sampled systematically. Although the successive phases characteristic of fresh to most altered rock can be recognized megascopically, at most places they are relatively small features. Where the layered rocks have segregations of minerals, it is difficult to obtain a representative sample. In one case the sample was a composite of the rock type within a mine from along the strike of a single zone over a distance of tens of feet. Samples selected generally weighed 500 grams or more. A limited number of standard rock and rapid rock analyses were made on representative rock suites, and semiquantitative spectrographic analyses were made on splits from these samples. Semiquantitative spectrographic analyses also were made for additional similarly prepared suites of rock samples.

Other data were gathered through mineral separations, rock gravity determinations, powder pH measurements of dry ground rock, and mine and ground water pH measurements. Mineral separations were by heavy liquid and water sedimentation, ultrasonic and magnetic methods, and by hand picking using a binocular microscope. Bulk densities of rock were determined on wax-coated samples following the method employed by a Geological Survey rock analysis laboratory (L. C. Peck, written communication, 1957). Measurements of pH were obtained by a glass electrode pH meter. The powder pH of rocks was made on a rock powder dispersed for 24 hours in distilled water. All other pH measurements were made in the laboratory as soon as possible after sample collection.

GEOLOGIC SETTING

The central part of the Front Range mineral belt consists chiefly of folded metasedimentary rocks and several types of igneous rocks of Precambrian age, some of which are metamorphosed. These rocks are intruded by small plutons and dikes of early Tertiary age. Rocks of both ages are cut by many intersecting faults that contain gold- and silver-bearing base-metal sulfide vein deposits. Uranium is an important constituent of some veins.

PRECAMBRIAN ROCKS

The principal Precambrian rocks of the region, listed in the order of their relative ages, are given in table 2, which also gives the equivalent names used in other reports. The metamorphic gneisses are largely metasedimentary rocks that belong to the amphibolite facies; the granitic rocks are intrusive igneous rocks that range in composition from quartz diorite to granite. Most of the rocks consist principally of quartz, K-feldspar, sodic or intermediate plagioclase, and biotite or hornblende, but in different proportions; thus they reacted similarly during alteration by hydrothermal solutions. The less abundant, iron-rich (and K-feldspar-deficient) rocks, such as quartz diorite and associated amphibolite, lime-silicate gneiss, and skarn, are less stable in the hydrothermal environment than are the more abundant, relatively iron poor, K-feldspar-rich rocks.

TABLE 2.—*Principal Precambrian rock units, central part of Front Range mineral belt, in order of probable relative age*

[Rocks are listed from youngest to oldest, except as noted]

Rock units	Probable origin	Remarks
Pegmatite	Igneous	
Granitic rocks:		
Biotite-muscovite granite	do	Equivalent to Silver Plume granite at Silver Plume, Colorado. (K–A age 1280 m.y., Davis and others, 1958)
Quartz diorite and associated amphibolite.	do	
Granodiorite	do	Probably equivalent to Boulder Creek granite.
Granite gneiss and pegmatite	Uncertain	Also constitutes felsic layers in migmatite.
Gneissic rocks:		
Microcline-quartz-plagioclase-biotite gneiss.	Metasedimentary.	Age relations among these rock units are unknown. Most of the rocks previously were grouped in the Idaho Springs formation. The microcline-quartz-plagioclase-biotite gneiss is the granite gneiss of Bastin and Hill (1917); it was referred to as quartz monzonite gneiss in previous reports (Sims, 1956; Tooker, 1956).
Cordierite-amphibole gneiss	do	
Biotite-quartz-plagioclase gneiss.	do	
Sillimanitic biotite-quartz gneiss.	do	
Lime-silicate gneiss	do	
Skarn and related rocks	do	
Amphibolite	do	
Quartz gneiss	do	

The metasedimentary rocks that underlie a large part of the central Front Range are well foliated, medium grained, and range in color from light to dark gray. The most abundant of these rocks, the microcline-quartz-plagioclase-biotite gneiss, is a light- to dark-gray, medium-grained gneiss that occurs as a thick, folded layer in the Central City (Sims, Drake, and Tooker, 1963) and Idaho Springs districts, and as thinner layers in other areas. The other less abundant metasedimentary rocks are mainly biotite gneisses that constitute an interlayered lenticular sequence of irregular thickness. The biotite gneisses may contain moderate or abundant granite gneiss and pegmatite, commonly as thin layers; these rocks, which are typical migmatites, are highly variable in composition and in thickness.

Granodiorite occurs as a large irregular mass in the Chicago Creek area (Harrison and Wells, 1959), and elsewhere as smaller, lens-shaped bodies. It is a competent, medium- to dark-gray, medium-grained rock composed of quartz, plagioclase, K-feldspar, biotite, and a minor amount of hornblende.

Quartz diorite and amphibolite are found chiefly in the Lawson-Dumont-Fall River district, but do not crop out well because they weather rapidly. The quartz diorite is dark gray, medium to coarse grained, and locally foliated; it is composed of plagioclase, hornblende, biotite, and quartz. The amphibolite is a black, medium- to coarse-grained rock composed principally of hornblende; it is generally not foliated.

Biotite-muscovite granite occurs in stock, dike, and sheetlike bodies throughout the area studied, but it is most abundant in the Freeland-Lamartine area. The granite is a medium-gray to salmon-pink generally medium grained rock with an equigranular to seriate-porphyritic texture. The rock is composed of sodic plagioclase, K-feldspar, quartz, biotite, and muscovite. Large K-feldspar crystals locally produce a weak planar structure; in most of the fine-grained facies the biotite flakes form an indistinct foliation. Mortar structure was also observed in some areas.

Sills, dikes, and podlike bodies of granite pegmatite cut all the older rocks. The pegmatite is medium to coarse grained, inequigranular, and composed chiefly of K-feldspar, quartz, plagioclase, and mica; biotite and magnetite are abundant locally.

TERTIARY ROCKS

Abundant, small, irregular intrusive dikes and plutons of radioactive porphyritic rocks of early Tertiary age cut the Precambrian rocks and are preore in age. The intrusives are particularly common in the metasedimentary gneisses, especially in the thicker and more widespread units of biotite gneiss. Wells (1960) has classed them according to petrographic character, relative ages, and geographic distribution into four groups: (1) hornblende granodiorite (oldest), (2) leucocratic granodiorite, (3) quartz monzonite, and (4) bostonite (youngest). The hornblende granodiorite group occurs only in the western part of the area; the quartz monzonite and leucocratic granodiorite groups are more abundant in the eastern part of the area, and form small irregular plutons and radial dikes. The bostonite occurs throughout the area, mostly as dikes that at places have narrow chilled margins; these dikes are somewhat finer grained than the older rocks.

Biotite-quartz latite dikes occur locally in the southeastern part of the area where they crosscut veins formed during the main period of ore formation and thus are the youngest Tertiary intrusive rocks.

STRUCTURE

The gneissic Precambrian rocks are folded and cut by faults and joints. These structures, together with primary lithologic features, have produced inhomogeneities in the rocks that have caused some variations in the character and extent of rock alteration. The folds, which trend northeastward, range from broad, symmetrical anticlines that have subsidiary open folds along their flanks, such as those in the Central City district (Sims, Osterwald, and Tooker, 1955), to complex tight or overturned folds, such as those in the Freeland–Lamartine area (Harrison and Wells, 1956). Cataclastic structures and younger northeast-trending folds are superposed locally on the earlier folded rocks (Moench, Harrison, and Sims, 1954; Sims, Moench, and Harrison, 1959).

The faulting took place at two separate times. An early period of fracturing that preceded the emplacement of the Tertiary intrusive rocks, produced northwest- and north-northeast-trending faults, and a later period of fracturing following the emplacement of the igneous rocks yielded east-, east-northeast-, and northeast-trending faults. The older faults may have formed in Precambrian time, but many were rejuvenated in Laramide time; the younger faults formed during the Laramide orogeny in early Tertiary time. Metallization followed the formation of the younger faults.

Except for some of the early northwest-trending faults, movements along the fractures generally were small. Recurrent movements took place along many faults during alteration and metallization (Sims, 1956; Harrison and Wells, 1959). These movements provided local open spaces in which the ore and gangue minerals were deposited.

In some parts of the region the major east- and northeast-trending faults, which were the principal channelways for hydrothermal solutions, are essentially parallel to the folds and to the metamorphic fabric of the country rock. Consequently, solution movement upward along the faults and outward into the wallrocks was partly controlled by these previously established structural features. The solutions were able to migrate a significant distance into the walls only where the fault zones provided a wide zone of broken rock, or intersected prominent joints or rock layers composed mostly of iron- and calcium-rich silicates.

MINERAL DEPOSITS

The ore deposits of the mining districts in the central part of the Front Range mineral belt are pyritic quartz veins, chiefly valuable for their gold and silver content, but also containing important but variable amounts of copper, lead, zinc, and uranium. The ratio of Pb^{207}:U^{235} in pitchblende for the Central City area indicates an absolute age of 59±5 million years or an early Tertiary age (Eckelman and Kulp, 1957, p. 1128).

The veins are hydrothermal fillings of fault fissures, and are similar in mineralogy, texture, and structure to deposits classified by Lindgren (1933, p. 530) as mesothermal. They range from simple filled-fissures with well defined, smooth, sharp walls to complexly branching lodes consisting of subparallel or intersecting fractures. The veins average 1 or 2 feet in width, but some are several feet wide. The ore shoots were deposited in structurally controlled open spaces along the fractures (Sims, 1956; Sims, Drake, and Tooker, 1963; Harrison and Wells, 1956).

The veins throughout the region contain the same suite of minerals, but they are present in different proportions from district to district. The common primary metallic minerals are pyrite, sphalerite, galena, chalcopyrite, and tennantite. Other minerals that are abundant locally are pitchblende, enargite, marcasite, polybasite, pearceite, tellurides of gold and silver, and free gold or electrum. The dominant gangue is silica in several forms, but various carbonates of the calcite group are common, and fluorite and barite occur locally. The common secondary minerals are chalcocite, covellite, and hydrous iron oxides.

Although vein filling took place during a single period of mineralization, the ore and gangue minerals were deposited in 2 main stages in most veins and in 3 stages in others (Sims, 1956, p. 745–746). The dominant mineralization, which yielded the sulfide ore minerals, began with the deposition of pyrite and quartz (pyrite stage) and concluded with the deposition of base-metal sulfide and gangue minerals (base-metal stage). Pitchblende and sparse pyrite and quartz were deposited in some veins during an early pyrite stage that preceded the base-metal sulfide mineralization; gold tellurides were precipitated locally following the base-metal stage. Vein structures were reopened periodically throughout the area separating stages of mineralization; more localized fracturing took place during the stages of filling.

The veins have been classified into two distinct types (Sims, 1956, p. 744), based on variations in the quantities of the principal vein-forming minerals. One type (pyrite veins) contains abundant pyrite and variable amounts of base-metal sulfides; the other principal type (galena-sphalerite veins) contains abundant galena and sphalerite and sparse pyrite. Transitional veins contain pyrite, substantial quantities of galena and sphalerite, and in addition, considerable amounts of copper minerals.

A regional concentric zonation of metallic minerals is indicated by the distribution of the different types of veins (pl. 1). The pyrite veins that are largely devoid of base-metal sulfides occur in a large, irregular core area which is designated the central ore zone on plate 1. This zone is in turn surrounded by a discontinuous area, the peripheral ore zone, which contains predominantly galena-sphalerite veins. The veins that contain substantial quantities of both pyrite and base-metal sulfides occur in an intermediate area or zone between the two principal zones. In the Freeland–Lamartine district (Harrison, 1955) and in other places, longitudinal zonation along a single vein or a vein system is also recognized. Although a depth zonation also may be present, it is not as evident as longitudinal and concentric zonation. Most of the mines in the area expose veins for relatively short distances down the dip. The Freeland and Lamartine veins, however, are exposed for at least 2,000 feet, and Harrison and Wells (1956, p. 87) noted that on the north-northeast Freeland vein, galena occurred near the surface, but not with pyrite and gold ore at depth.

ALTERED WALLROCKS

The alteration of rocks into zones about vein structures may vary owing to the type and proportion of rock minerals present and to the vein and rock structures present as well as to the physical and chemical character of the altering solutions—whether they be hydrothermal or supergene.

HYDROTHERMALLY ALTERED WALLROCKS

Fresh and hydrothermally altered wallrocks along veins in the area characteristically have a sequence of mineralogical and textural zones that can be recognized megascopically. The fresh rock, zone 1 of this report, grades veinward through weakly argillized rock, zone 2, to strongly argillized rock, zone 3, to sericitized rock, zone 4.

Exceptions to this zonal sequence are rare in the central part of the Front Range mineral belt, but where they do exist they can be related directly to differences in the primary mineralogy of the host rock. Most of the rocks in the region contain quartz, K-feldspar, and plagioclase as essential minerals. These rocks nearly always alter to a soft outer zone in which clay minerals are dominant, and a hard, bleached, sheared sericite-quartz-K-feldspar zone adjacent to the vein. Some of the wallrocks that are deficient in quartz and K-feldspar and which contain an abundance of iron-bearing minerals and plagioclase alter to a soft rock instead of to a hard sericitized rock adjacent to the vein. The soft rock contains greenish clay and at most places a secondary biotite. Another exception to the usual alteration pattern is evident where plagioclase and mica are absent in the host rock; these rocks alter to a montmorillonite-rich phase.

Zone 1, fresh rock.—Fresh rock, as used in this report is characterized in hand specimen by the clear, reflective appearance of the constituent minerals, and in thin sections by the lack of observed alteration of plagioclase, biotite, and hornblende. The plagioclase crystals in quartz diorite, however, may be altered (deuterically?) to clay minerals, and a clouding effect along a crystal boundary, crack, or an irregular patch within the crystal area commonly is observed. Sparse and irregular sericitelike inclusions also occur in many of the "fresh" rocks. The transition from zone 1 into zone 2 is gradual, but the boundary is placed where clay minerals formed from plagioclase definitely become megascopically recognizable. Evidence of widespread deuteric or early pervasive propylitic types of rock alteration, characteristic of many areas, perhaps are obscured here by regional metamorphism.

Zone 2, weakly argillized rock.—The rock is hard, and the original structures and textures are preserved. The plagioclase grains, white to light gray, have a dull, chalky luster that is due to incipient alteration on cleavage and fracture faces. Hornblende, when present, also is partly altered to gray-green montmorillonite, illite, and mixed-layer clay. At the variable, often indistinct boundary with zone 3, the rock has been partly softened owing to the alteration of hornblende and plagioclase. Biotite, K-feldspar, and quartz are unaltered.

Zone 3, strongly argillized rock.—Strongly argillized rock is soft, composed of light-green to white clay mineral, quartz, biotite, and K-feldspar, and it retains much of the original structure and texture. Clay minerals have nearly completely replaced plagioclase and hornblende crystals, and biotite may be incipiently altered to clay minerals, but quartz and K-feldspar remain unaltered.

The strongly argillized zone may be subdivided into several subzones based on the predominant clay mineral within a mixed clay assemblage as determined from X-ray diffraction data. The outermost subzone, a, is montmorillonite-rich rock that grades veinward into a less distinct subzone, b, of illite-rich rock, that in turn grades into the kaolinite-rich rock subzone, c, adjacent to zone 4. The boundary of zone 3 and zone 4 is generally sharp, and is marked by distinct changes in texture and mineral content.

Zone 4, sericitized rock.—The sericitized rock, which is the most intensely altered wallrock bordering the veins, is hard and bleached light greenish gray or white. Original structures and textures are mostly obliterated, and shearing and recrystallization are common features on the veinward side of the zone. The rock is composed chiefly of sericite, quartz, and K-feldspar, and contains scattered crystals of pyrite that appear to be localized by altered biotite.

SUPERGENE ALTERATION AND ROCK WEATHERING

Supergene solutions derived from the weathering of the sulfide vein minerals probably have modified the minerals of the altered wallrocks as well as the veins, but the products of supergene alteration are not readily distinguished from those of hydrothermal alteration in the zones of argillic and sericitic alteration. Rock weathering processes in wallrocks may have also affected the altered and fresh rocks close to the surface, giving rise to some of the same minerals that were deposited by hydrothermal solutions.

The principal observable effect of supergene solution alteration is the widespread oxidation in the vein of some iron-bearing minerals to limonite, less commonly the conversion of copper-bearing sulfides to chalcocite, and of uraninite to sooty pitchblende and other secondary uranium-bearing minerals. Fournier (written communication, 1960) points out also that supergene alteration is not limited to oxidation-reduction reactions. A very important aspect of supergene alteration, borne out by Hemley's (1959) studies, is the ability of relatively cool alkaline solutions to leach alkalies from silicates and thus form montmorillonite, mixed-layer clay, and even kaolinite. The oxidation of pyrite to form a very acid solution enhances alteration and favors the formation of supergene kaolinite. Biotite may lose iron as a result of an oxidation reaction, but in nonoxidation reactions supergene chlorite and vermiculite are formed. The mineralogical and chemical character of the wallrock host minerals as well as of the supergene solution therefore affects the results of supergene activity in a manner very similar to that of weak hydrothermal activity. Thus only where we recognized the results of oxidation-reduction reactions were we able to distinguish supergene action with any certainty.

The depth to which supergene solutions are known to have been active varies in the districts, but averages 50 to 150 feet in most mines; however, Bastin and Hill (1917, p. 151) observed chalcocite as much as 700 feet below the surface in some mines.

Some water-filled mine workings ultimately connect with old drainage tunnels (such as the Argo tunnel) through caved workings or open-vein structures and thus provide for a locally pronounced downward circulation of oxidizing solutions below the water table.

Where readily oxidizable iron in the veins also is available, as at the E. Calhoun mine, limonite and limonite-stained clay minerals occur in the vein and along foliation planes that extend into sericitized microcline-quartz-plagioclase-biotite gneiss walls for a considerable distance below the surface; but in the deepest parts of mine workings limonite and pyrite are observed only in veins. In contrast to these observations, the oxidation of uranium minerals in undrained and recently water-filled workings of the Carroll mine occurs close to the surface (Sims, 1956, p. 752); uraninite is found on the 228-foot level, but is altered to sooty pitchblende and metatorbernite on the 102-foot level. Pyrite is not as abundant a vein mineral here as in central and intermediate ore zone veins. Limonite-stained clay commonly forms a thick coat on vein walls and penetrates into the walls along foliation and joint planes of walls in the upper part of the vein, but is only a thin wall coating where present on the 228-foot level.

In contrast, many workings close to the surface are well drained and contain much evidence of supergene action in veins and wallrocks. The Mammoth mine workings contain sooty chalcocite and limonite in the vein, but microcline-quartz-plagioclase-biotite gneiss and bostonite dike wallrocks are relatively free of obvious supergene alteration. The R.H.D.-McKay shaft workings vein, and the amphibolite and biotite-quartz-plagioclase gneiss wallrocks, contain limonite as a prominent alteration product of iron-rich sulfides and silicate minerals, and metatorbernite presumably altered from pitchblende (Sims and Tooker, 1955).

Rock weathering at the surface produces weak argillic alteration characteristic of zone 2 hydrothermal and (or) supergene alteration. Studies of oxygen isotope ratios offer hope that in time we may distinguish one product from another.

WIDTH OF THE ALTERED WALLROCK ZONE

The width of alteration selvages is roughly proportional to the width of the veins unless modified by an unusual wallrock composition or vein-opening geometry. Veins in the central zone at Central City have wider zones of altered wallrock (table 3) than veins in the intermediate or peripheral zones.

Detailed measurements of the width of the altered zones was restricted to relatively few places. The miners carefully followed vein structures and excavated the sulfides and gangue minerals, and only as much of the sericitized-silicified wallrock as was necessary to continue the operation; crosscuts away from the veins are uncommon. Thus the widths given in table 3 more commonly are for small branch veins rather than for the major veins in the district.

TABLE 3.—*Widths of some small veins and their associated wallrock zones*

[Rock types are: A, microcline-quartz-plagioclase-biotite gneiss; B, granodiorite; C, metasedimentary gneiss]

Ore zone	Mine[1]	Rock type	Width		Ratio of altered rock to vein[3]
			Vein (feet)	Altered rock[2] (feet)	
Central	Mammouth (7)	A	0.3	4.0	13.3
	Hayes and Wheeler (8)	B	.05	1.6	32.0
Central to intermediate	E. Calhoun (9)	A	1.0	.84	.8
	Old Stagg (Mendick) (24)	C	.2	.37	1.3
Intermediate	Two Brothers (12)	A	.2	1.2	6.0
			3.5	7.6	2.1
Intermediate to peripheral.	Cherokee (10)	C	.2	1.5	7.5
			.25	1.2	4.0
Peripheral	Dumas-Kinney (1)	A	.6	3.2	5.3
	Carroll (4)	A	.6	4.0	6.6
	Essex (5)	A	3.0	8.0	2.6
			5.0	4.0	.8
	Banta Hill (10)	C	.2	.8	4.0
	M. and M.-Dixie (21)	B	.1	.3	3.0
		B	.02	.84	42.0
		B	.2	2.2	11.0

[1] Number refers to mine location shown on pl. 1.

[2] Measurement for alteration zone generally possible only on one side of vein; altered zone generally nearly symmetrical.

[3] Altered zone includes both sides of the vein.

The alteration selvage generally is wider in host rocks containing appreciable amounts of calcic plagioclase and hornblende than in the more common rocks that are felsic. The contrasts in the width of altered rock zones along veins in granodiorite and biotite-muscovite granite (table 4) illustrate this point. A generalization of the relative widths of altered wallrock zones in different types of country rock bordering a vein is shown in figure 2. The width of the openings along the vein fissures at the time of alteration may not have been of equal width at the time of metallization (p.90); accordingly, this factor may account for some of the anomalous altered rock to vein size ratios in similar kinds of rock such as occur in the Essex, M. and M. Dixie, and E. Calhoun mines (table 3).

TABLE 4.—*Estimated widths, in feet, of some minor branch veins and their altered wallrock zones*

[ND, not determined]

Host rock	Vein	Altered-rock zones			Mine
		4	3	2	
Granodiorite	0.05	0.2	0.2	1.2+	Hayes and Wheeler.[1]
Do	.02	.18	.16	.5+	M. and M.-Dixie.[2]
Do	.2	.4	.3	1.5+	Do.
Biotite-muscovite granite	1.0	.3	.2	ND	Diamond Mountain.[2]
Do	3.5	ND	--------	--------	Nabob.[2]
Do	.9	.5	ND	--------	Do.
Do	12.0	2.0	ND	--------	Jo Reynolds.[2]
Microcline-quartz-plagioclase-biotite-gneiss.	.1	.6	.15	.1	E. Calhoun, 6th level.[3]
Do	.01	1-.2	.5	.1-.2	Do.
Do	[4] 1.5	.16	.1	ND	Do.
Do	.2	3-.4	ND	--------	Do.
Do	.05-.1	.35	.1	.1	Do.
Do	.1	.35	.1	.1	Do.
Do	.1-.15	.12	.5	ND	Do.
Do	.3	.65	.2	ND	Do.

[1] Central ore zone

[2] Peripheral ore zone.

[3] Intermediate ore zone.

[4] Vein measured at ore stope.

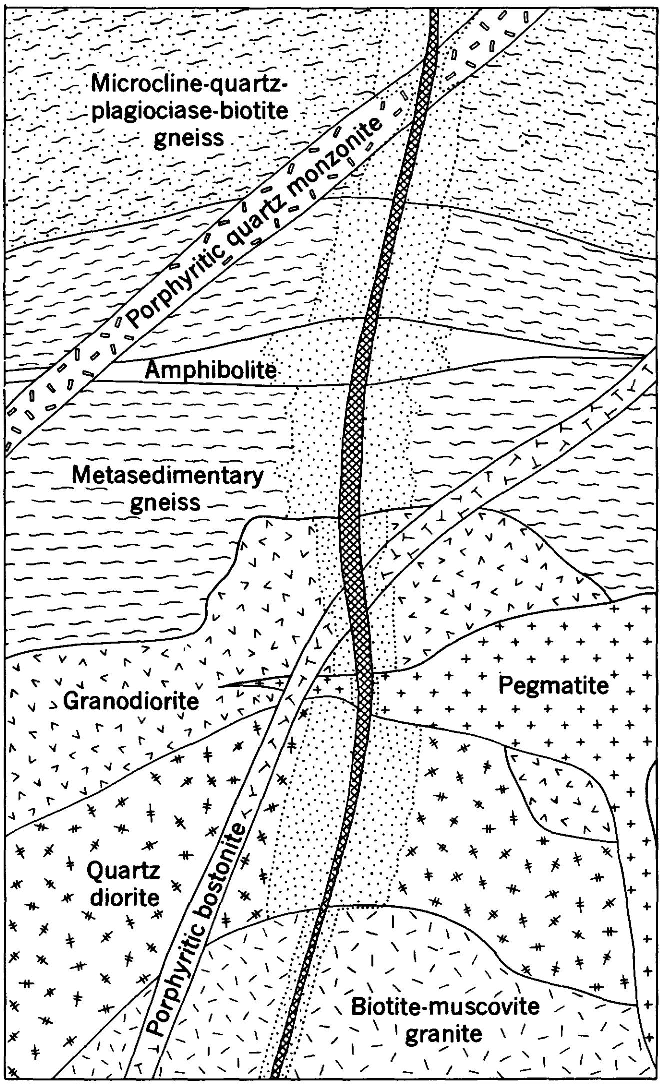


FIGURE 2.—Diagrammatic section showing comparative widths of alteration zones (stippled) in different types of rocks bordering a vein.

FACTORS THAT PRODUCE VARIATIONS IN THE CHARACTER AND DISTRIBUTION OF THE ALTERED WALLROCKS

Among the most obvious factors that produce variations in the character and distribution of altered wallrocks are differences in (1) the mineralogy and structure of the wallrocks, (2) the structure of the vein fissures, and (3) the hydrothermal and supergene environments of the veins.

Variation in the character of alteration is directly related to differences in the mineralogy of the wallrock and, consequently, to the areal distribution of a given rock type. Most of the rocks in the area studied consist essentially of quartz, K-feldspar, plagioclase, biotite, and hornblende. Thus, the altering solutions reacted with the same host minerals, but in different proportions in the different rocks. Biotite-muscovite granite, found mostly in the southwest part of the area, does not contain prominent zones of argillic alteration because most of the minerals forming this rock were not readily attacked by the altering solutions. Most altered biotite-muscovite granite is hard, and the altered rock zone is narrow. In contrast, quartz diorite and amphibolite, quantitatively only minor rocks in this region, contain only small amounts of quartz and K-feldspar but comparatively large amounts of plagioclase and hornblende, which alter readily to clay minerals. The resulting altered rock is soft, and the argillic zone commonly is broad.

Differences in the temperature, pressure, and composition of the hydrothermal solutions also resulted in variations in the distribution and thickness of altered wallrocks. In general, wider zones of altered wallrocks border veins in the central ore zone—the center of more intense hydrothermal activity—in the Central City district than in the intermediate or peripheral ore zones, but this generalization may not be strictly applicable to other parts of the central zone, as defined on plate 1. A factor that also may contribute to the notably wide alteration halos in the central zone at the Central City district is the presence of unusually wide and more persistent veins in this area which in turn permitted a greater and more direct flow of solutions than the smaller, less continuous fractures in other parts of the region.

The variations in the character and distribution of the altered wallrocks that are related to the competency and structure of the wallrocks are not easily separated from those related to the relative openness of fissures. Bastin and Hill (1917, p. 95) and later workers (Sims, Drake, and Tooker, 1963; Harrison and Wells, 1956, p. 83) have noted that the relative competency of wallrock exerted a marked effect upon the extent and character of the fractures. In a general way, fractures in relatively competent granitic rocks form complexly branching, persistent, open vein fissures, whereas fractures in the less competent schistose rocks, which were affected by similar stresses, are short, closely spaced, and commonly filled with gouge clay. Veins passing from granite to granitic gneisses into schistose gneisses at a low angle to the foliation generally split into a series of small branching fractures nearly parallel to schistosity and die out; veins that cut schistosity at nearly right angles, however, although not common, may persist for a substantial distance into the schistose gneiss.

In texturally homogenous competent rocks, such as in the microcline-quartz-plagioclase-biotite gneiss of the Central City district, the vein fissures are wide, closely spaced, and persistent, and correspondingly have wide alteration selvages. These features are especially evident in the central and intermediate mineralogic zones (pl. 1) but are less marked in the peripheral zone. In contrast, the texturally inhomogeneous, incompetent rocks, which often are complexly folded (such as the migmatites and metasedimentary rocks characteristic of the Freeland-Lamartine district), commonly contain only narrow, moderately spaced and mostly discontinuous fissures. Also, the brittleness of the competent rocks in Central City district fostered the formation of wide breccia zones, which permitted a pervasive penetration of altering solutions where the fissure veins were close spaced. The incompetent and therefore relatively impervious, foliated wallrock of the Freeland-Lamartine district restricted solution flow to the principal fissures. This has resulted in the development in the schistose rocks of narrow, longitudinal altered zones that stand in contrast to the wide, altered zones typical of those bordering the veins near Central City.

As may be expected, the altered zones commonly are wider at intersections of veins and along portions of the veins where movement along irregularities in the surface of the fissure has caused openings. Cross fissures are not common, but joints and foliation surfaces at places were paths for limited movement of solution outward from the vein fissures. However, alteration effects are generally weak along such cross-breaking features.

ALTERED ROCK MINERALS

PRIMARY AND SECONDARY MINERAL STABILITY

Quartz, K-feldspar, and muscovite are most resistant to weathering and hydrothermal alteration, and biotite, plagioclase, and amphibole are progressively less resistant and more quickly and completely altered. Generalized ranges and amounts of minerals in fresh and progressively more altered rocks in the central part of the Front Range mineral belt (fig. 3) indicate the relative susceptibility of minerals in these rocks to alteration. Generally, the relative susceptibilities of the minerals follow Goldich's (1938) sequence of weathering susceptibility. The ranges of minerals in individual rocks are given later in this report.

Quartz, SiO_2, commonly constitutes about 30 percent of the rocks, and occurs uniformly across the altered rock zones. It occurs in anhedral, strained, medium-size crystals, and is altered only in the sense that some of it is commonly recrystallized in zone 4, with a loss of undulatory extinction and the formation of smaller, complexly interlocking grains.

K-feldspar, a potassium-rich alkali feldspar but not a pure potassium feldspar, generally constitutes less than 10 percent of the rocks in which it occurs; it is chiefly microcline (fig. 4) in the Precambrian rocks and is orthoclase or cryptoperthite (fig. 5) in the Tertiary intrusive rocks. K-feldspars in microcline-quartz-plagioclase-biotite gneiss and in pegmatite commonly are slightly perthitic, and according to Sims (written communication, 1960) contain approximately 20 percent Na-feldspar in blebs and solid solution. In general, the K-feldspar occurs as white to salmon-pink, subhedral to anhedral, tabular, well twinned phenocrysts ranging in size from 0.02 to 1.5 mm. Microcline also occurs in blebs in antiperthite crystals. K-feldspar apparently is unaltered through zones 1, 2, 3, and 4 (fig. 6). At places, small amounts of clay minerals incipiently and randomly replace K-feldspar along crystal margins, cleavages, and fractures. However, such replacement may be of supergene origin. X-ray examination of hand-picked white clay replacing K-feldspar in a pegmatite dike revealed the presence of kaolinite, illite (or sericite), and a minor amount of montmorillonite. Sparce amounts of anhedral adularia(?) enclosed by clay minerals occur in zone 4 in some of the rocks; however, this phase has not been identified as a prominent product of wallrock alteration in these areas.

Muscovite, $K(OH)_2Al_3Si_3O_{10}$, present in biotite-muscovite granite and sillimanite-bearing metasedimentary gneisses in amounts less than 5 percent, is unaltered in nearly all of the alteration zones. This 2M, dioctahedral mica occurs in elongate subhedral cleavage plates (fig. 4*A*), which may be partly recrystallized to coarse sericite grains in zone 4.

Biotite, which according to Winchell and Winchell (1951, p. 373) may have a general composition $K(OH)_2(Mg,Fe,Al)_3(Si,Al)_4O_{10}$, makes up as much as 20 percent of some rocks, and has an intermediate range of stability in comparison with other rock-forming minerals when altered in the hydrothermal environment. The biotite is trioctahedral (table 5), and occurs as elongated cleavage plates (fig. 7) that are generally brown to black, or less commonly, a greenish brown. Biotite alters to a variety of alteration products. In zone 4 it is often completely and pseudomorphically altered to clay mineral-sericite aggregates. In many rocks the boundary between zones 3 and 4 is sharp; on one side of the contact, biotite is relatively unaltered, but on the other side it is completely altered.

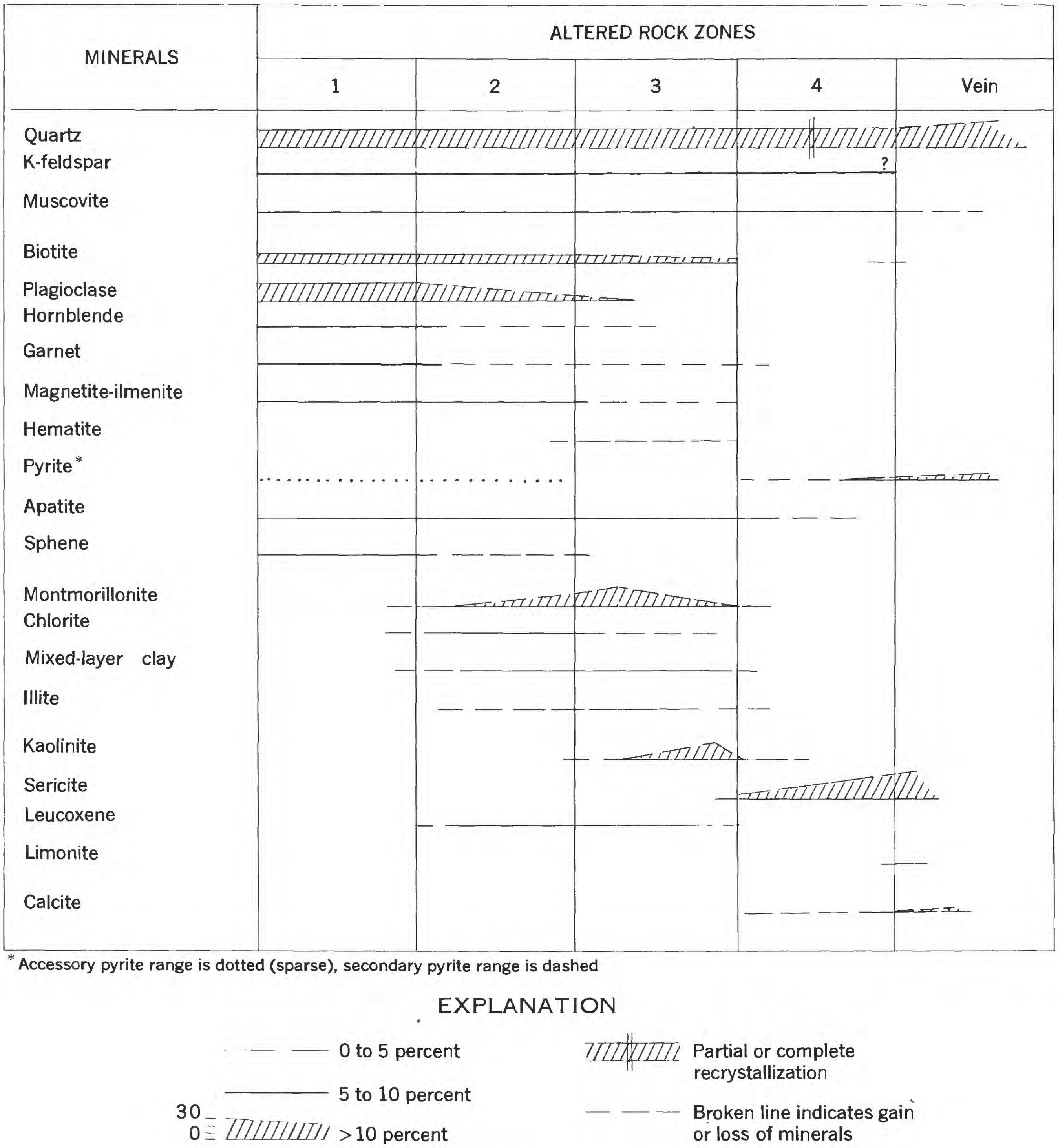
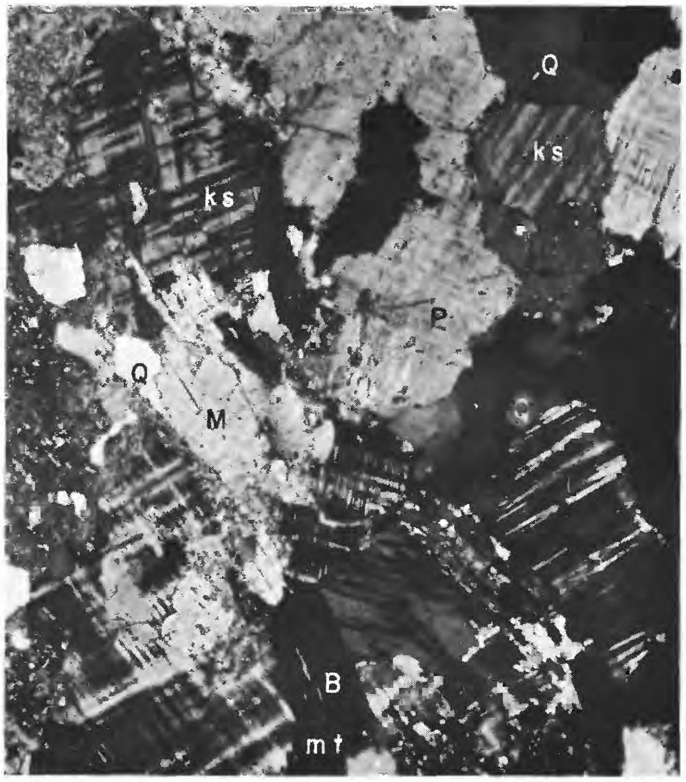
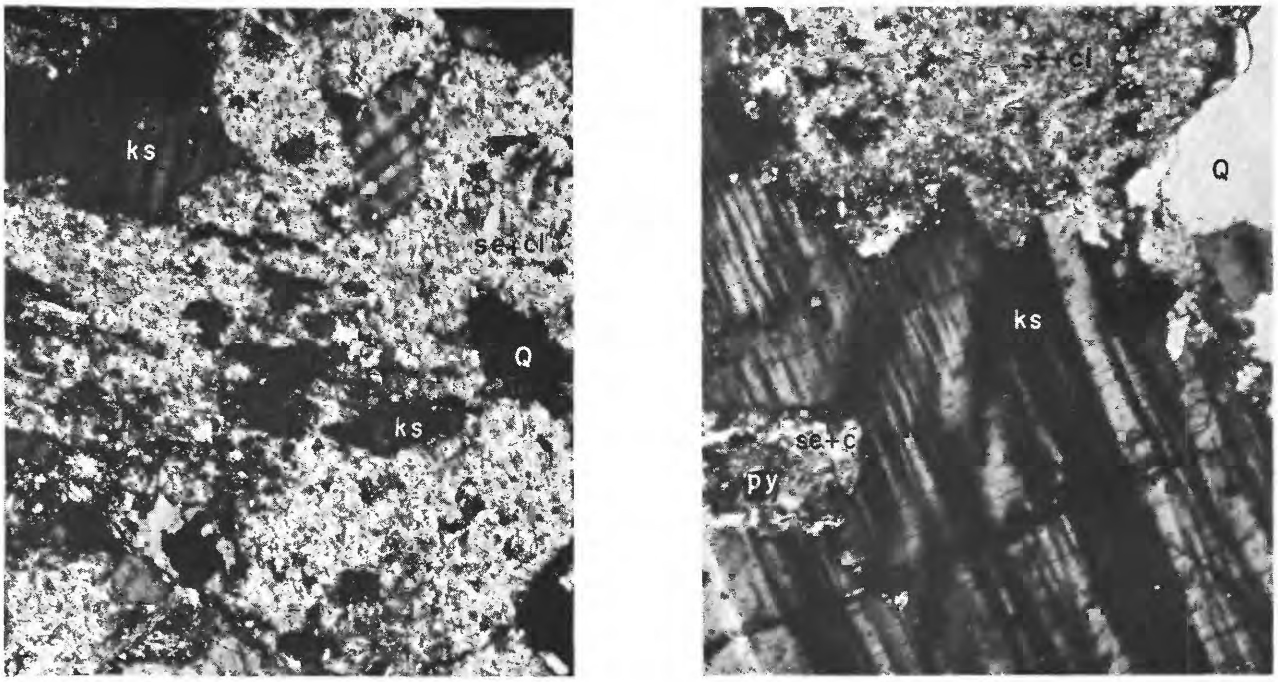


FIGURE 3.—Generalized ranges and amounts of the most common minerals in fresh and altered wallrocks.



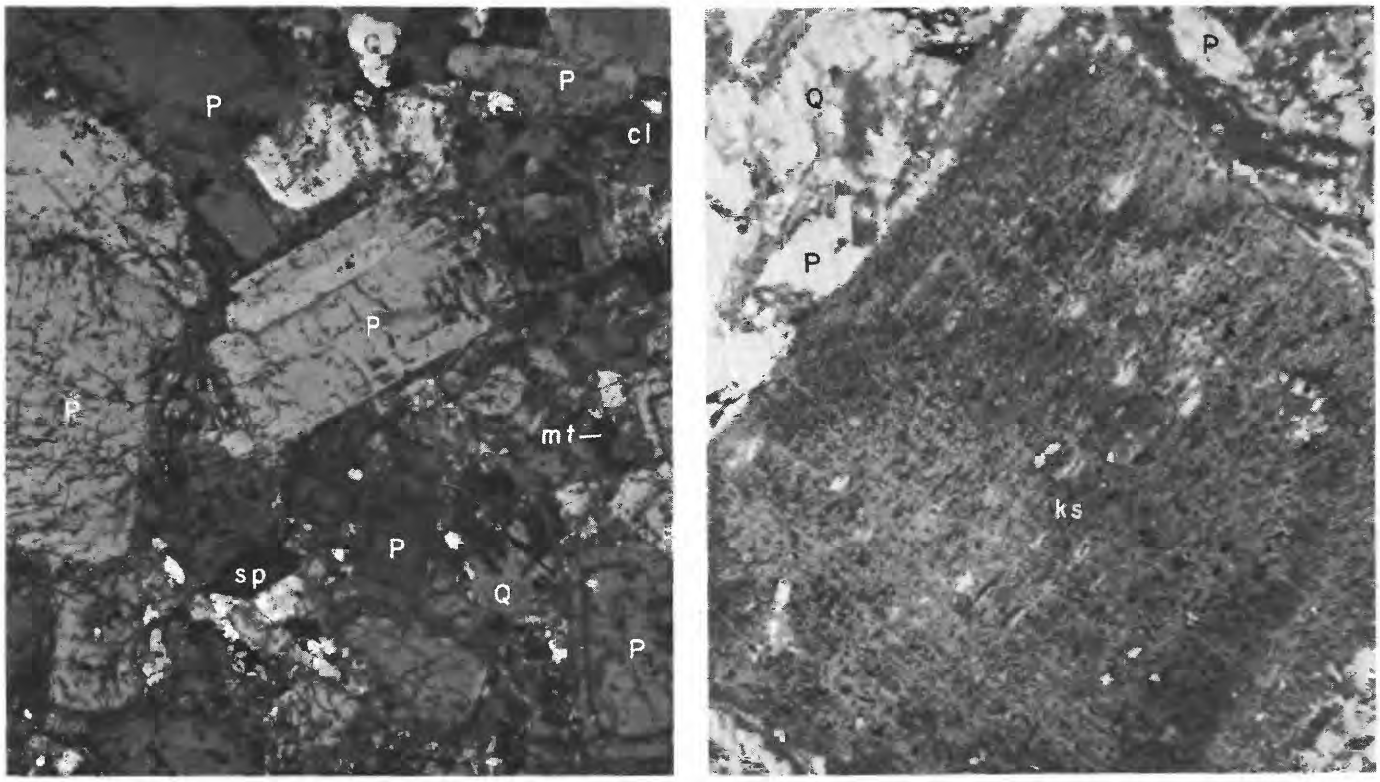
A, Zone 2–3; biotite-muscovite granite, Jo Reynolds mine; quartz (Q), microcline (ks), biotite (B), muscovite (M), and magnetite-ilmenite (mt) are unaltered; plagioclase (P) is altered partly to clay minerals, and biotite (not shown) is altered partly to chlorite; crossed nicols, 46×.



B, Zone 3–4; microcline-quartz-plagioclase-biotite gneiss, Essex mine; quartz (Q) and microcline (ks) are unaltered, plagioclase (P) and biotite (B) are altered to clay minerals (cl), sericite (se) and an unidentified iron oxide; crossed nicols, 90×.

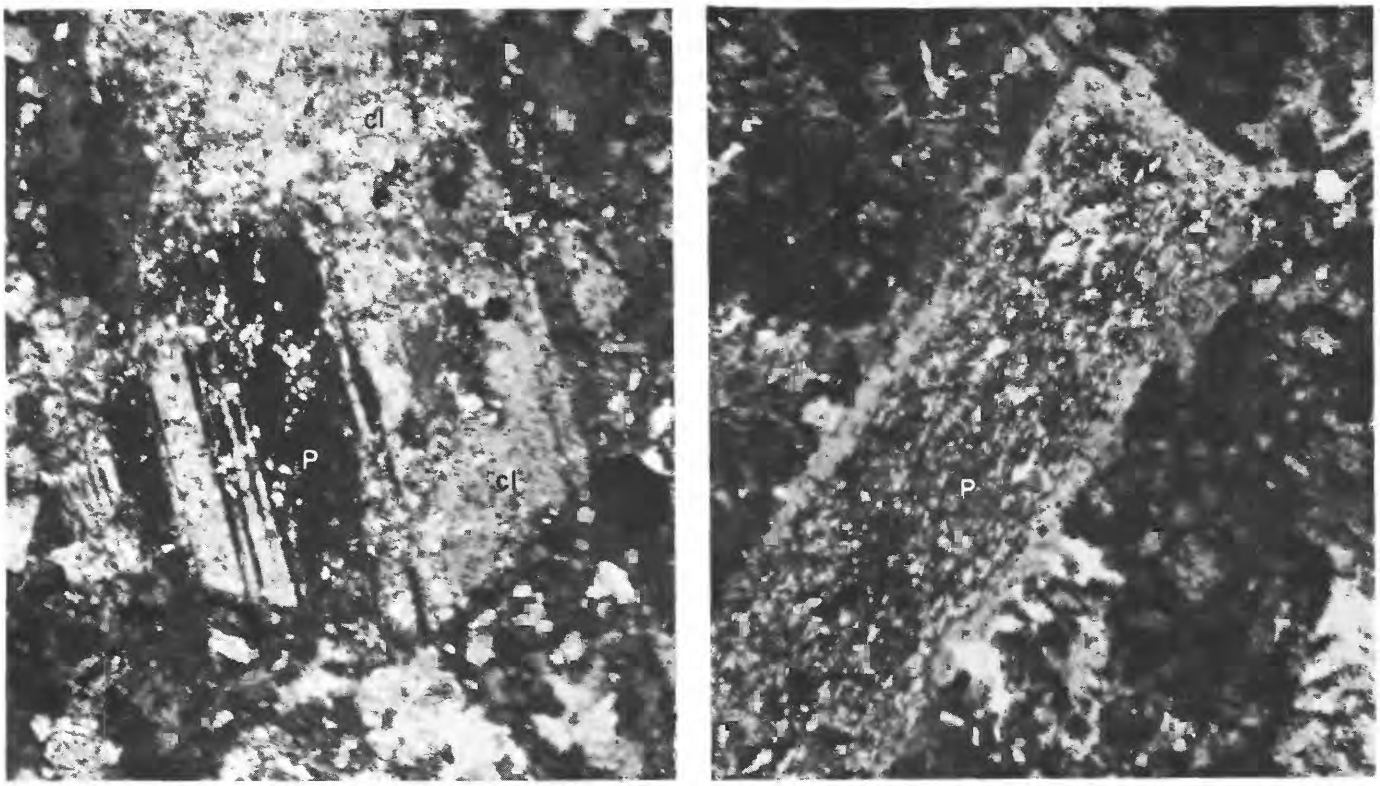
C. Zone 4; biotite-muscovite granite, Nabob mine; quartz (Q) and microcline (ks) are unaltered, plagioclase (P) is altered to sericite (se) and minor clay minerals (cl), biotite (B) is altered to sericite (se), pyrite (py), and calcite (c); crossed nicols, 76×.

FIGURE 4.—Photomicrographs of fresh and altered microcline-rich rocks.



A, Zone 1; porphyritic quartz monzonite, Silver Age mine; quartz (Q), plagioclase (P), hornblende (not shown) are unaltered phenocrysts; magnetite (mt), sphene (sp), and clay minerals (cl) are included in the groundmass; crossed nicols, 60×.

B, Zone 1; porphyritic bostonite, Diamond Mountain mine; unaltered K-feldspar (ks) phenocrysts in quartz (Q) and weakly altered plagioclase (P) groundmass; crossed nicols, 130×.



C, Zone 3-4; porphyritic bostonite, Banta Hill mine; plagioclase (P) phenocryst altered in part to clay minerals (cl); crossed nicols, 130×.

D, Zone 2-3; porphyritic bostonite, Banta Hill mine; altered zoned plagioclase (P) phenocryst in partially altered groundmass; crossed nicols, 130×.

FIGURE 5.—Photomicrographs showing the manner of alteration of phenocrysts in Tertiary intrusive rocks.

TABLE 5.—*X-ray data from biotite separated from zones of progressively more altered granodiorite, Hayes and Wheeler mine*

[Fe/Mn radiation: s, strong; m, medium; w, weak; vs, very strong; b, broad lines, d, diffuse lines

Sample in zone shown—													
1		2		3a		3c			4				
41		42		43		45a			45b		45c		
d angstroms	Index	d angstroms	Index	d angstroms	Index	d angstroms	Index	1M hkl	d angstroms	Index	d angstroms	Index	2M hkl
10.01	s(b)	10.10	vs	10.10	vs	10.10	vs(b)	001	10.01	vs	10.01	vs	002
		7.09	vvw						7.13	vw	7.19	vw	
									6.55	vw	5.98	w	
									5.00	m			004
4.64	w	4.60	w(b)	4.64	w	4.62	w	020	4.52	ms			021
											4.49	w(bd)	110
									4.24	m			
									4.12	vw			022
3.96	vw			3.93	vw	3.96	vw		3.96	vw			112
									3.79	m	3.76	w	023
3.70	vw	3.69	w	3.68	vw	3.69	w	$11\bar{2}$	3.67	w(d)	3.66	w	
		3.54	vw								3.51	m	$11\bar{4}$
				3.42	m				3.48	w			
3.37	vs(b)	3.37	vs	3.35	vs	3.37	vs	003	3.34	vs(b)	3.36	vs(b)	006
									3.23	m	3.23	w	114
3.17	w(d)	3.17	m	3.14	m	3.16	w	112			3.12	w	
3.05	vw	3.05	vw						3.07	vw			
									2.99	w	3.00	vw	025
2.94	w(d)	2.94	mw(d)	2.93	mw	2.94	w	$11\bar{3}$	2.91	w(b,d)	2.91	vw(d)	115
		2.84	vw						2.81	w	2.81	m(d)	$11\bar{6}$
		2.79	vw						2.77	vw			
2.73	w			2.71	w	2.72	vw	023	2.68	w	2.70	ms	
2.63	s	2.63	s	2.63	s	2.64	s	130, $20\bar{1}$	2.63	w	2.62	mw	
									2.57	mw(b)	2.57	vw	116
2.52	m(b)	2.53	m-w	2.51	m	2.52	m	004, 113					
								131, $20\bar{2}$			2.52	w(b)	008
		2.45	m	2.44	m-s	2.45	m-s	$13\bar{2}$, 201	2.45	vw(d)	2.42	w	$13\bar{3}$
		2.32	w	2.31	vw	2.31	vw	$22\bar{1}$	2.39	vw(d)			
2.28	vw(d)	2.28	vw	2.25	vw	2.27	w(d)	$20\bar{3}$	2.28	vw			
								220, 132	2.26	vw	2.22	w	041
		2.18	m	2.18	m	2.19	m-s	041	2.17	w	2.18	vw	$22\bar{3}$
2.11	ms								2.13	w			
2.01	ms(b)	2.05	m	2.02	ms	2.02	s	005	2.01	wm	2.02	w(d)	0010
		2.00	m	1.99	m	2.00	m	$20\bar{4}$, 133	1.99	wm	1.99	w	$13\bar{7}$
1.92	vw	1.92	w	1.91	vw	1.92	w	$13\bar{4}$, 203			1.91	w	
											1.89	vw	
											1.84	vw(d)	
									1.82	w			
									1.80	vw	1.80	vw	
									1.75	vw	1.74	vw	$13\bar{9}$
1.68	ms	1.68	m(b)	1.68	ms(b)	1.70	s-m	$13\bar{5}$, 204	1.68	w(d)	1.68	w(d)	150
											1.63	m	
1.55	ms(b)	1.54	m	1.54	s(b)	1.54	s	060	1.54	w	1.54	mw	060
1.53	w	1.53	w	1.52	w	1.53	vw	061, 330					
1.48	vw(d)								1.50	w	1.50	w	
1.44	vw(b,d)	1.44	w	1.43	w	1.44	w				1.45	vw	
1.36	mw(b)	1.37	w(b)	1.36	m(d)	1.37	w(b)				1.36	vw(b,d)	

Biotite within successive alteration zones is bleached from dark brown to light greenish brown, and subsequently to yellow to silvery white (fig. 7*B*, 7*C*). This loss in dark color is accompanied by a loss of pleochroism and birefringence. X-ray diffraction patterns (table 5) of a series of progressively more bleached biotite grains indicate that in zone 4 the white flakes contained spacings characteristic of both 2M dioctahedral and 1M(?) trioctahedral structures. Kaolinite (or perhaps chlorite) and montmorillonite also are present.

Locally, biotite flakes are replaced irregularly along certain cleavage planes by chlorite (fig. 7), which occurs either as discontinuous or patchy, randomly oriented aggregates of small crystals or as continuous thin plates. Under the microscope both the aggregates and plates of chlorite are characterized by low relief, low birefringence, and an index of refraction greater than balsam; this is in contrast to residual mica (degraded biotite) that has weak pleochroism, high relief, and strong birefringence. The iron liberated by this chloritization may be fixed as magnetite or other iron oxides in irregular streaks that also are parallel to the cleavage planes (fig. 7*E*).

The chlorite is altered subsequently to kaolinite and montmorillonite-illite, as shown in figure 7*E*, *F*. In zone 4 the chlorite and clay minerals are altered to sericite, and the iron oxides to pyrite (fig. 7*G*, *H*). Additional pyrite may be introduced subsequently by ore-forming solutions.

Some secondary biotite forms during the alteration process. In rocks such as amphibolite and biotite schist, small randomly oriented flakes of biotite are irregularly distributed through the soft-green montmorillonite-rich clay-mineral zone adjacent to the vein as noted on p. 38. These flakes appear genetically related to finer grained, more fragile, and light-colored mica (sericite?) crystals in the same rock. In garnet-quartz gneiss, narrow veinlets of secondary biotite

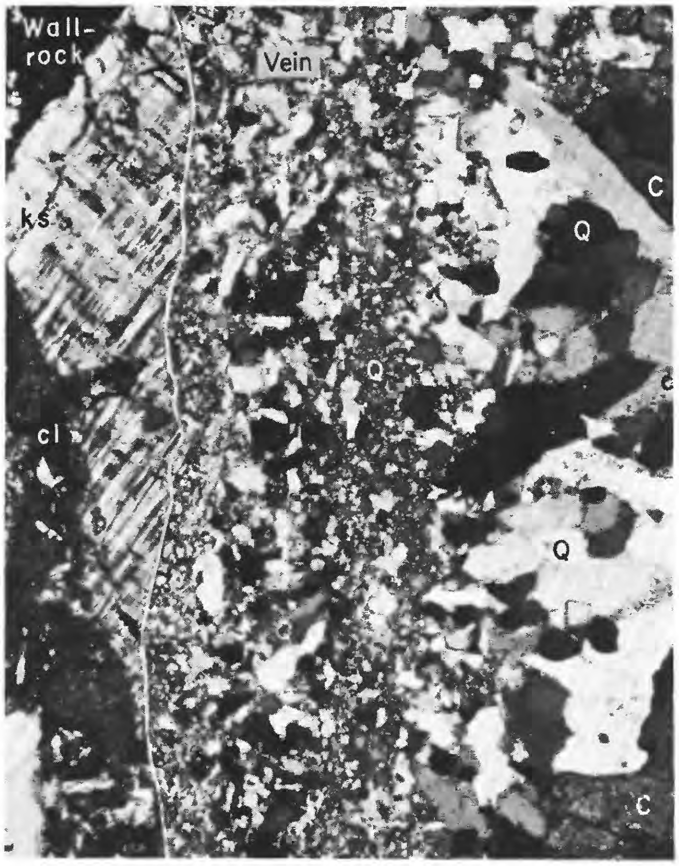


FIGURE 6.—Photomicrograph showing unaltered microcline in zone 4 along a vein in sheared biotite-muscovite granite, Nabob mine, crossed nicols, 60×. Note quartz crystal (Q) growth into vein (toward right margin of photograph) and calcite (C) filling spaces between quartz grains; sheared fine-grained quartz and clay minerals (cl) border the microcline (ks) on the vein wall.

flakes were observed along altered fracture surfaces in garnet crystals.

Plagioclase, generally within a range of composition of An_{18-30}, makes up as much as 35 percent of the host rocks, and is the least stable major rock constituent in the alteration environment. In the fresh rock, it occurs as white, subhedral to anhedral, prominently twinned, medium-sized grains. In the outermost fringes of alteration, the more calcic variety, An_{28-30}, is incipiently altered along twin plane traces on cleavage surfaces to patchy aggregates of clay (fig. 8 *C*, *D*). In the early literature (Bastin and Hill, 1917, p. 103) these aggregates are referred to as sericite or kaolin, but X-ray studies indicate that they are actually composed of a mixture of montmorillonite, illite, and random mixed-layer montmorillonite-illite clay minerals. In zone 2 plagioclase is altered to fine-grained aggregates of montmorillonite, other clay minerals, and some calcite. As observed in thin section, the clay minerals appear to be composed of several irregularly intermixed components that have variable optical properties and stain. Altered patches and layers increase in size and abundance until they coalesce (fig. 8), leaving a few clear and fresh "island" remnants, or cloudy, partly altered, and twinned crystals. The more sodic plagioclase alters less readily in zone 2 and is not completely altered until in zone 3. This phenomenon is readily observed in the zoned phenocrysts in some porphyritic bostonites (fig. 5). The calcic plagioclase centers of crystals are completely altered, whereas the sodic rims are less altered. Kaolinite and illite are more prominent products of alteration of plagioclase in the mixed-clay assemblage in zone 3 closer to the vein. The clay minerals are almost completely altered to sericite in zone 4. The sericite occurs as radial plumose aggregates, parallel bundles of flakes, or most commonly as irregularly matted aggregates of moderately birefringent (white to straw-yellow) grains of clay size or larger.

Hornblende, a hydrous calcium, sodium, magnesium, iron, and aluminum silicate, is a common mineral in granodiorite, quartz diorite (fig. 8), amphibolite, and in some lime-silicate rocks. This mineral ranges in amount from 1 to 48 percent, and is slightly more stable than plagioclase in the altered rock zones since plagioclase is completely altered to clay minerals and sericite before all hornblende is altered. Hornblende is incipiently altered in zone 2 (fig. 9*A*), but in zones 2–3 and 3 it is completely replaced by clay minerals and mica. The clay minerals form along fractures and cleavage surfaces (fig. 9*B*) and eventually entirely replace the original mineral. The earliest alteration product is a low index, low birefringent, fibrous-appearing mineral aggregate resembling chlorite mixed with calcite(?). Where well developed, these alteration products give way to a coarser, higher birefringent clay mineral, probably illite, and finally to sericite. In some rocks, hornblende appears to be converted to a greenish, biotitelike mineral, later to a chlorite mica.

Garnet, which constitutes up to 20 percent of the garnet-quartz gneiss layers in the Fall River area, commonly is a sparsely disseminated mineral elsewhere in metasedimentary gneiss layers, and is moderately stable in altered rocks. The garnets are red to red-brown and, according to Harrison and Wells (1956), fall generally into the almandite type. The garnet in a garnet-quartz gneiss from the Golconda mine contained 7.5 percent Mn (quantitative spectrographic value), and 0.7 percent Mg, 3.0 percent Ca, 0.015 percent Ti, and Si, Al, and Fe each in amount greater than 10 percent[2] (Sims, written communication, 1960). The stability of garnet in the alteration environment is not yet fully understood. In zone 3, garnet is altered along cracks to a fine-grained calcite and clay-mineral assemblage, but it persists into zone 4 where the alteration of garnet is complete only

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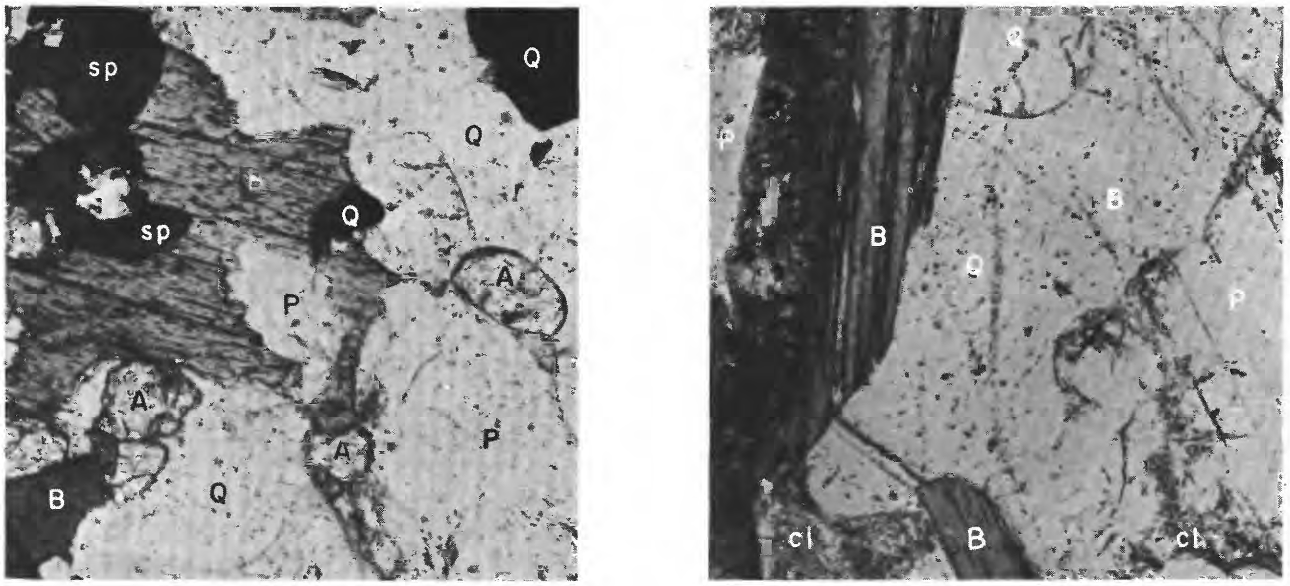
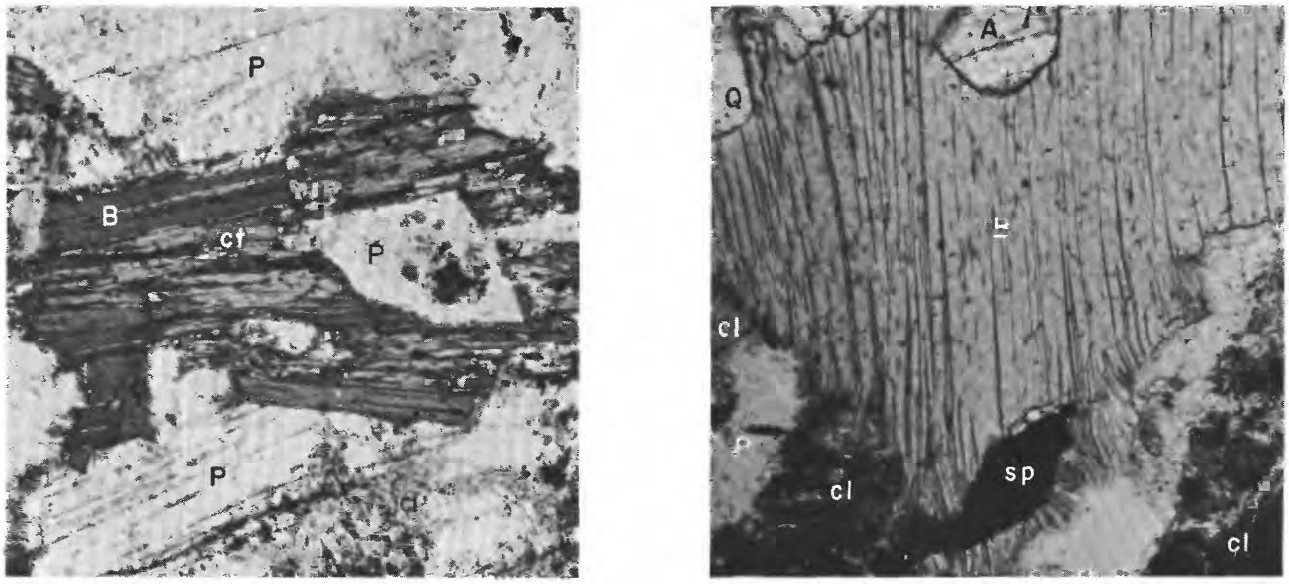
[2] Footnote, p. 50 indicates the significance of these semiquantitative spectrographic values.



A, Zone 1; granodiorite, Hayes and Wheeler mine; biotite(B), quartz(Q), plagioclase(P), sphene(sp), Apatite(A)-and magnetite-ilmenite (not shown) are unaltered; plane-polarized light, 60×.

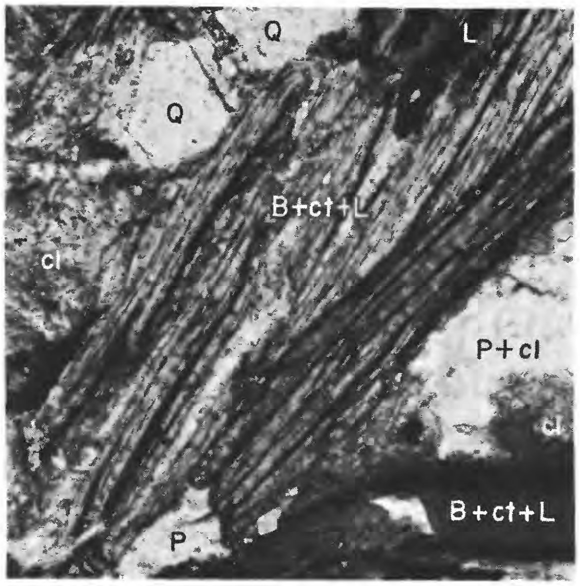
B, Zone 2; biotite-quartz-plagioclase gneiss, Essex mine; quartz(Q) is unaltered, plagioclase(P) and biotite(B) mostly are unaltered to clay minerals(cl) but are incipiently to well altered along a small fracture in the rock (across left side of photograph); plane-polarized light, 130×.



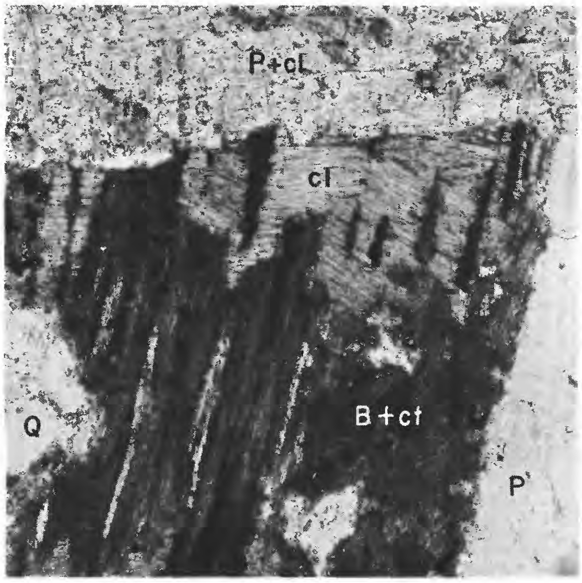
C, Zone 2–3; biotite-quartz-plagioclase gneiss, Essex mine; biotite flake(B) is partly altered to chlorite(ct), light gray, along cleavage; adjoining plagioclase crystals(P) are weakly altered to clay minerals(cl); plane-polarized light, 130×.

D, Zone 3; granodiorite, Hayes and Wheeler mine; biotite(B), quartz(Q) and apatite(A) are unaltered, plagioclase(P) is altered to brown-stained clay minerals(cl), sphene(sp) is partly altered to leucoxene; plane-polarized light, 130×.

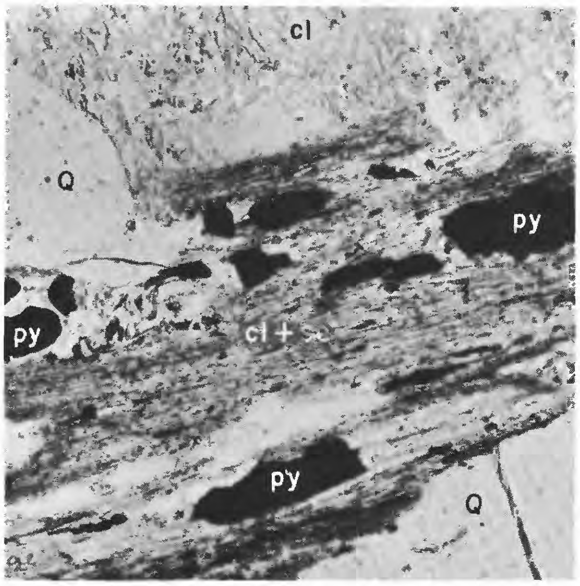
FIGURE 7.—Photomicrographs illustrating



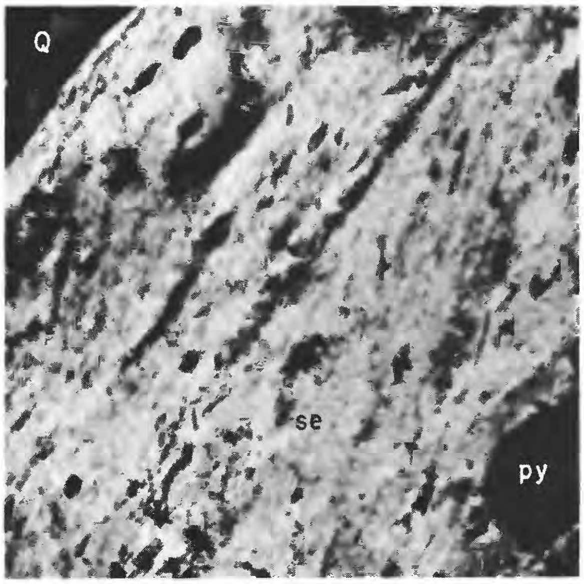
E, Zone 3; biotite schist, Almadin mine; quartz(Q) is unaltered, biotite(B) is altered along cleavages to chlorite(ct), clay minerals(cl) and iron oxide(L); most of the plagioclase(P) is strongly altered to a structureless mass of iron-stained clay minerals; plane-polarized light, 130×.



F, Zone 3; biotite-muscovite granite, Jo Reynolds mine; quartz(Q) is unaltered, biotite(B) is altered along cleavages initially to bleached biotite, chlorite(ct), and clay minerals(cl) and gradually, as the crystal is altered, a vermicular clay mineral(kaolinite?) growth is observed; plagioclase(P) is altered, but its crystal structure is not yet completely masked by clay minerals; plane-polarized light, 130×.

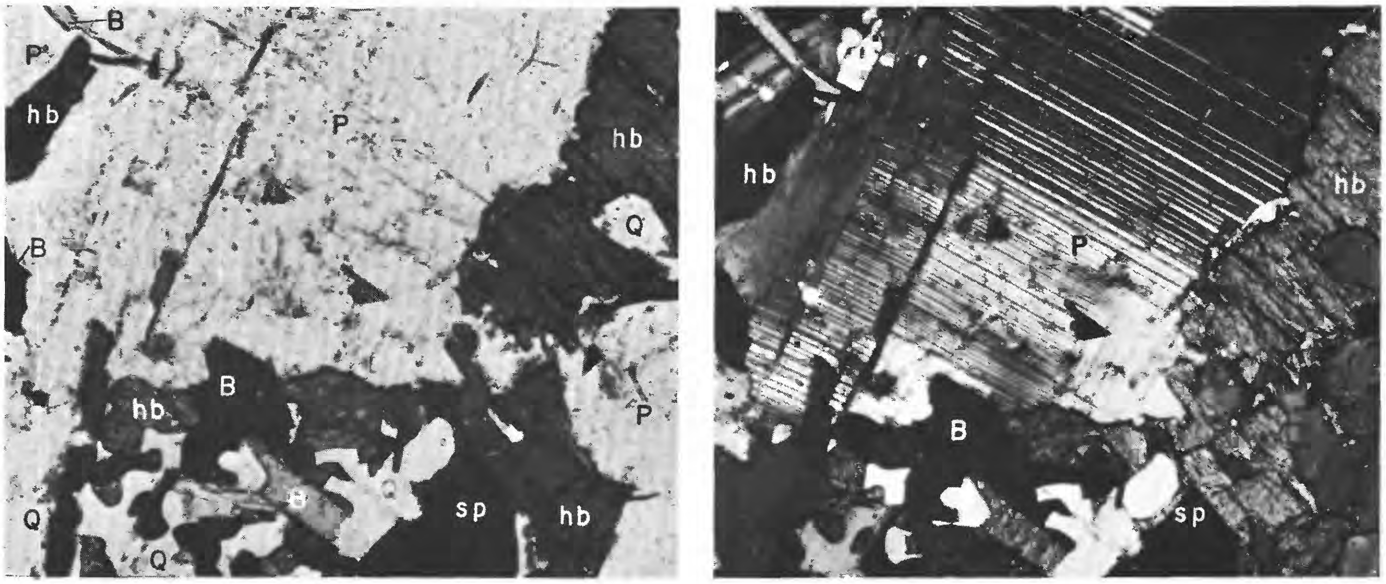


G, Zone 4; microcline-quartz-plagioclase-biotite gneiss, E. Calhoun mine; quartz(Q) is unaltered, argillized plagioclase and biotite crystals are almost completely altered to sericite(se) and pyrite(py); plane-polarized light, 130×.

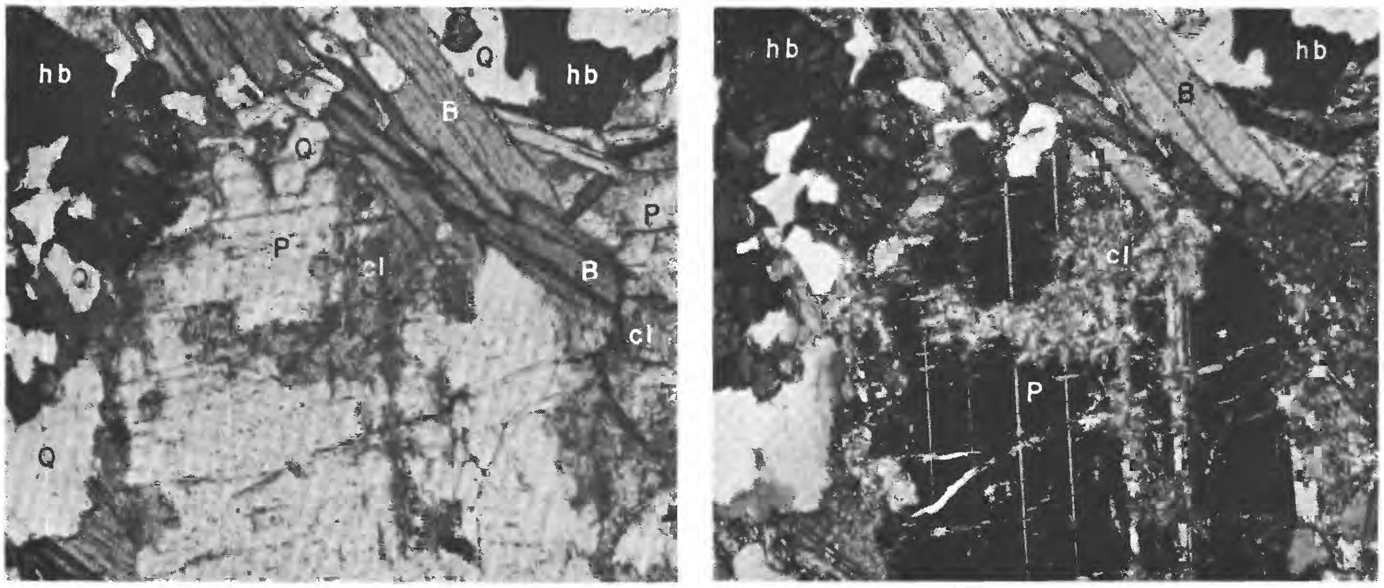


H, Zone 4; microcline-quartz-plagioclase-biotite gneiss, Essex mine; quartz(Q) is unaltered, biotite crystal is completely altered to sericite(se) and pyrite(py), crossed nicols, 130×.

the manner of alteration of biotite.

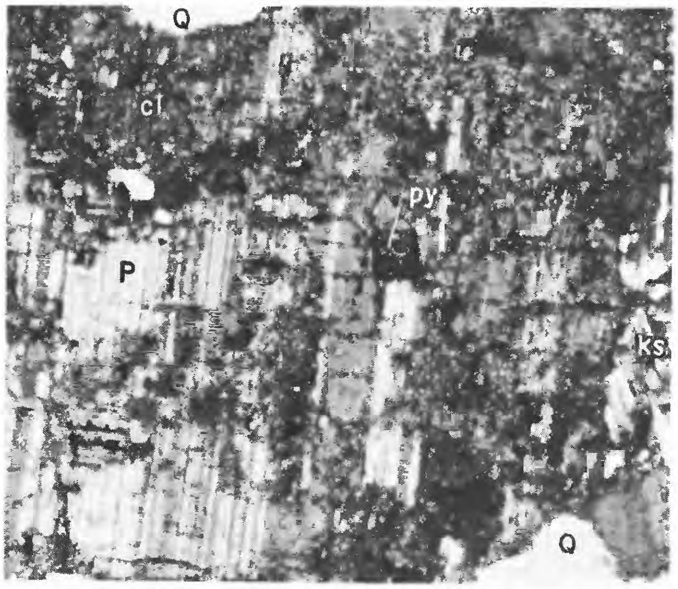


A, *B*, Zone 1; quartz diorite, Jo Reynolds mine; quartz(Q), plagioclase(P), hornblende(hb), biotite(B), sphene(sp) are unaltered; plane-polarized light and crossed nicols, 60×.

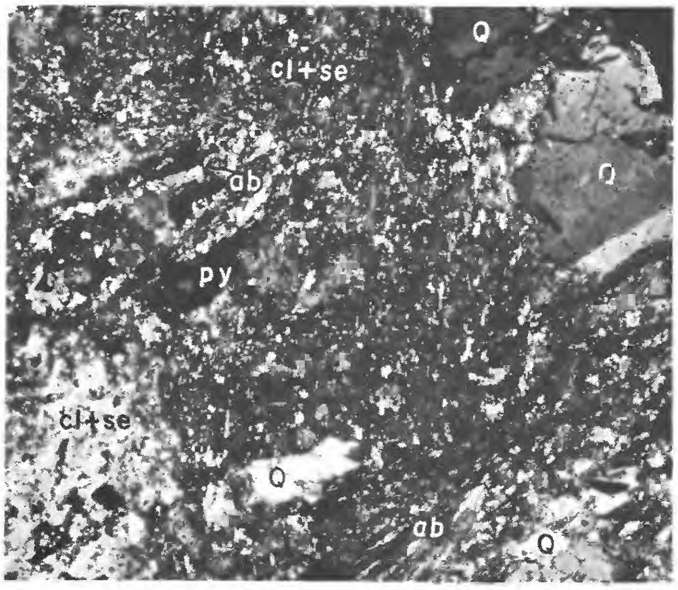


C, *D*, Zone 2; quartz diorite, Nabob mine; quartz(Q) and biotite(B) are unaltered, plagioclase(P) and hornblende(hb) are altered to clay minerals(cl) along cleavage planes and in irregular patches; plane-polarized light and crossed nicols, 60×.

FIGURE 8.—Photomicrographs illustrating



E, Zone 2; microcline-quartz-plagioclase-biotite gneiss, Essex mine; microcline(ks), quartz(Q), and pyrite(py) (lost in sectioning, only crystal outline shows) are unaltered, plagioclase(P) is weakly altered into patches of clay mineral(cl) irregularly across the crystal crossed nicols, 130×.



F, Zone 4; granodiorite, Hayes and Wheeler mine; quartz(Q) is unaltered, plagioclase(P) and biotite(B) are altered to clay minerals(cl), sericite(se); altered biotite grain at (ab).

the alteration of plagioclase in wallrocks.

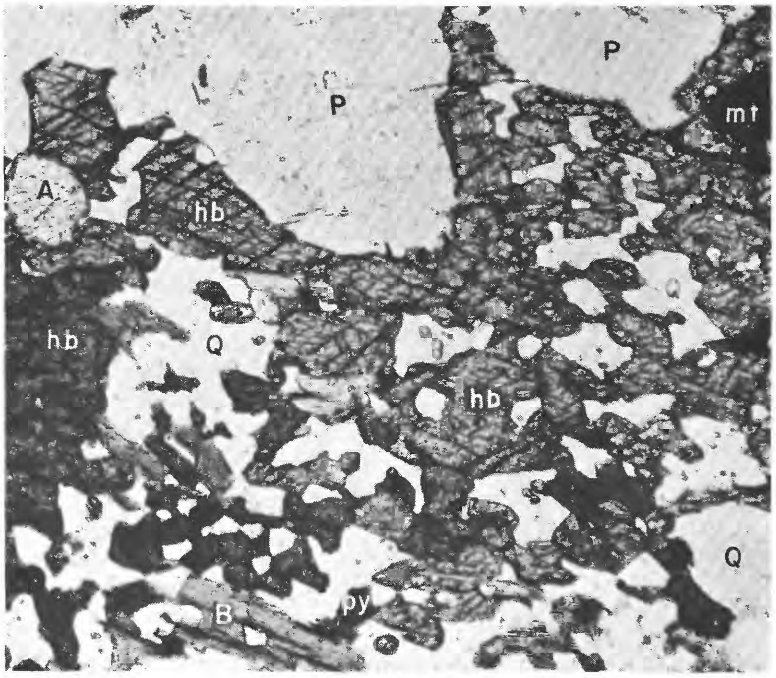
where biotite has been completely altered. Fractures and crystal margins of garnets are the loci of early alteration (fig. 10); as the rim of alteration expands, residual garnet islands are formed, and finally even the islands are overwhelmed, leaving recognizable textural pseudomorphs composed of clay, secondary biotite, sericite, and calcite.

Sphene, $CaTiSiO_5$, is widespread as a subhedral to wedge-shaped accessory mineral in many rocks but is especially conspicuous in the granodiorite in the Central City district (fig. 7). It is not stable in the hydrothermal environment, and begins to be altered in zone 2 (complete in zone 3) to the amorphous white material, leucoxene, which is an opaque whitish aggregate alteration product of titanium-bearing minerals (Allen, 1956, p. 832). However, the leucoxene is dissipated or indistinguishable in zone 4.

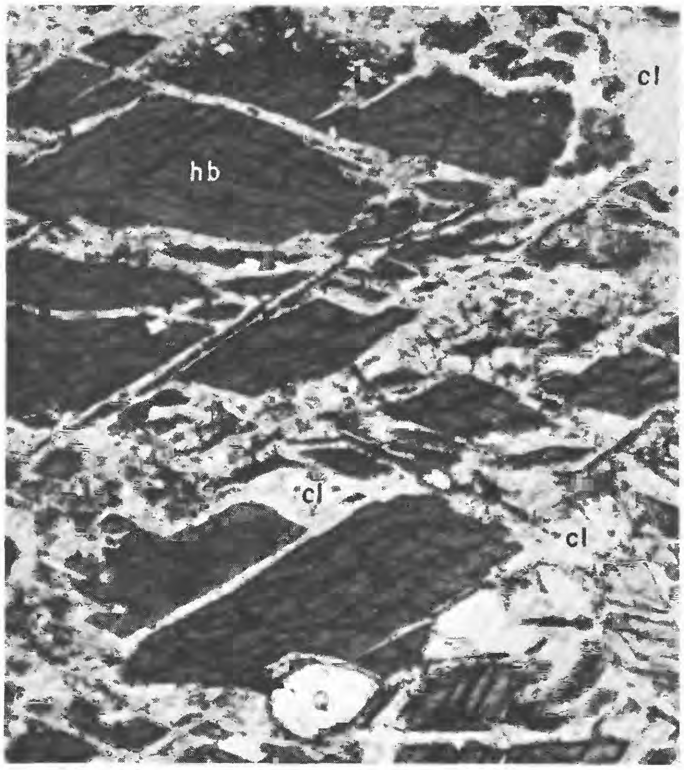
In the presence of acid supergene waters, accessory pyrite, FeS_2, oxidizes to limonite, $Fe_2O_3{\cdot}nH_2O$, which occurs generally as anhedral amorphous-appearing yellow masses and as a stain on host rock minerals in and near the veins.

Calcite, $CaCO_3$, is most common in fractures and replacing plagioclase and hornblende crystals close to veins. It is generally subhedral to anhedral, and medium to extremely fine grained. Calcite also is a common vein gangue mineral.

Clay minerals are the most common secondary minerals in altered rocks and may be subdivided into several species, each of which have distinct stability ranges. These mineral species are best characterized and identified by their distinctive lattice structures and are well described by Grim (1953, p. 43–83), and diagrammed in figures 11 and 12. Two basic structural units, tetrahedral and octahedral layers, are combined to form either the three-layer (mica-type, fig. 11) or the two-layer (kaolinite-type, fig. 12) clay minerals. The close similarity in mica-type structures allows for the common random interstratification on a unit-cell scale of chlorite, illite, and montmorillonite in mixed-layer clay minerals. Halloysite structure is that of a variably hydrated kaolinite with random displacements in the *a* and *b* crystallographic directions. Details of the polymorphism of three-layer clay minerals have not been determined.



A, Zone 2; quartz diorite, Jo Reynolds mine; quartz(Q), plagioclase(P), biotite(B), hornblende(hb), magnetite-ilmenite(mt), pyrite(py), and apatite(A) are unaltered; plane-polarized light, 130×.



B, Zones 2-3, amphibolite, R.H.D.-McKay shaft; quartz(Q) is unaltered, hornblende(hb) is altered to clay minerals(cl) along cleavages; plane-polarized light, 130×.

FIGURE 9.—Photomicrographs of fresh and altered hornblende.

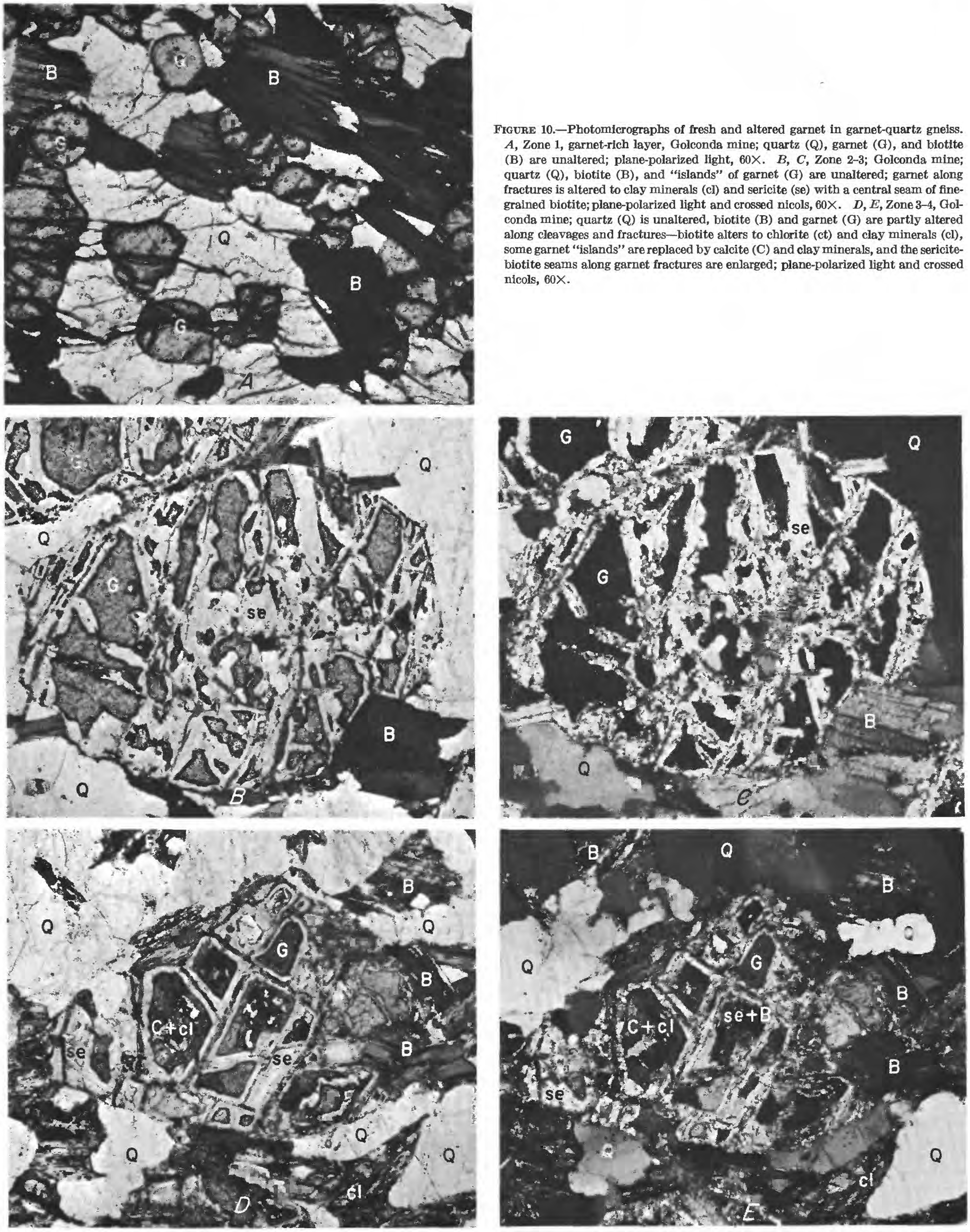


FIGURE 10.—Photomicrographs of fresh and altered garnet in garnet-quartz gneiss. *A*, Zone 1, garnet-rich layer, Golconda mine; quartz (Q), garnet (G), and biotite (B) are unaltered; plane-polarized light, 60×. *B*, *C*, Zone 2–3; Golconda mine; quartz (Q), biotite (B), and "islands" of garnet (G) are unaltered; garnet along fractures is altered to clay minerals (cl) and sericite (se) with a central seam of fine-grained biotite; plane-polarized light and crossed nicols, 60×. *D*, *E*, Zone 3–4, Golconda mine; quartz (Q) is unaltered, biotite (B) and garnet (G) are partly altered along cleavages and fractures—biotite alters to chlorite (ct) and clay minerals (cl), some garnet "islands" are replaced by calcite (C) and clay minerals, and the sericite-biotite seams along garnet fractures are enlarged; plane-polarized light and crossed nicols, 60×.

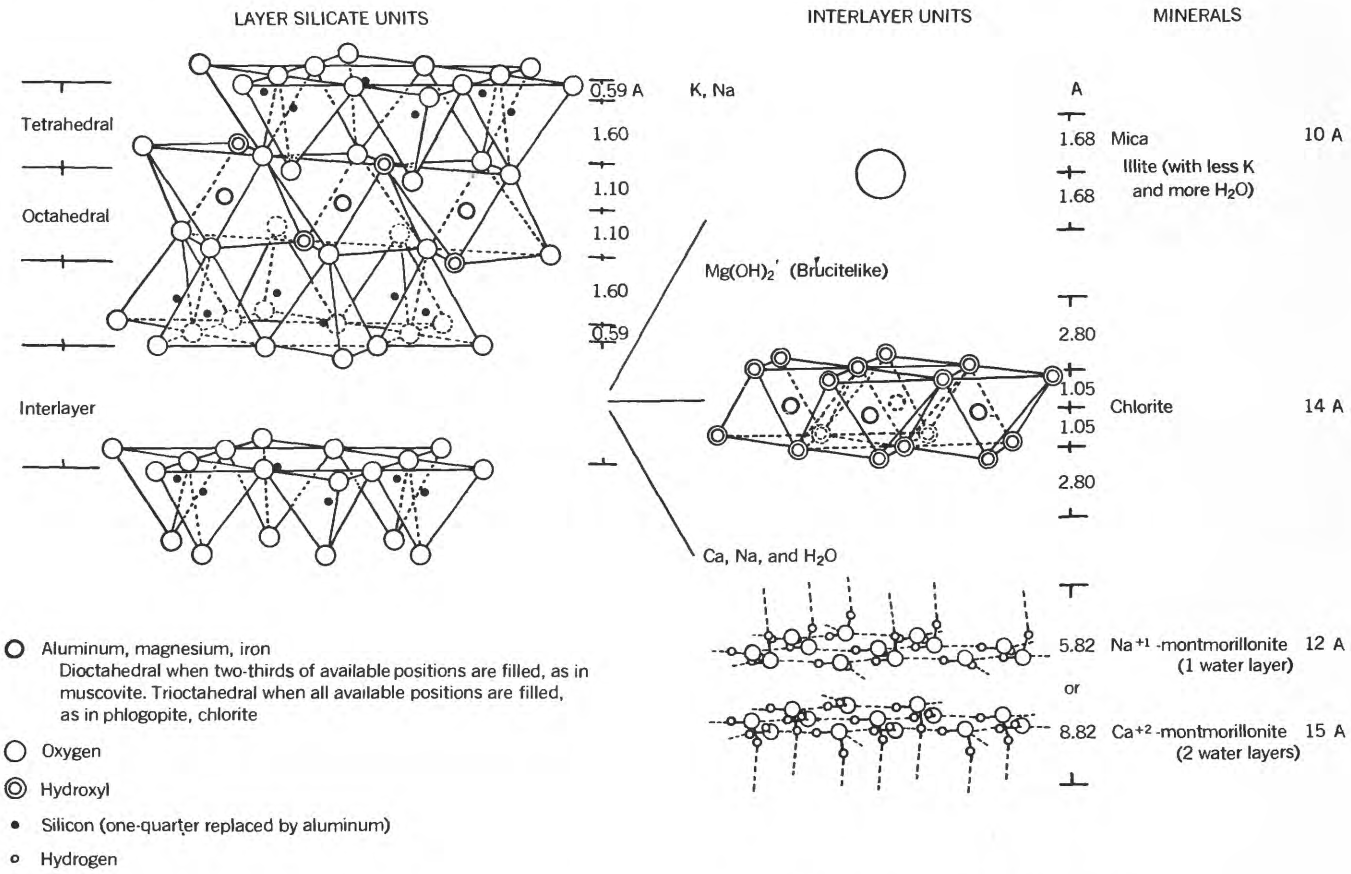


FIGURE 11.—Diagrammatic sketch showing the structure of some three-layer clay minerals (after Grim, 1953).

The indices of refraction characteristic of these clay materials obtained from oriented aggregates of clay minerals are shown in table 6. The particle size of the clay mineral is below the effective resolution size for the microscope; hence the image viewed usually is that of an aggregate of clay minerals. Grim (1953, p. 294) notes, therefore, that, "* * * the identification of clay minerals in thin section is usually difficult and often impossible * * *," although he feels (p. 293) that, "The petrographic microscope is * * * useful in studying the texture of clays * * *." The X-ray determinations, therefore, are combined with optical studies to form the basis for the following general

TABLE 6.—*Mean indices of refraction for oriented mixed-clay mineral and sericite assemblages in altered rock zones*

Clay mineral [1]	Index of refraction [2]	Altered rock zone	Host rock	Sample
Montmorillonite	1.542	2	Microcline-quartz-plagioclase-biotite gneiss.	22
Do	1.552	2	Granodiorite	125
Do	1.554	2	do	41
Do	1.520	3	do	43
Do	1.536	3	do	44
Do	1.550	2	Quartz-diorite	80a
Do	1.544	3	do	79a
Illite	1.552	3	do	78
Do	1.548	3	Microcline-quartz-plagioclase-biotite gneiss.	58
Kaolinite	1.546	3	do	21
Do	1.576	3	do	261
Do	1.544	3	do	242
Sericite	1.558	G		79
Do	1.566	V	Microcline-quartz-plagioclase-biotite gneiss.	51
Do	1.545	4	Biotite-quartz-plagioclase gneiss.	23

[1] Indicates most conspicuous clay mineral constituent in the aggregate.
[2] Mean value for α and γ indices observed normal to 001 face.

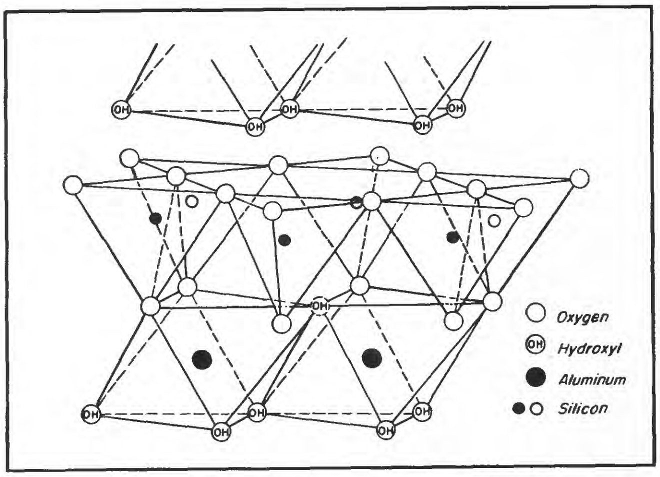


FIGURE 12.—Diagrammatic sketch of the two-layer structure of kaolite (after Gruner, 1932).

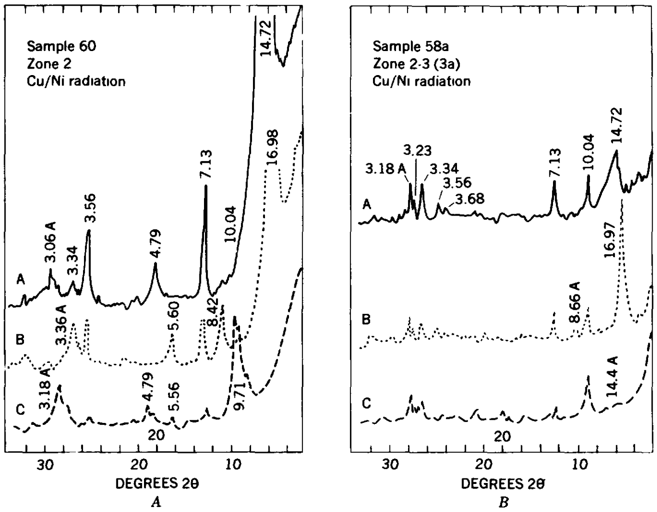


FIGURE 13.—Smoothed X-ray diffraction traces of oriented montmorillonite-rich clay mineral assemblages (typical of zones 2 and 3a) that occur in altered microcline-quartz-plagioclase-biotite gneiss, Essex mine: montmorillonite, kaolinite, mixed-layer clay, illite, and chlorite are present. *A*, untreated; *B*, ethylene-glycol treated; *C*, heated to 500° C.

discussions of the stability of clay minerals. The methods of X-ray identification and estimation of relative amounts of clay minerals used in this study are described by Weaver (1958).

The typical montmorillonite of the alteration zones bordering the veins in the central part of the Front Range mineral belt is indicated by X-ray studies to be chiefly of the calcium type (Ca as prime interlayer ion), which when air dried has interlayer water two molecular layers thick (fig. 11). The mineral is most common in zones 1, 2, and 3a, but it is found in small amounts in zone 4 as well. Weakly to moderately altered rock is characterized by the replacement of plagioclase crystals by a cream-white to greenish clay mineral that is primarily montmorillonite. Figure 13 shows the typical X-ray diffraction patterns of a montmorillonite-rich clay assemblage in zones 2 and 2–3. The X-ray patterns are partly smoothed by reduction in scale causing the deletion of very minor extraneous instrumental responses. Curve *A* is a trace of a separated but otherwise untreated clay-mineral assemblage. Curve *B* is the X-ray diffraction trace of the same sample after ethylene-glycol treatment, which expands the unit-cell *c*-dimension component and helps to resolve the identity of illitic and montmorillonitic components. Curve *C* is a trace of the same sample subsequently X-rayed after heat treatment at 500° C.

Incipient formation of clay minerals on cleavage surfaces of plagioclase in least altered rock (zone 1) is not readily observable in hand specimen. In thin section the twinned plagioclase crystals are mottled with specks and small irregularly shaped indistinct mineral masses. The index of refraction of large aggregates in thin sections is less than that of the plagioclase; birefringence is very low; and color in plane and polarized light is gray. The most abundant clay mineral is montmorillonite and it is formless, but the aggregates on most crystals are alined with the cleavage or twin planes. When observed on flat twin surfaces, the clay appears randomly distributed and exhibits no discernible crystallographic orientation.

In zone 2, the aggregates of montmorillonite-rich clay materials are enlarged and at some places completely cover the crystal surface; however, the plagioclase twinning, through obscured, is still visible. As the clay becomes more prominent it forms darker brown translucent masses that coalesce and broaden along cracks and along cleavage planes in the crystal. The earliest formed clay minerals appear to be replaced by a clay having a slightly higher index of refraction (> plagioclase) and higher birefringence that is possibly partly illitic in composition. Individual clay flakes are larger near the boundary with zone 3, and the altered crystals are cut by ramifying veinlets of clay.

In zone 3 the clay veinlets widen, and the entire crystal is replaced. Not all of the early, low-index material is completely changed over, and "island" remnants remain to give the clay pseudomorphs a mottled appearance.

In zone 4 most of the clay minerals are lost, having been replaced by sericite.

Chlorite, $(Mg,Fe,Al)_6(OH)_8(Si,Al)_4O_{10}$, is not a common secondary clay mineral; however, it was observed in thin section as an alteration product of hornblende in zones 1 and 2, and in part, of biotite in zones 3 and 4. Chlorite occurs about edges and along cleavages of these minerals (fig. 7), and it is recognized by the light pleochroic green-yellow color, low index of refraction, and low birefringence. In intensely altered rocks, chlorite is in turn replaced by the micalike clay minerals—montmorillonite, illite, and mixed-layer clays—and ultimately in zone 4 by sericite. The X-ray trace of a sample of chlorite heated to 500° C is shown in figure 13, curve *C*, sample 58a. The 7A peak persists at a lower intensity, and the 14A peak emerges, following the collapse of montmorillonite 001 spacing to that of the dehydrated phase close to 10A. In general, chlorite is not an abundant product of argillic alteration in these rocks, and commonly is obscured by the more prominent clay phases.

Illite, general composition $K_{1.33}(OH)_4\ (R_4^{+3},M_6^{+2})Si_{6.67}Al_{1.33}O_{20}$ where $R=Al^{+3}$, Fe^{+3}, or Ti^{+3}, and $M=Mg^{+2}$, Fe^{+2}, or Li^{+1} (Yoder and Eugster, 1955, p. 252), is the mica-type clay mineral with nonexpandable lattice. The optical properties for illite are similar to those for montmorillonite, and illite is texturally indistinguish-

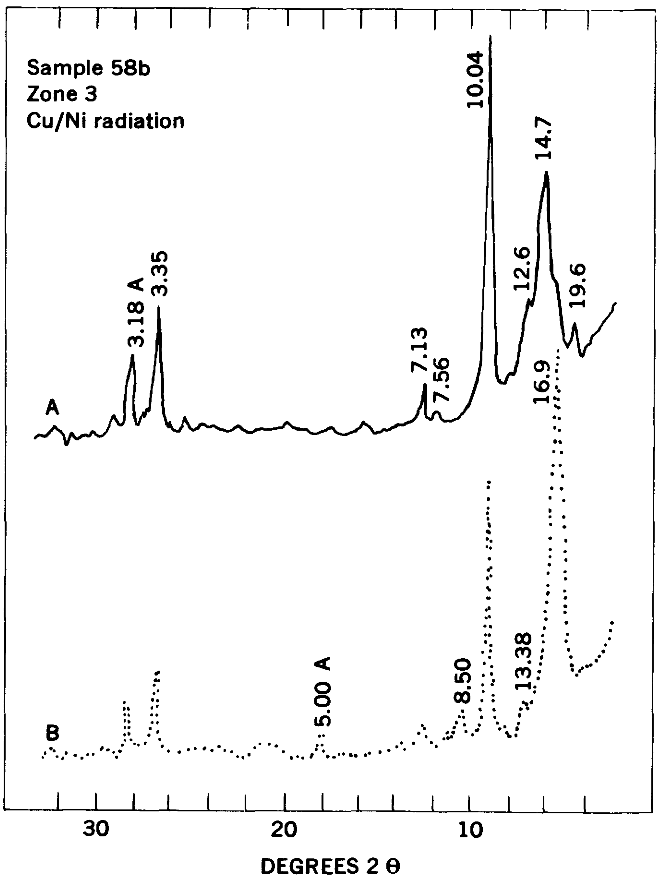


FIGURE 14.—Smooth X-ray diffraction traces of an illite-rich clay mineral assemblage (typical of zone 3b) that occur in altered microcline-quartz-plagioclase-biotite gneiss, Essex mine: illite, montmorillonite, mixed-layer clay and kaolinite are present. *A*, untreated; *B*, ethylene-glycol treated.

able in thin section from other similar clay minerals. X-ray studies, especially of material treated with ethylene glycol or heated prior to study, clearly show the unexpanded 10 A basal order sequence and its predominance in zone 3 (fig. 14); illite occurs throughout the zones of altered rock in less abundance. Yoder and Eugster (1955) describe material that previously was believed to be illite actually to be related instead to 1Md or 1M mica polymorph crystallization or to mixed-layer clay mineral type. It was not possible to determine polymorphic form of mica in these analyses, thus the name illite was retained to describe conveniently clay material whose nonexpanding X-ray spacings are more diffuse than those for sericite or mica, and whose composition may be intermediate between montmorillonite or chlorite and biotite.

Mixed-layer clay minerals result from an interstratification of layered clay minerals on the order of a single or a few sheets. Mixed-layer structures thus formed are as stable as those composed of symmetrically layered alumino-silicate sheets because of the close similarity between the structures of individual clay minerals (fig. 11). The composition of the mixed-layer clay is of course regulated by the composition of the constituent mixtures of clay minerals. Mixing, discussed in detail by Weaver (1956), is of two types: (a) regular interstratification along the *c* axis resulting in a unit cell that is equal to the sum of component layers, allowing regular 001 reflections; or (b) random interstratification or irregular mixing, which results in a nonintegral sequence of 001 reflections.

Mixed-layer clays are common in altered wallrocks in zones 1, 2, and 3 but are identified only by X-ray studies. Since mixed-layer clays were not specifically identified optically, their textural importance cannot be discerned. Both irregular and regular mixed-layer clays occur in altered wallrocks in zones 2 and 3 but the randomly interstratified clays are most common; however, mixed-layer clays are absent or in very small amounts in zone 4 rock. The precise determination of the mixed-layer type is difficult in the rapid X-ray scan used, but in figure 15 examples of mixed layering are indicated. The montmorillonite and illite reference basal reflections are shown under each trace and an asterisk indicates a change from lower to regular sensitivity for subsequent peaks.

Kaolinite, $(OH)_8Al_4Si_4O_{10}$, is most abundant in zone 3c as a sparse alteration product of K-feldspar or biotite, and more commonly as a replacement for earlier formed clay minerals. It occurs also in zones 1 through 4 in minor amounts. Zone 3 clays are generally white and pseudomorphically replace host-rock feldspars. Kaolinite is inferred in thin section by the presence of low index and low birefringent clay replacing montmorillonite and possibly biotite in the boundary region between zones 3 and 4. Measurements of indices of refraction of oriented clay-mineral aggregates rich in kaolinite from zone 3c are shown in table 6. The index values, slightly higher than those of Grim (1953), reflect the intermixing of clay minerals (fig. 16) common to this type of replacement. The sample shown in curve *c*, figure 16, was heat treated at 300°C. Kaolinite structure persists. Treatment at 500°C destroys the kaolinite structure, but not that of chlorite, which is a phase commonly obscured by the kaolinite X-ray pattern.

Halloysite, $(OH)_8Al_4Si_4O_{10}$, with basal spacing close to 7.5 A occurs sparsely in zone 3c and more abundantly in zone 4 of some rocks (in altered garnet-quartz gneiss). It is closely related to kaolinite in crystal structure as well as in mineral texture and appearance in thin section. The X-ray data indicate that the mineral was partly hydrated when examined (Grim, 1953, p. 54).

Sericite, according to Winchell and Winchell (1951, p. 369), "is a fine scaly or fibrous kind of muscovite;

the name is usually confined to white mica which is secondary." Yoder and Eugster (1955, p. 255) consider sericite to be a fine-grained white mica, whereas Wright and Shulhof (1957) class sericite as a clay mineral. In this report sericite is defined as a secondary white mica within the clay-mineral particle size range which yields a well-crystallized mica X-ray pattern. X-ray diffraction data for a typical sample of sericite (table 7) indicate that it is a dioctahedral, 2M mica.

Sericite is most common in zone 4 rock, in vein gouge clay, and in altered brecciated wallrock included in the vein. A transitional boundary between zone 3 illite and zone 4 sericite is difficult to delineate at

Table 7.—*X-ray diffraction powder data for sericite (—2μ) in zone 4, Essex mine*

[Indices for mica from Yoder and Eugster (1955, p. 247) correspond to similar d values· M, montmorillonite, K, kaolinite impurities, Copper/nickel radiation: s, strong; m, moderate; w, weak relative intensities; vs, very strong; b, broadline]

d angstroms	Index	Mica hkl	d angstroms	Index	Mica hkl
9.96	vs	002	2.80	vw	$11\bar{6}$
7.13	vvw	K	2.68	vw (b)	
4.98	ms	004	2.57	m	$13\bar{1}$, M
4.57	mw	$11\bar{1}$	2.49	w	008
4.37	vw	021	2.47	w	$13\bar{3}$(?)
4.27	w	111	2.39	w	133(?), $20\bar{4}$, K
4.13	vw	022	2.15	vw (b)	222
3.90	vw	$11\bar{3}$	2.14		043, 135
3.66	vw	023(?), K	1.99	ms	0010
3.32	vs	006, 024	1.82	vw	K, M
3.22	w (b)	114	1.66	w (b)	$20\bar{10}$
3.07	mw	M	1.65		314(?)
3.01	w	025	1.64		
2.94	vw		1.54	vw	M
2.86	vw	115	1.51	mw	060

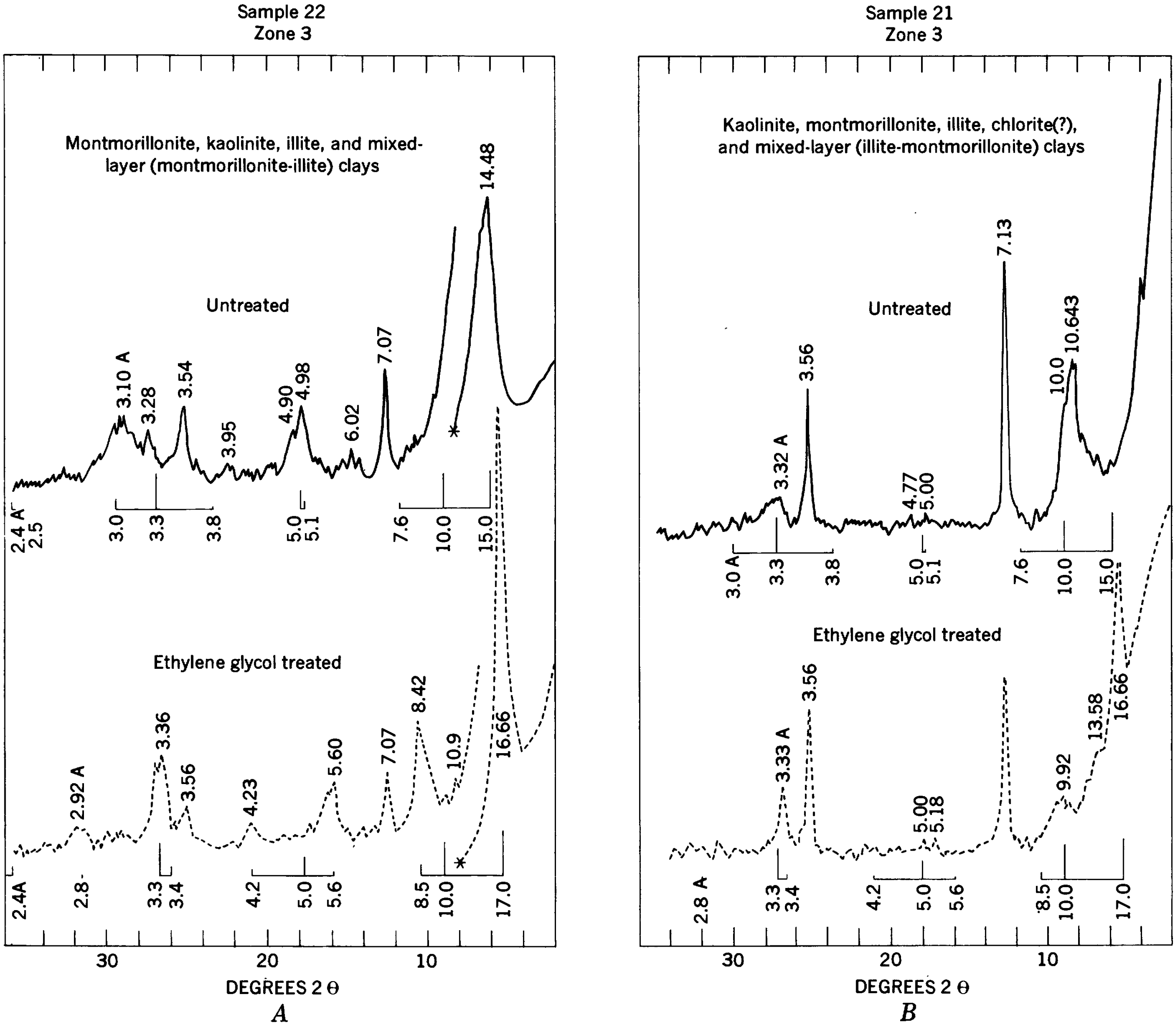


Figure 15.—Smoothed X-ray diffraction traces of clay-mineral assemblages, including mixed-layer clay, from altered biotite-quartz-plagioclase gneiss, Essex mine.

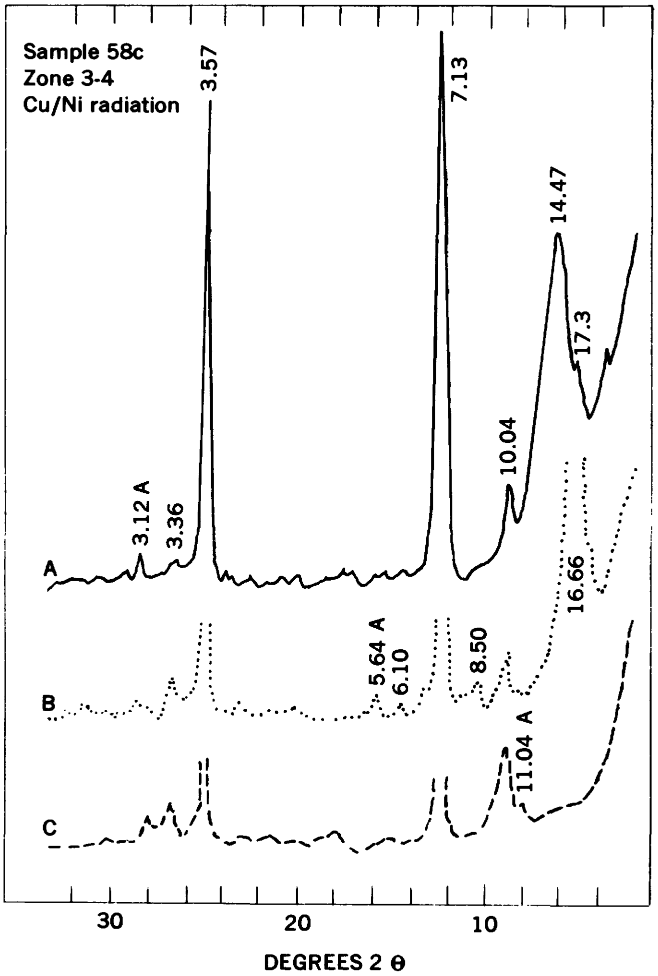


FIGURE 16.—Smoothed X-ray diffraction traces of a kaolinite-rich clay-mineral assemblage (typical of zone 3c) that occurs in altered microcline-quartz-plagioclase-biotite gneiss, Essex mine: kaolinite, montmorillonite, mixed-layer clay, and illite are present. *A*, untreated; *B*, ethylene-glycol treated; *C*, heated to 300° C.

most places. Sericite in zone 4 is in part a replacement of plagioclase, K-feldspar (perthitic variety), and biotite following their alteration first to clay minerals (fig. 17). Typical sericite textures are shown in figure 18, and result from randomly placed radial clusters, oriented elongated bunches of flakes, and random patches.

Handpicked feldspar grains from granodiorite in zone 1 that contained material identified microscopically as sericite was crushed, the clay-sized fraction separated, and X-rayed. This material (fig. 19) consists of a mixture of illite, montmorillonite, chlorite, and random mixed-layer clay. X-ray sensitivity was about two times greater than normal to obtain these traces.

EXPERIMENTAL STUDIES OF MINERAL STABILITIES

Recent experimental synthesis data may be useful as an aid to interpret geologic data, and several reviews of the literature by Ingerson (1955a, 1955b), Yoder (1955), Lovering (1955), and Kerr (1955) present a wide selection of such data. However, many of these data are of limited value for geologic interpretation because mineral stabilities are considered in terms of bulk compositions of solids and as simple hydration-dehydration reactions (such as Roy, 1954) rather than as they occur in nature as assemblages in equilibrium with a single or succession of solutions. J. J. Hemley (written communication, 1960) points out that in the alteration process it is the solutions that do the work; these solutions move in an open system thus approaching equilibrium in a succession of local states over a range of chemical conditions. His experimental work (1959) shows that the K^+/H^+ activity ratio and temperature are the most important controls on the fields of stability in the K_2O-Al_2O_3-SiO_2-H_2O system. At a given temperature and with increasing K^+/H^+ activity ratio, the fields of kaolinite, mica, and K-feldspar are suc-

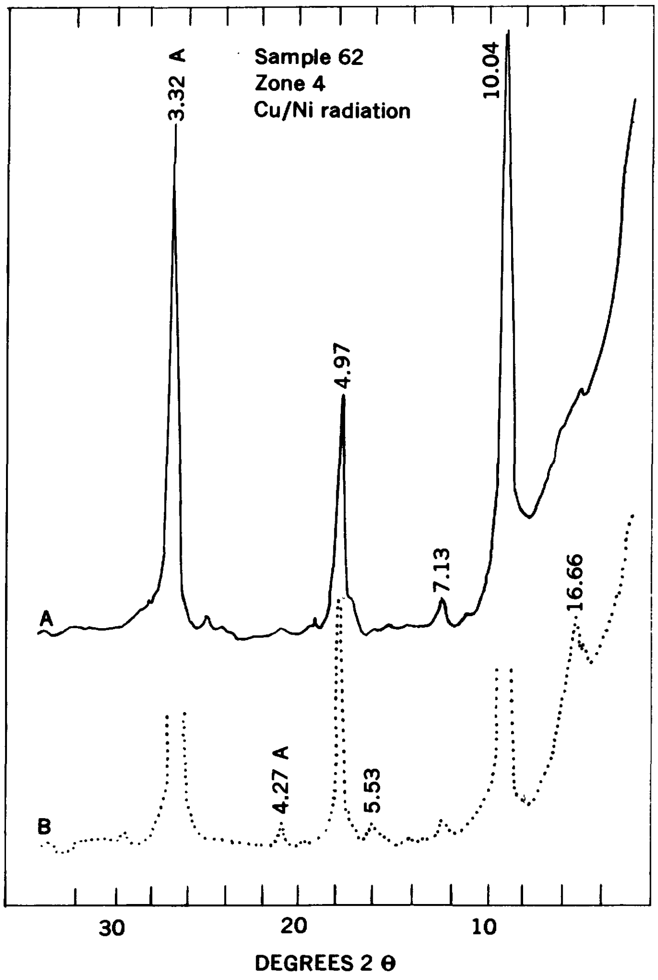
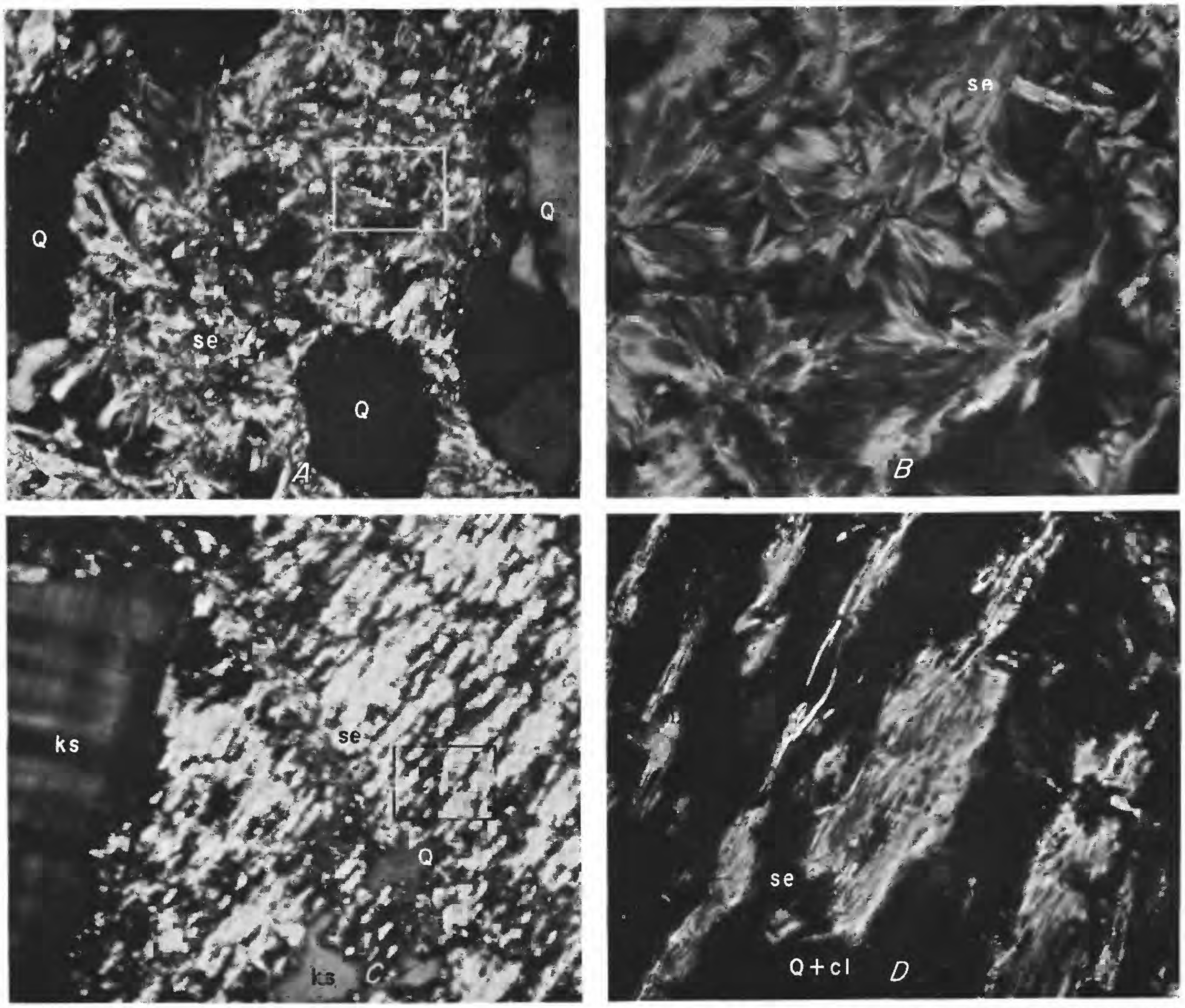


FIGURE 17.—Smoothed X-ray diffraction traces of sericite in most intensely altered microcline-quartz-plagioclase-biotite gneiss, Essex mine: montmorillonite and kaolinite are trace constituents. *A*, untreated; *B*, ethylene-glycol treated.



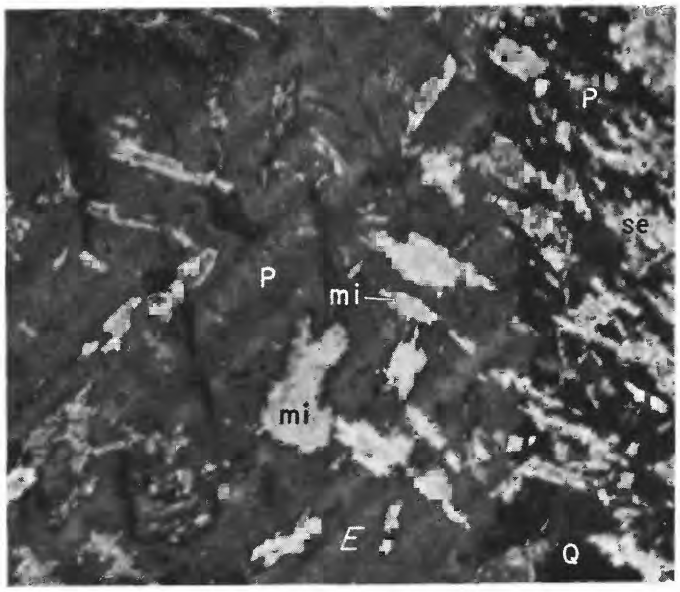


FIGURE 18.—Photomicrographs illustrating the textures of sericite. *A*, Zone 4; microcline-quartz-plagioclase-biotite gneiss, Alma-Lincoln mine; quartz (Q) is unaltered, sericite (se) replaces plagioclase; area within rectangle is enlarged in *B*; crossed nicols, 130X. *B*, Enlargement of rectangle area of *A*, showing radial clusters of bladed sericite crystals in detail; crossed nicols, 320X. *C*, Zone 4, microcline-quartz-plagioclase-biotite gneiss, Essex mine; microcline (ks) and quartz (Q) are unaltered, sericite (se) replaces plagioclase; area within rectangle is enlarged in *D*; crossed nicols, 130X. *D*, Enlargement of rectangle area of *C*, showing elongate parallel bundles of sericite blades separated by clay minerals and quartz; crossed nicols, 320X. *E*, Zone 3–4; biotite-muscovite granite, Jo Reynolds mine; quartz (Q) is unaltered, the left side of the plagioclase crystal (P) is partly replaced by large, random, sparse sericite (se) or mica (mi) flakes, and the right side is almost completely altered to oriented bundles of sericite blades separated by clay minerals (Cl) and quartz(?); crossed nicols, 130×.

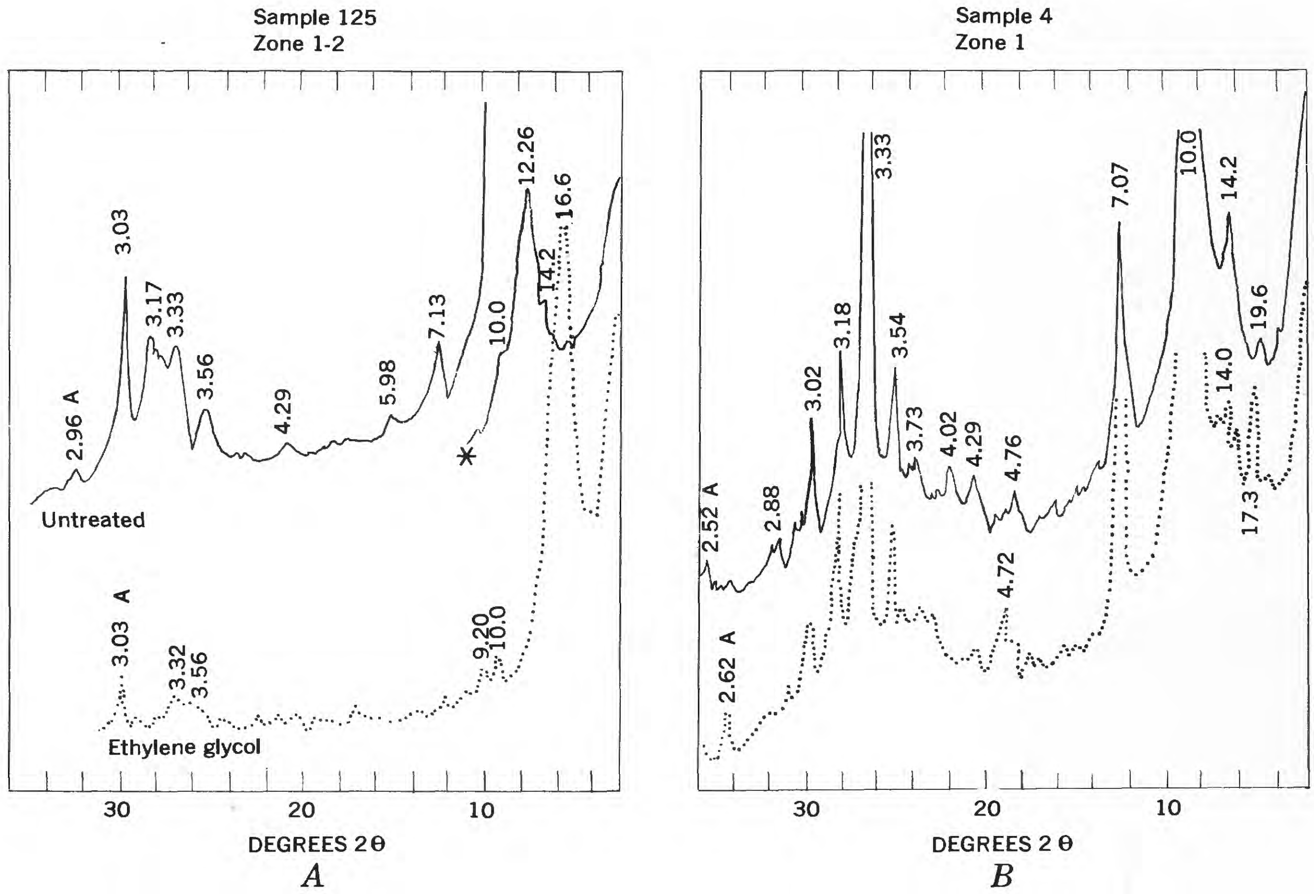


FIGURE 19.—Smoothed X-ray diffraction traces of clay minerals separated from fresh granodiorite by the ultrasonic separator: illite, montmorillonite, chlorite, and mixed-layer clays are present.

cessively traversed. With constant K^+/H^+ ratio and increasing temperature, the same sequence may be obtained.

MINERAL STRUCTURES AND THEIR INHERITANCE

Mineral structures and the possible inheritance of structures are of limited value to thermodynamic considerations of equilibrium chemistry, yet they are significant for the understanding of mineralogical-chemical relations in alteration processes. This aspect of the alteration process will be emphasized, but an attempt will be made to relate these data to the thermodynamic considerations of Hemley (1959). The basic factors governing the spatial and energy relations between ions in silicate minerals are well summarized by Hauth (1951) and Henderson (1951). Of fundamental importance in the following discussion is the observation of the large size of the O^{-2} ions compared with Ca^{+2}, Mg^{+2},Fe^{+3}, Al^{+3}, and Si^{+4} ions. Silicate crystal structures are best described as regular close packings of the larger oxygen ions with associated smaller cations in the interstices. Mason (1952, p. 219) points out that dense structures, such as those composed of single tetrahedra, are generally more economical of space than are framework structures having a more open lattice. This explains, in part, why large ions may be replaced by smaller ions during rock alteration without loss of volume.

The concept that parent mineral structures exert an influence over subsequent secondary mineral structures is confirmed in part by the textures of certain altered rocks seen in thin section, that is, the formation of clay minerals and sericite along certain crystal directions in altered feldspar (fig. 8) and mica (fig. 7) crystals. The persistence of K-feldspar in the kaolinite argillized zone demonstrates that the arrangement of ions in stable mineral structures may impede formation of the normally expected chemical equilibrium assemblages. Bradley and Grim (1951) have demonstrated the structural lineage in the thermal decomposition reaction of kaolinite to mullite. However, it has not been possible to demonstrate conclusively with laboratory methods that there is or is not structural continuity in the alteration of feldspar to clay mineral (Brindley and Radoslovitch, 1956).

ALTERATION OF REPRESENTATIVE WALLROCKS

The various types of igneous and metamorphic rocks in the region have altered differently, mainly because of differing mineral makeup. Six of the rocks[3] or groups of rock are considered representative of those in the area: (1) biotite-muscovite granite and microcline-quartz-plagioclase-biotite gneiss, (2) granodiorite and biotite-quartz-plagioclase gneiss, (3) quartz diorite, (4) amphibolite, (5) garnet-quartz gneiss, and (6) bostonite. In addition to the differences between rock types, there is variability within each rock type, as is shown in figure 20.

The rocks diagrammed in figure 20 may be classified into two general groups: (1) rocks in which K-feldspar is a major constituent, and (2) those in which it is a minor constituent or only an accessory. Biotite and plagioclase are constituents in both groups of rock, but hornblende is abundant only in the K-feldspar deficient rocks.

BIOTITE-MUSCOVITE GRANITE AND MICROCLINE-QUARTZ-PLAGIOCLASE-BIOTITE GNEISS

These K-feldspar-bearing rocks are the most abundant rock types observed; the granite and its altered phases were sampled along lead-zinc-silver veins in the peripheral ore zone in the Freeland-Lamartine and Lawson-Dumont-Fall River districts, whereas the microcline-rich gneiss was sampled along typical veins in the three ore zones and in all of the mining districts. The fresh biotite-muscovite granite is a medium-gray to pink generally medium grained rock composed of plagioclase, K-feldspar (microcline, and perthitic intergrowths in plagioclase), quartz, biotite, and muscovite; sphene, apatite, zircon, magnetite-ilmenite, and pyrite are present in minor amounts (table 8). Clay minerals, hematite, leucoxene, and chlorite are sparse secondary minerals that locally replace plagioclase, magnetite-ilmenite, and biotite in even the freshest rock.

Weakly argillized granite is hard and possesses a characteristic granitic texture. In thin section, plagioclase is seen to be converted partly to clay minerals found by X-ray diffraction studies to be montmorillonite and illite. Biotite is partly replaced by chlorite (fig. 7*c*), clay minerals (table 9), iron oxide, and locally by minor amounts of epidote. K-feldspar and muscovite are not visibly altered in zone 2. The stability ranges of the primary and secondary minerals of the biotite-muscovite granite and their relative abundance in this and subsequent zones are shown in figure 21.

TABLE 8.—*Modal analyses, in volume percent, of fresh and altered biotite-muscovite granite, Nabob and Jo Reynolds mines*

[Chemical and spectrographic analyses; samples 1, 2, and 3 are composites of 138 139, 140, 141a, and 141b]

Mineral	Zones							
	1–2		2	3	4			
Sample	141a	141b	138	140	154	139	136	143
Quartz	29.3	29.4	29.8	28.8	31.3	35.1	28.7	35.0
Plagioclase	31.3	6.8	5.1	3.9		15.5	2.5	3.2
K-feldspar	21.1	26.1	36.5	36.3	33.7	32.6	23.4	36.3
Biotite	7.6	9.8	3.9	2.2				
Muscovite	2.7	1.8	.7	.9	2.1	.3	7.0	4.2
Sphene	2.3	.3	.2				.2	.1
Apatite		1.2	.2	.4	.9	.6	.7	.3
Zircon		.1						
Hematite	.2	.1	.2	.7	.3	.9	.4	.1
Magnetite-ilmenite	.3		.1					
Pyrite			.1				Tr.	Tr.
Leucoxene	.6	.4	.6	.3			.4	
Clay minerals	2.4	[1] 22.2	[1] 21.5	[1] 25.2	30.4	12.8	35.6	19.9
Chlorite	2.2	1.4	.9	1.1			.7	
Calcite				.2	1.1	1.7		.8
Total	100.0	99.6	99.8	100.0	99.8	99.5	99.6	99.9
Points counted	1,013	1,025	1,140	1,122	1,038	1,033	928	1,090

[1] Mostly clay splotches in recognizable plagioclase grains.

Localities where samples were collected:
- 141a, Jo Reynolds mine, about 1½ feet into hanging wall.
- 141b, Jo Reynolds mine, adjacent to 141a, but closer to vein.
- 138, Jo Reynolds mine, crosscut away from main vein zone.
- 140, Jo Reynolds mine, crosscut away from main vein zone.
- 154, Nabob mine, rock in horse between vein splits.
- 139, Jo Reynolds, red alteration stain on K-feldspar.
- 136, Nabob mine, along main vein zone.
- 143, Jo Reynolds, silicified zone 1 foot thick on hanging wall.

Strongly argillized granite in zone 3 is a moderately hard rock with recognizable original rock textures in which clay minerals almost completely replace plagioclase and biotite crystals. The large amounts of unaltered quartz, muscovite, and K-feldspar, which remain in zone 3 (fig. 4*B*), preclude softening of the altered granite as is commonly observed in other strongly argillized rocks. Scattered pyrite and disseminated blebs of calcite occur sparsely throughout the rock.

TABLE 9.—*Clay minerals and sericite representative of altered-biotite-muscovite granite, Diamond Mountain and Nabob mines*

[Relative amounts of constituents: s, sparse; m, moderate; a, abundant; va, very abundant; vs, very sparse; tr, trace]

Altered-rock zone	Clay minerals					Sericite	Sample
	Montmorillonite	Mixed-layer[1]	Chlorite	Illite	Kaolinite		
V–4	s	s				va	156
G	s–m	m[2]	tr(?)	a	vs		151
4	vs	tr			vs	va	154
4	vs	tr				va	155
3	a	s	tr(?)	m	m–s		130a
2–3	m–a	m–s		m	a		131

[1] Mixed-layer clay minerals are predominantly of montmorillonite-illite type.
[2] Mixed-layer clay minerals of illite-montmorillonite type.

Localities where samples were collected:
- 156, Nabob mine, wallrock included in vein zone.
- 151, Nabob mine, footwall gouge clay.
- 154, Nabob mine, silicified rock adjacent to a subsidiary vein.
- 155, Nabob mine, greenish silicified zone adjacent to main vein zone.
- 130a, Diamond Mountain mine, rock from hanging wall.
- 131, Diamond Mountain mine, altered granite 6 in. into hanging wall.

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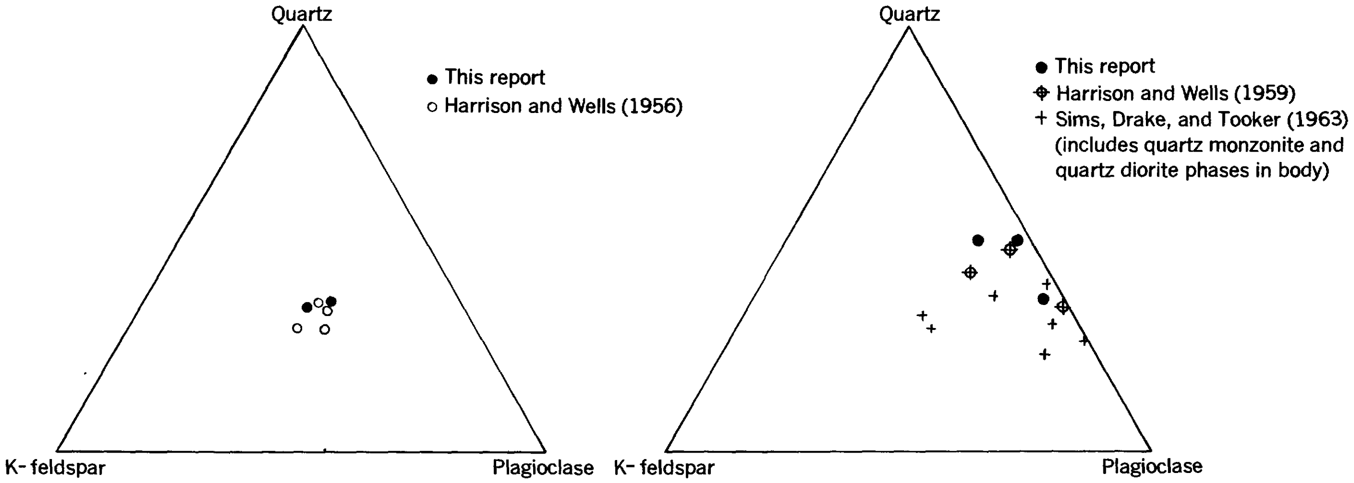
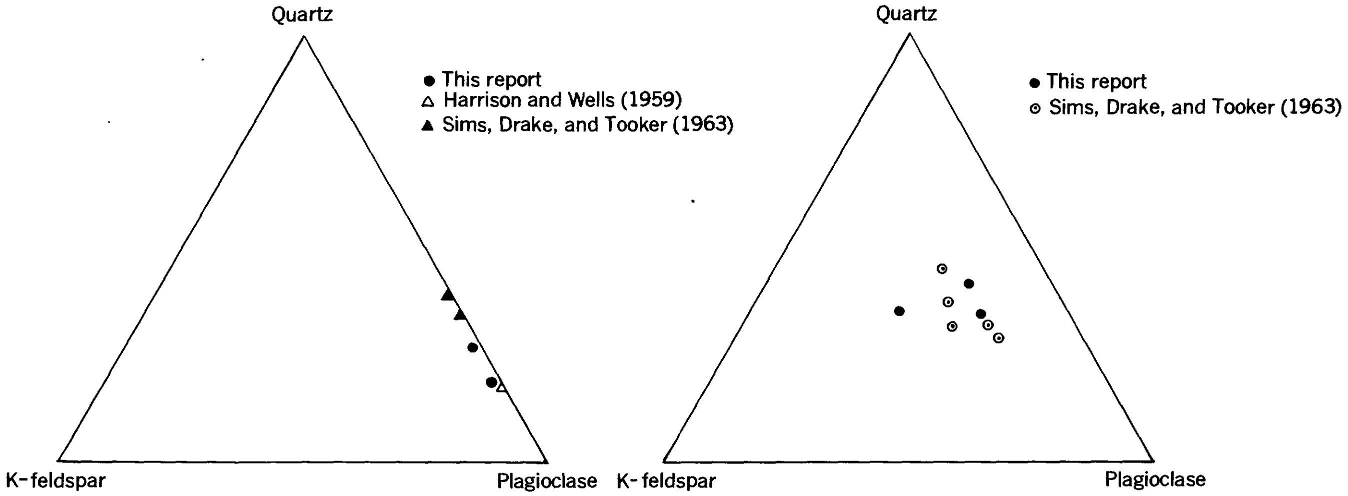
[3] The specific rock names used in this report conform to those selected by coworkers who mapped these units on the surface and underground, and whose reports this paper is meant to amplify. Therefore the names of rocks described in the text are based on average compositions of the rock determined elsewhere in the region by modal analyses of a large number of specimens taken from a multitude of rock outcrops.



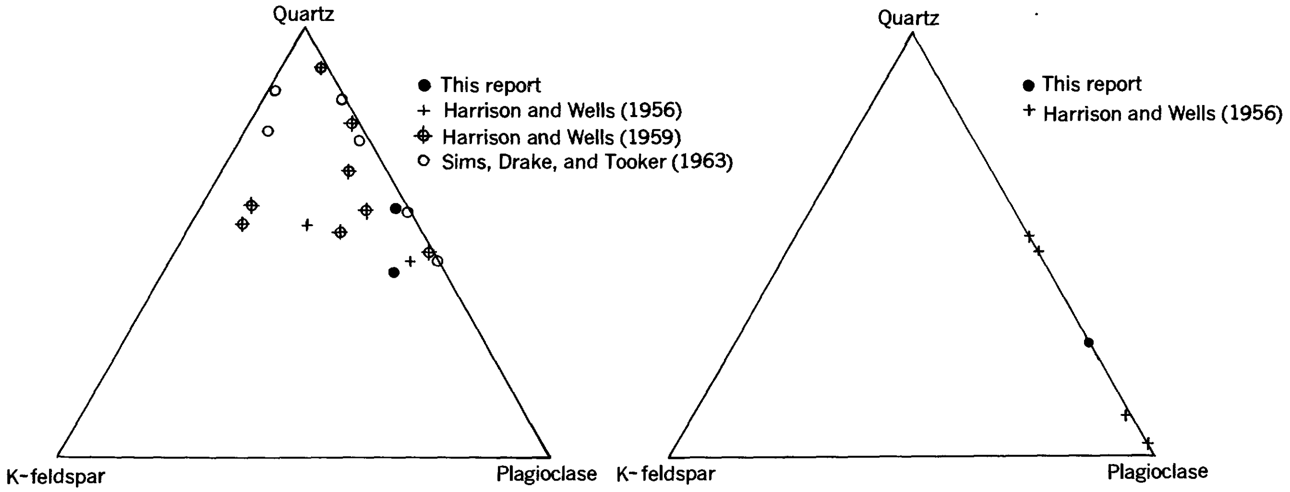
A. Biotite-muscovite granite

B. Granodiorite



C. Quartz diorite

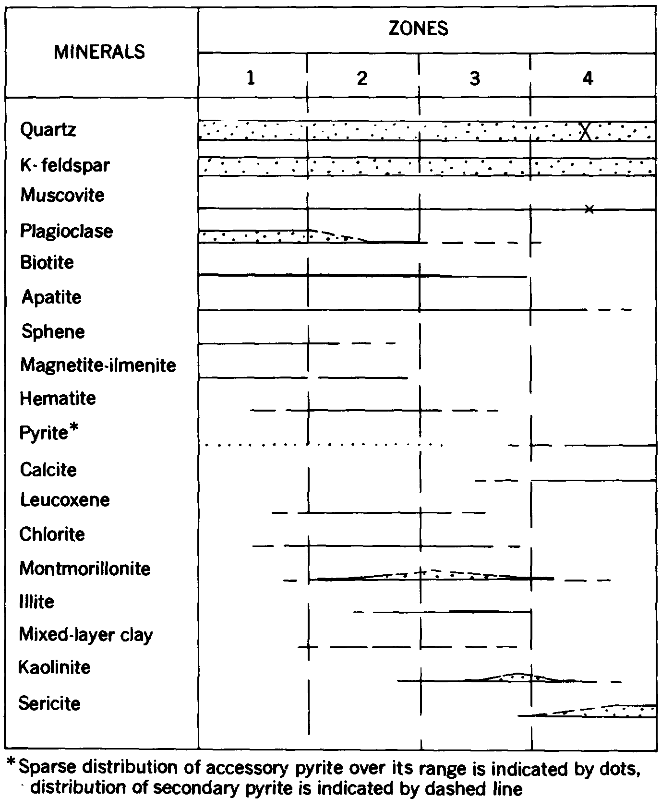
D. Microcline-quartz-plagioclase-biotite gneiss



E. Biotite-quartz-plagioclase and related gneisses

F. Amphibolite

FIGURE 20.—Triangular diagrams showing variations in mineralogic composition of rocks described for this report and of similar rocks examined by others.



*Sparse distribution of accessory pyrite over its range is indicated by dots, distribution of secondary pyrite is indicated by dashed line

EXPLANATION

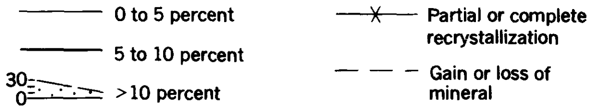


FIGURE 21.—Stability and relative abundance of minerals in the progressive alteration of biotite-muscovite granite.

Sericitized granite in zone 4 (fig. 4*c*) is a hard rock and is composed chiefly of K-feldspar, sericite, clay minerals consisting of sparse kaolinite, montmorillonite and illite, and partly recrystallized quartz. Residual muscovite and secondary pyrite are present also. Calcite, which occurs in shears and as disseminations in this rock, may have been introduced during the ore-depositing stage.

A similar sequence of mineral changes (table 10) is obtained in zone 2 of altered microcline-quartz-plagioclase-biotite gneiss. In contrast to those in the granite, however, zones 3 and 4 in gneissic rock are wider owing mostly to the greater abundance of unstable minerals such as plagioclase. Strongly argillized gneiss is of an intermediate hardness; clay minerals almost completely replace plagioclase, but biotite retains its textural character (fig. 22) and fresh greenish-black color in zone 3. Quartz and K-feldspar are unaltered.

Where examined in the intermediate ore zone at the Essex mine, sericitized gneiss of zone 4 typically is hard and sheared, and contains recrystallized quartz, microcline, sericite, clay minerals, and pyrite (figs. 4*C* and 7 *G*, *H*). Sericite occurs as radial clusters or matted aggregates of elongate flakes in clay minerals pseudomorphic after plagioclase (fig. 18, *A*–*D*). Most of this clay-sized material is sericitic in composition. Coarse sericite is most abundant adjacent to the vein and diminishes in abundance and crystal size toward the zone 3 boundary; sericite generally is not present where biotite is unaltered. The clay between sericite clusters possibly is kaolinite or montmorillonite; however, the precise position of kaolinite and montmorillonite observed in thin sections of these rocks is uncertain. Staining of the sample with sodium cobaltinitrite reveals that K-bearing minerals are somewhat concentrated at the zone 3–zone 4 interface. Thin sections of the rock indicate the presence of microcline, slightly altered biotite, and an illitic clay mineral altered partly from plagioclase. Very little chlorite was formed from the biotite; the alteration of biotite is to illite, kaolinite, and sericite. Most modes and thin section observations of altered gneiss from zone 4 show that quartz, although partly recrystallized, is not markedly lost in the alteration process, and the hardness of the altered rock apparently is related to recrystallization rather than introduction of quartz. However, fine-grained quartz is commonly found within shears at vein margins and coating openings along foliation planes, and comb quartz fills some of the more open spaces in the rock. Microcline locally contains a kaolinitelike clay filling along fractures, but otherwise is unchanged.

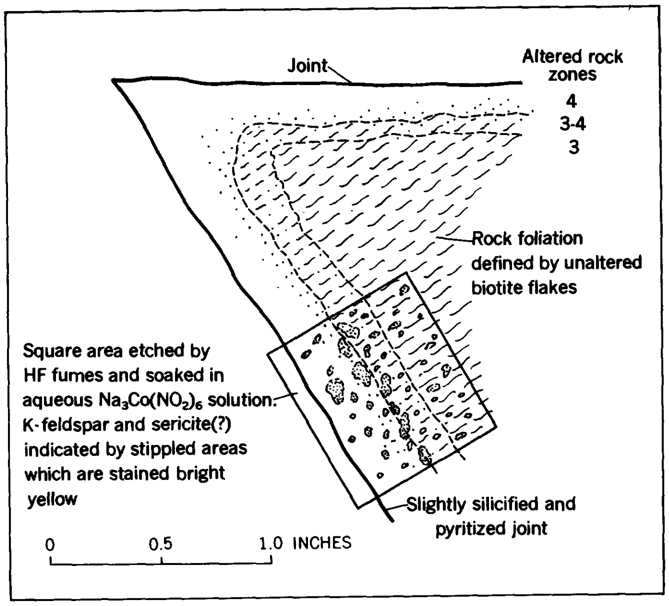


FIGURE 22.—Sketch of small-scale wallrock alteration effects along slightly mineralized joints in the host rock.

TABLE 10.—*Modal analyses, in volume percent, of fresh and altered microcline-quartz-plagioclase-biotite gneiss, Essex mine*

Zone	1		2		3		3-4		4	
Sample	46	47	48a	63	58	49	64	48	58a	56
Mineral										
Quartz	31.2	28.5	32.2	40.7	24.3	27.2	36.8	19.8	16.7	38.4
Microcline	14.7	27.9	12.8	17.8	15.5	20.9	20.0	18.8	14.3	3.4
Plagioclase	38.9	24.9	30.8	26.7	1.2	21.3	1.4	5.5		
(Antiperthite) [1]	2.3	1.7	4.1	6.4		2.9				
Biotite	6.2	8.3	7.5	3.5	[2] 10.1	1.6				
Hornblende	.9	3.5								
Magnetite [3]	1.3	.8	2.2	2.2	2.9	2.9		2.3		1.9
Pyrite		.2						.4		
Leucoxene	.1		.9			.4		1.4		
Apatite	1.1	.8	.6	.4		1.8				.8?
Sphene		.1								
Clay minerals	2.6	2.2	6.3	2.2	45.8	19.7	41.9	51.0	66.7	54.5
Calcite [4]	.6	.8	2.5			1.1		.9	2.4	.8
Epidote		.2								
Total	99.9	99.9	99.9	99.9	99.8	99.8	100.1	100.1	100.1	99.8
Points counted	837	874	500	480	168	191	579	218	84	466

[1] Consists of scattered, large Na-plagioclase crystals that contain small, irregular patches of microcline.
[2] Unusually biotite-rich layer in gneiss.
[3] Magnetite is commonly altered to hematite.
[4] The amount of calcite increases in samples 48a, 48, 49 and 58a due to proximity to ore veins that have calcite in gangue or disseminated in adjacent wallrock.

Localities where samples were collected:
46, 3 ft into footwall of vein.
47, 1½ ft into footwall of vein.
48a, 1½ in. into footwall.
63, 4½ in. into footwall.
49, 2 in. into footwall.
58, 8-10 in. into footwall, sheared rock.
64, 3½ ft into footwall.
48, About 1 in. into the footwall.
56, 2 in. into footwall of 2-ft vein.
58a, Silicified part of sample 58 (same thin section).

GRANODIORITE AND BIOTITE-QUARTZ-PLAGIOCLASE GNEISS

These rocks of moderate abundance in the area represent those that contain moderate to sparse amounts of K-feldspar. Fresh and altered mafic phases of the granodiorite were sampled in the M. and M.-Dixie and Wallace tunnels in the Chicago Creek area, and in the Hayes and Wheeler tunnel in the Central City district (pl. 1). Samples of biotite-quartz-plagioclase gneiss from several mines in the districts, including those from the Essex, Banta Hill, and Cherokee mines reported here, were studied.

TABLE 11.—*Modal analyses, in volume percent, of progressively more altered granodiorite, M. and M.-Dixie tunnel*

Zone	1		2		3	4	
Sample	4	124	125	126	127	128	129
Mineral							
Quartz	32.6	34.2	30.3	30.2	16.9	32.0	24.7
Plagioclase	26.1	33.4	14.7	1.2			
K-feldspar	6.2	2.1	2.5	1.5	1.2		
Biotite	22.4	21.2	20.1	19.3	21.8	16.2	
Hornblende	.6	.4					
Sphene	4.5	.7	.3	.2			
Apatite	1.6	.7	.3	.8	.3	1.3	.3
Pyrite	.3	.4		.1	.6	.2	.3
Hematite			.7	4.0	1.5	1.9	1.9
Magnetite	2.4	2.3	.2				
Clay minerals	2.6	4.5	28.0	38.7	53.9	45.1	68.8
Leucoxene			2.5	3.8	3.6	1.9	2.9
Carbonate	.2		.2	.2	.3	1.2	.9
Total	99.5	99.9	99.8	100.0	100.1	99.8	99.8
Points counted	1,359	719	722	522	330	510	312

The fresh granodiorite is a medium- to dark-gray medium-grained, competent rock composed of quartz, plagioclase biotite, K-feldspar, and accessory minerals (tables 11, 12). A compositional layering consisting of alternating biotite-rich and quartz-feldspar rich layers in part accounts for the mineralogical variations (observed in samples) of the rock.

TABLE 12.—*Modal analyses, in volume percent, of progressively more altered mafic border facies of granodiorite intrusive, Hayes and Wheeler tunnel*

Zone	1	2	3b	3c	4	
Sample	41	42	44a	44b	45a	45b
Mineral						
Quartz	24.3	20.2	15.9	21.3	20.0	24.2
K-feldspar	1.2	2.2	4.6	2.2	9.4	2.2
Plagioclase (An $_{26-28}$)	40.0	28.2				
Antiperthite [1]	.9	4.0				
Biotite	18.7	28.8	21.2	1.2	.3	
Clay minerals	4.6	8.0	48.3	60.4	54.3	59.5
Calcite	tr	1.0	tr	tr	tr	2.6
Chlorite	.4	.1	tr	2.7		
Mica [2]					3.4	2.4
Magnetite	2.0	2.1	3.5	5.4	1.1	.2
Pyrite	.6	.1	.5	.5	2.8	3.6
Hematite				.2	1.6	.2
Leucoxene		2.6	3.1	3.9	5.4	2.4
Fluorite			tr		tr	.2
Apatite	2.7	2.5	3.0	2.2	1.6	2.4
Sphene	4.8	tr	.2			
Total	100.2	99.8	100.3	100.0	99.9	99.9
Points counted	1,045	801	426	409	385	417

[1] Consists mostly of microcline blebs in plagioclase.
[2] Secondary muscovitelike mica altered from biotite.

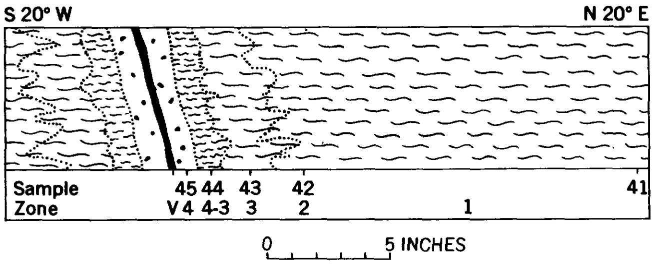


FIGURE 23.—Diagrammatic section of altered granodiorite, Hayes and Wheeler tunnel. Foliation shown by a long dash representing fresh biotite, a shorter dash represents altered biotite. Dots represent isolated remnants of biotite and no foliation. Approximate zone boundaries are marked by fine dotted lines. Note irregular contacts.

The boundaries between the fresh rock of zone 1 and the weakly argillized rock of zone 2 in granodiorite are remarkably gradational and are only slightly less so between zone 2 and the soft, light-green strongly argillized rock of zone 3. These gradational contacts contrast markedly with the abrupt contact between zone 3 and the hard, sericitized rock of zone 4. The boundaries between zones at the Hayes and Wheeler tunnel (fig. 23) are irregular; and it is believed that the ragged nature of the zone contacts result from the nearly right-angle intersection of the vein and the rock foliation, which is concordant with compositional layering. In contrast, at the M. and M.-Dixie mine, where the vein is nearly parallel to the foliation, the contacts between zones are more regular.

Granodiorite in zone 2 is moderately hard and appears texturally and mineralogically similar to rock in zone 1; a slight local softening is caused by progressive alteration of plagioclase (fig. 8, *C–F*), and hornblende into clay minerals. Sphene is altered to leucoxene. The order of abundance of clay minerals, as indicated by X-ray diffraction patterns is montmorillonite> illite>mixed-layer clay>kaolinite>chlorite (table 13). Biotite generally is unaltered except near microscopic fractures and open foliation planes now filled with calcite. Quartz, microcline, magnetite, pyrite, and apatite are unaltered in zone 2 (fig. 24); however, sphene is almost completely altered to leucoxene. Calcite, which occurs in veinlets and fractures, seems to have been introduced from the vein; some of the Ca, though, may have been furnished through the alteration of the plagioclase and hornblende.

Intensely argillized granodiorite of zone 3 is characteristically a soft clay mineral-biotite-quartz rock. The boundary with hard, weakly argillized rock of zone 2 is completely gradational, but the boundary with zone 4 rock is a sharp mineralogical-textural boundary but a less abrupt hardness boundary. Zone 3 may be arbitrarily subdivided into (a) soft, crumbly, greenish-white clay mineral-biotite-quartz rock, and (b) moderately soft, white clay mineral-biotite-quartz rock which may appear slightly darker gray on the veinward side.

TABLE 13.—*Representative clay minerals and sericite in altered granodiorite, Hayes and Wheeler and M. and M.-Dixie tunnels*

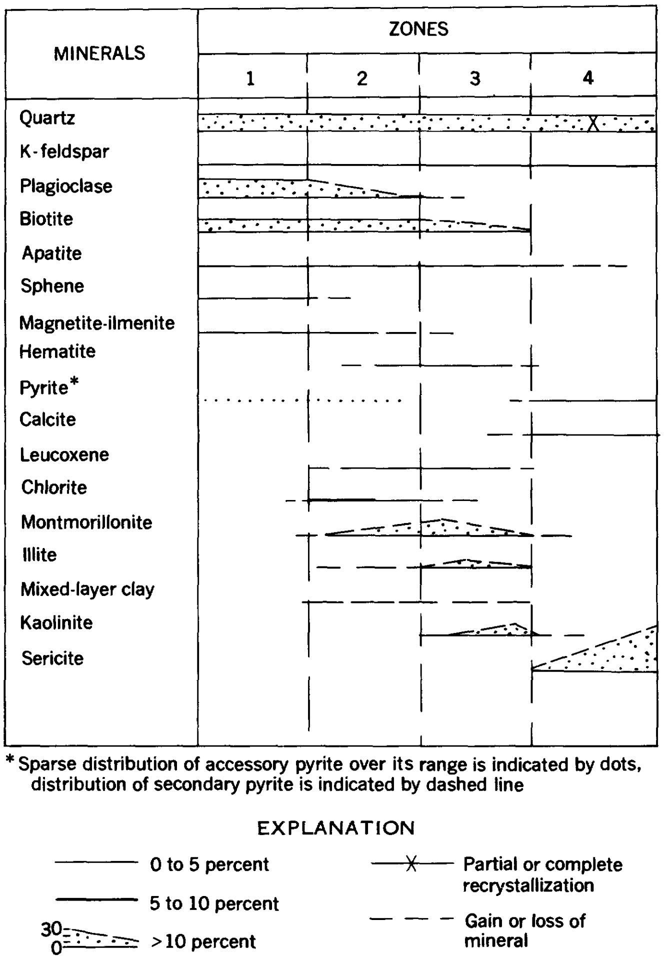
[s=sparse, m=moderate, a=abundant, va=very abundant, vs=very sparse, ?=uncertain, ---=not identified]

Altered-rock zones	Clay minerals					Sericite	Sample
	Montmorillonite	Mixed-layers [1]	Chlorite	Illite	Kaolinite		
1	s	m	s	a	s	---	4
2	a	s	s	m	m–s	---	42
2	s	s	---	m	m–a	---	126
3	va	s	s	vs	m	---	43
3	a	m	---	s	m–a	---	127
3	a	s	?	s	a	---	[2] 44
3–4	m	vs	---	?	m	m–a	45
3–4	s	m	---	?	m–a	m	129

[1] Mixed-layer clay mineral, generally random illite-montmorillonite mixtures.

[2] Separation of illite and sericite in sample on basis of crystallinity as shown by X-ray diffraction peak.

Localities where samples were collected:
4, M, and M.-Dixie, fresh wallrock on dump.
42, Hayes and Wheeler, fresh rock 5.5 in. into hanging wall from edge of vein.
126, M. and M.-Dixie, greenish clay zone.
43, Hayes and Wheeler, green clay zone 3 in. into hanging wall from edge of vein.
127, M. and M.-Dixie, white clay zone.
44, Hayes and Wheeler, white clay zone 1.5 in. into hanging wall from edge of vein.
45, Hayes and Wheeler, silicified zone 0.5 in. into hanging wall from edge of vein.
129, M. and M.-Dixie, green clay silicified zone adjacent to vein.



*Sparse distribution of accessory pyrite over its range is indicated by dots, distribution of secondary pyrite is indicated by dashed line

FIGURE 24.—Stability and relative abundance of minerals in progressively more altered granodiorite.

The soft greenish-white clay mineral aggregate of subzone (a) consists dominantly of montmorillonite, some kaolinite, and less illite; a mixed-layer (illite-montmorillonite) clay mineral and chlorite are also present in sparse amounts. The whole clay mineral of subzone (b) is dominantly kaolinite.

Biotite is not pervasively altered in zone 3 (fig. 7), but where it is partly altered along cleavage planes and at ragged edges it is bleached and replaced by mixed-layer silicates (illite and montmorillonite) and calcite. Chlorite is also present along the edges of some biotite flakes. Quartz is not appreciably altered, but in part is recrystallized; the number of strained crystals decreases, but the content of quartz as determined from modal analysis remains nearly constant across the entire zone of altered wallrocks. Apatite is unaltered, some magnetite is stained with hematite, and pyrite is more abundant than in zone 2.

Sericitized granodiorite (zone 4) is a hard, dense, fine-grained, medium greenish-gray rock. The boundary with zone 3 is distinct and is marked by the absence of unaltered biotite. The appearance and modal content of K-feldspar and most of the quartz are not changed, but some quartz may be partly recrystallized. Minor amounts of kaolinite and montmorillonite are found in zone 4. Apatite, hematite, and pyrite are present; however, apatite and hematite disappear toward the vein whereas pyrite increases veinward.

Biotite-quartz-plagioclase gneiss, which is similar to the granodiorite in mineralogic composition (table 14) is similarly altered. However, in the eastern part of the area, and in the peripheral ore zone (pl. 1), a different alteration pattern is evident in these rocks. At the Banta Hill and Cherokee mines, for example, the pronounced argillization phase is absent, and zone 2 grades abruptly into sericitized rock of zone 4. In general the entire alteration zones bounding the veins is narrow, amounting to only a foot or so.

TABLE 14.—*Modal analyses, in volume percent, of fresh and altered biotite-quartz-plagioclase gneiss, 150-foot level, Essex mine*

Zone(s)	1	2	2–3	3a	3c	3–4	4
Sample	19	20	21b	21a	22	23a	21c
Mineral							
Quartz	38	32	40	39	37	43	40
K-feldspar	9	--------	2	2	2	6	3
Plagioclase	[1] 42	22	38	--------	--------	--------	--------
Biotite	5	6	7	14	4	--------	--------
Clay minerals	3	37	9	29	52	42	43
Calcite	.7	1	1	16	5	8	14
Hematite	.1	1.9	1.6	--------	--------	.2	--------
Magnetite	.1	.2	--------	--------	.3	.6	.2
Pyrite	.1	--------	.6	--------	--------	--------	--------
Leucoxene	.9	.3	.5	--------	--------	--------	--------
Apatite	.2	.1	.1	--------	.2	.2	.2
Sphene	.4	--------	--------	--------	--------	--------	--------
Total	99.5	100.5	99.8	100	100.5	100.0	100.4
Points counted	785	900	759	256	523	437	482

[1] 2 percent antiperthite included.

Localities where samples were collected:
19, Least altered gneiss, 5.5 ft into footwall.
20, Weakly altered gneiss, 4.5–5 ft into footwall.
21b, Moderately altered gneiss, 3 ft into footwall, but close to unmineralized shear.
21a, Moderately altered gneiss (white clay-biotite rock).
22, Moderataly altered gneiss adjacent to silicified shear, 3-12 in. into footwall of main vein adjacent to silicified zone.
23a, Intensely altered gneiss on hanging wall.
21c, Intensely altered gneiss (hard, silicified rock) along shear 18 in. into footwall from main vein.

In strongly foliated biotite-quartz-plagioclase gneiss at the Essex mine (fig. 25) zone 4 follows certain planes into the rock, and sulfides were deposited farther from the vein than was possible in granodiorite.

QUARTZ DIORITE

Quartz diorite is not quantitatively an abundant rock and occurs chiefly in the Freeland-Lamartine and Lawson-Dumont-Fall River districts, but because of the high proportion of minerals unstable in the hydrothermal (and weathering) environment it is more completely and abundantly altered. The fresh rock is dark gray, medium to coarse grained, and has a fair to good foliation where metamorphosed. It is composed (table 15) chiefly of quartz, hornblende, plagioclase (An_{38-54}), and biotite (fig. 8); K-feldspar is sparse or absent. The biotite content varies in the fresh rock (Harrison and Wells, 1956, p. 46), and no tendency toward biotitization is implied in table 14 by the apparent increase in biotite. Secondary minerals include clay minerals, calcite, chlorite, leucoxene, and epidote. The altered quartz diorite near silver-bearing galena-sphalerite veins is a hard, weakly argillized rock (zone 2) that in turn grades into a broad, soft-green to white, strongly argillized clay mineral-biotite-quartz rock (zone 3). A very narrow, silicified rind (zone 4) adjacent to the vein contains little or no clay mineral, but abundant sericite. The relative distribution of minerals

TABLE 15.—*Modal analyses, in volume percent, of progressively more altered quartz diorite, Jo Reynolds and Nabob mines*

Zone	1	1	2	2	3	4
Sample	70	70a	[1] 16	80	77	[2] 75
Mineral						
Quartz	14.8	11.4	10.7	13.2	22.0	25
Plagioclase	37.4	43.7	--------	30.7	--------	--------
K-feldspar	.4	.4	1.2	.2	--------	tr
Biotite	7.9	7.0	20.0	23.1	20.3	--------
Hornblende	25.3	23.4	25.8	4.0	--------	--------
Apatite	.7	1.2	.8	.6	.6	--------
Sphene	.9	1.8	--------	--------	--------	--------
Magnetite	4.6	6.0	.4	.5	1.8	--------
Hematite	.1	--------	--------	.1	.5	--------
Pyrite	.7	--------	--------	.8	.2	--------
Clay minerals	6.6	5.1	40.7	26.0	53.9	65
Calcite	.5	--------	.3	--------	.1	10
Chlorite	--------	--------	--------	.6	.2	--------
Leucoxene	--------	--------	--------	.2	--------	--------
Epidote	.1	--------	--------	--------	--------	--------
Total	100.0	100.0	99.9	100.0	99.6	100
Points counted	895	500	725	966	807	--------

[1] Fine-grained facies.

[2] Megascopic estimation.

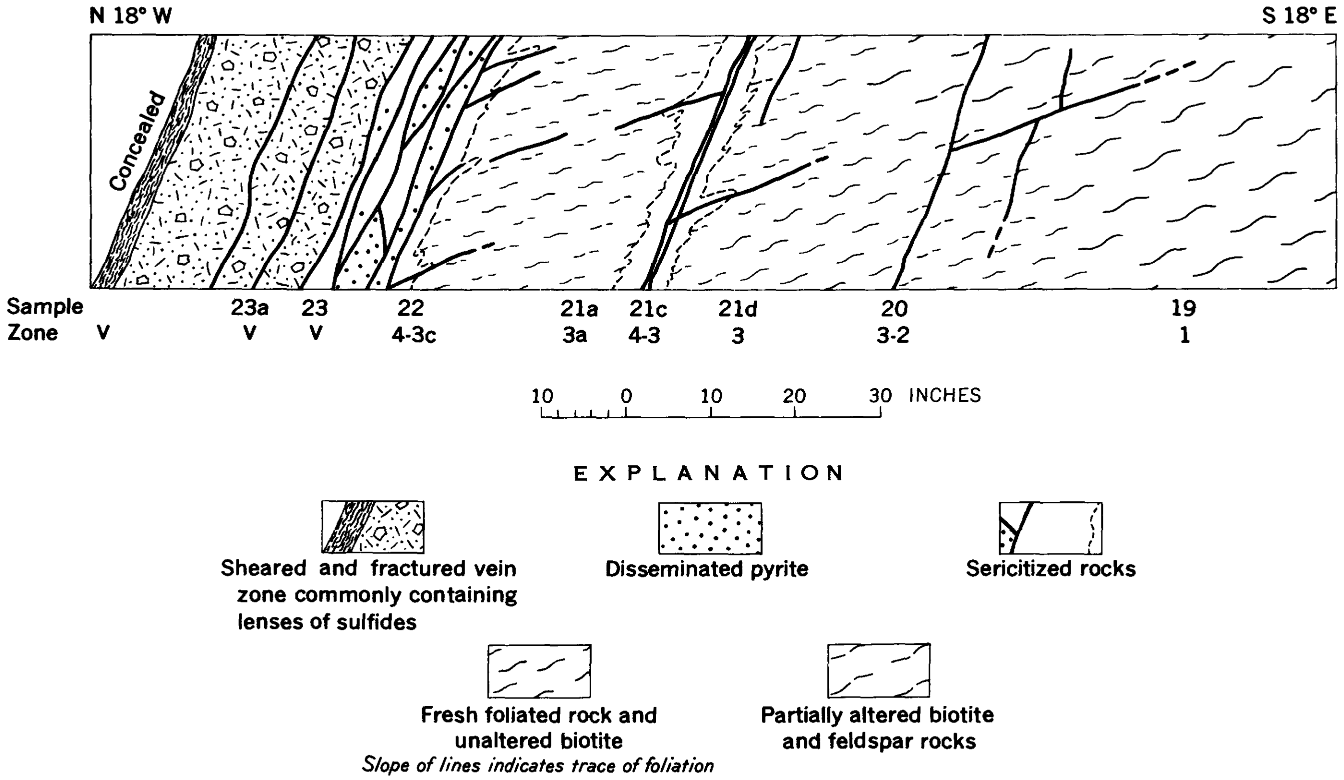


FIGURE 25.—Diagrammatic section showing altered biotite-quartz-plagioclase gneiss sample location, 150-foot level, Essex mine, Central City district, Colorado.

in the alteration zones (fig. 26) follows the pattern established in other rocks.

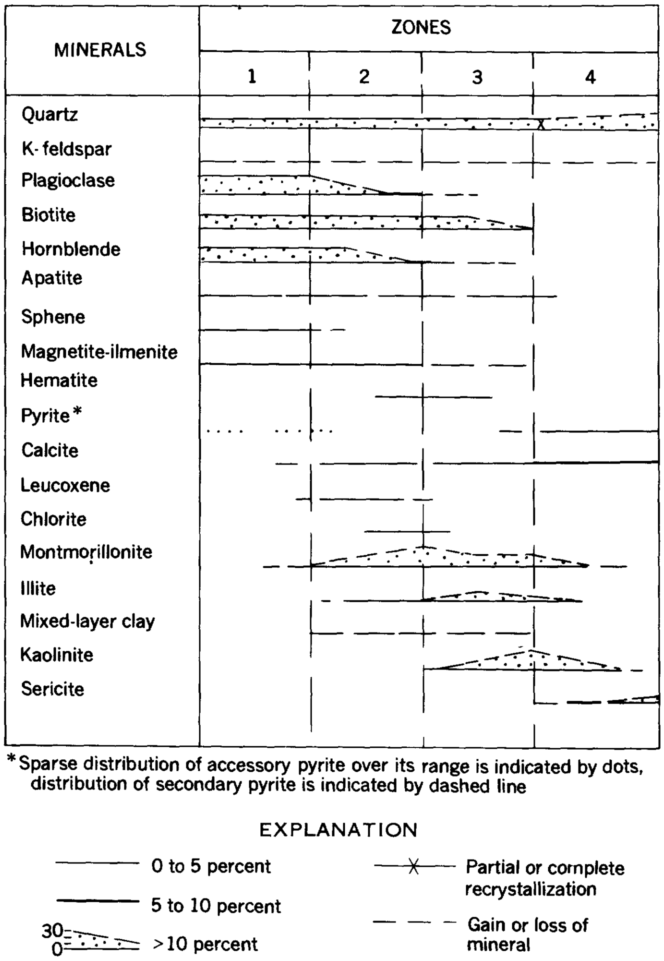
In the weakly argillized quartz diorite of zone 2, the plagioclase grains are replaced by clay minerals along fractures and twin and cleavage planes (fig. 8). With further replacement these patches and streaks of clay minerals expand until a large proportion of the crystal is enveloped. Hornblende is relatively unaltered in the outer part of zone 2, although there is incipient formation of clays along cracks, edges, and cleavages of the mineral. Toward the vein, however, hornblende becomes progressively more altered and is almost completely replaced by clay minerals at the boundary between zone 2 and zone 3. Biotite and quartz are not visibly altered in zone 2.

Montmorillonite is a prominent secondary constituent in zone 2, but mixed-layer (random montmorillonite-illite) clay minerals, chlorite, illite, and kaolinite are increasingly more important constituents in zone 3.

The strongly argillized quartz diorite in zone 3, which is unusually wide in proportion to the width of the associated vein, is composed dominantly of clay minerals and consequently is very soft. Near zone 2 the altered rock is green because of a high content of montmorillonite, but veinward it becomes progressively lighter in color until it is nearly white which results from kaolinite displacing montmorillonite as the dominant alteration product. The clay minerals almost completely replace the plagioclase and hornblende; biotite is unchanged in the montmorillonite-rich rock and is only slightly altered on the veinward size of the zone. Quartz is unaltered. Of the accessory minerals, magnetite is partly altered to hematite but apatite is unchanged.

Microscopic examination of the strongly argillized rock shows the clay minerals to replace the primary minerals as intimately intermixed assemblages in which the individual grains are virtually indistinguishable. In some specimens examined, however, the original plagioclase seems to be replaced chiefly by kaolinite, and the hornblende, which is recognized by its textural relations with quartz (fig. 9*A*), is replaced by a mixture of montmorillonite, illite, and random mixed-layer clay minerals.

In the veins examined, the intensely altered quartz diorite of zone 4 forms a narrow, hard, dark-brown, partly silicified rind bordering the vein. The biotite and clay minerals are largely altered to sericite; magnetite is oxidized to hematite, part of which has been replaced in turn by limonite. Quartz persists essentially unaltered, but some grains appear to have been recrystallized. Sericite and calcite commonly are also present in fractures and the latter also is disseminated through the sericitized rock.



*Sparse distribution of accessory pyrite over its range is indicated by dots, distribution of secondary pyrite is indicated by dashed line

FIGURE 26.—Stability and relative abundance of minerals in progressively more altered quartz diorite.

AMPHIBOLITE

The amphibolite, also a minor rock constituent in the area, is a dark-gray to black, medium-grained rock composed mainly of hornblende, quartz, and plagioclase; biotite, microcline, apatite, sphene, and epidote (table 16) are minor constituents. The rock occurs mostly as small pods and lenses within metasedimentary gneiss units and was sampled in the R.H.D.-McKay and E. Calhoun mines in the Central City district, the Kitty Clyde and Foxhall mines in the Idaho Springs district, the Golconda mine in the Fall River district, and the Dorit tunnel in the Chicago Creek area. This rock commonly occurs also in other mines in the area studied.

No distinct weak and strong argillized zones are observed; hornblende and plagioclase are altered farthest from the vein and biotite persists closer to the vein. Adjacent to veins the amphibolite is extensively altered (fig. 9) to a soft, green rock containing montmorillonite, illite, mixed-layer clay minerals, and residual quartz (usually characteristic of zone 3). The quartz is not visibly recrystallized. Close to the vein sericite occurs sparsely as large, randomly oriented, fragile, biotite like flakes in the soft argillized rock.

TABLE 16.—*Modal analyses, in volume percent, of altered amphibolite, R.H.D.-McKay shaft*

Zone(s)	1-2	3
Sample	32	33
Mineral		
Quartz	12.1	14.4
Plagioclase	5.4	------
Hornblende	48.2	23.7
Epidote	2.6	------
Clay minerals	30.0	60.8
Accessories	2.2	1.6
Total	100.5	100.5
Points counted (approx.)	350	200

Localities where samples were collected:
32, Weakly altered gneiss approximately 1 ft. into footwall from vein.
33, Moderately altered gneiss on footwall.

GARNET-QUARTZ GNEISS

The garnet-quartz gneiss, which occurs as moderate-sized layers or lenses in the metasedimentary rock sequence, is a dense, weakly foliated, dark red-brown to black, fine- to medium-grained rock composed of garnet and quartz with biotite, and accessory pyrite, magnetite, carbonate, and apatite (table 17). The garnet and biotite content is considerably varied from interlayer to interlayer within this unit. Generally in the garnet-poor layers, biotite and plagioclase are the abundant minerals. Garnet-quartz gneiss was studied only in the Fall River area, and was sampled in the Golconda, Almadin, and Mary mines.

The alteration of garnet-quartz gneiss varies according to the rock mineralogy. The normal gneiss contains mostly garnet, quartz, and biotite. A narrow zone

TABLE 17.—*Modal analyses, in volume percent, of altered garnet-quartz gneiss, Golconda mine*

Zone(s)	1-2		2-3	4
Sample	380	382a	382b	382c
Mineral				
Quartz	42	33	31	32
Garnet	21	27	26	5
Biotite	15	25	22	------
Calcite	10	2	4	------
Clay minerals	6	7	13	55
Apatite	2	4	2	<1
Pyrite	3	------	1	6
Magnetite	<1	2	1	1
Total	<100	100	100	<100
Points counted	759	284	374	405

Localities where samples were collected:
380, Altered gneiss, 0.7 ft into hanging wall of vein 1.8 ft thick.
382a, Altered gneiss, 0.7–1.2 in. into hanging wall of veinlet 0.1-in. thick, 0.7–0.8 ft into footwall of hanging wall vein zone.
382b, Altered gneiss, 0.3–0.6 in. into hanging wall of veinlet.
382c, Altered gneiss, 0–0.3 in. into hanging wall of veinlet.

along the vein a few inches thick is changed to a hard bleached quartz-clay mineral rock that locally contains abundant disseminated pyrite; no soft argillized zone is present. The marked color changes in garnet and biotite are from reddish to greenish gray, and black to yellow brown. Some interlayers in the gneiss contain appreciable plagioclase, and these layers are altered to clay minerals as in other rocks. X-ray diffraction studies of bulk samples of altered garnet-quartz gneiss show that near the vein in zone 4 biotite is altered to muscovite like clay minerals and kaolinite, and garnet is partly altered along fractures to a mica-type clay mineral and quartz. Some secondary biotite also occurs in this zone. Unaltered, "islandlike", fragments of garnet are replaced locally by calcite near the vein. Thin-section studies indicate that biotite may be incipiently altered prior to alteration of the garnet, but garnet crystals are altered completely before all biotite grains are changed. Magnetite alters to pyrite. Quartz is unaltered, but near the veins and fractures fine-grained quartz is introduced or is the product of recrystallization.

BOSTONITE

Bostonite, which occurs mostly in dikes, is a lilac- to reddish-brown rock that contains abundant salmon-pink K-feldspar and sparse pyroxene phenocrysts in a fine-grained matrix consisting predominantly of plagioclase. The rock has been described by Phair (1952) and Wells (1960), and was sampled by the writer in the Mammoth, Banta Hill, Phoenix, and Diamond Mountain mines. Bostonite is not as abundant in the central and intermediate ore zones as it is in the peripheral ore zone; where it is cut by veins in these ore zones commonly it is bleached light gray to white, and the feldspar phenocrysts are softened and changed in color to green or chalk white. The matrix of the bostonite is composed largely of plagioclase which is completely altered to a soft white aggregate composed mostly of sericite and pyrite. Quartz is recrystallized. The K-feldspar phenocrysts are only partly altered. Magnetite is replaced by pyrite. Adjacent to the vein, a narrow hard silicified and pyritized rind is common.

In the eastern peripheral ore zone area, and specifically in the Banta Hill mine, the prominent soft clay-mineral zone was not observed in the altered bostonite. In alteration zones 2 and 3 along the Banta Hill vein the matrix and feldspar phenocrysts are argillized but the rock is not softened, and the magnetite is altered to pyrite. Zone 4 is a granular sericite-pyrite rock which faintly retains the original porphyritic texture.

GEOCHEMISTRY OF ALTERED ROCKS

Hydrothermal alteration bordering vein deposits has changed the relative abundance and distribution of elements in wallrock minerals. In the discussion of the geochemistry of altered wallrocks the data are considered in the light of three fundamental observations: (1) that the atomic structures of rock minerals contribute to physical-chemical behavior through (a) the unique character of constituent ions or atoms, (b) the bonding between ions, and (c) coordination possibilities between ions as set forth in Pauling's rules (Hauth, 1951); (2) that in the rocks close-packed oxygen ions, whose size is large in comparison with Ca^{+2}, Mg^{+2}, Fe^{+3}, Al^{+3} and Si^{+4} ions, constitute as much as 92 percent by volume, and the corollary that the remaining cations and anions fill the interstices; and (3) that the partial or complete replacement of ions in a crystal structure is common in oxygen-bearing rock minerals and is regulated by the space requirements of atoms and ions, the temperature, ionization potential, and the crystal structure (Rankama and Sahama, 1950, p. 125). In addition, in most environments Si^{+4} combines with O^{-2} to form silica tetrahedrons, $(SiO_4)^{-4}$. The construction of silicate minerals is based on the variable manner in which the sheets of silica tetrahedra can be joined to other silicate sheets by sharing one or more O^{-2} (Henderson, 1951). The background data for the discussion of ion position and ion replacements in crystal structures and the average distribution of ions in wallrocks and minerals has been drawn mostly from Bragg (1937), Rankama and Sahama (1950), Mason (1952), and Green (1953).

Consideration of the chemistry of altered wallrocks is facilitated by an arbitrary division of the subject into two parts: (1) the geochemistry of the elements that are the major constitutents of the alteration minerals, and (2) the "trace element" constituents occluded with the alteration minerals. The "major constituent" elements have been determined by conventional methods of chemical analysis, and the "trace element" constituents have been determined by semiquantitative spectrographic methods. The powder pH of a few representative rocks also has been determined.

MAJOR CONSTITUENT ELEMENTS

Chemical analyses of unweathered wallrocks and their altered equivalents provide a basis for calculating the abundance and distribution of elements in progressively more altered rocks. The major distribution trends observed in these studies are the veinward addition of K^{+}, Fe^{+2} (and total iron), C^{+4}, H^{+}, S^{-2}, and less commonly Al^{+3} ions concurrent with losses in Si^{+4}, Na^{+}, Ca^{+2}, Fe^{+3}, and Mg^{+2} ions.

METHODS FOR PRESENTING CHEMICAL ANALYSIS DATA

Several methods have been used for the rapid visual comparison of chemical data. Gravimetric weight percentages of oxides in the fresh and altered wallrocks (table 18) are not readily adaptable to a discussion of ion movement and are not easily compared with spectrographic data (figs. 36 to 62). Many writers have modified oxide weight-percentage analyses, although Schwartz (1953) preferred to discuss the data as reported. Lovering (1941) computed analyses into molecular percent of equivalent volumes, and Butler (1932), Butler and others (1920), Sales and Meyer (1948), and Lovering and others (1949), converted oxide weight percent into grams per cubic centimeter. Barth (1948, 1955) calculated the number of cations and anions in a "standard rock cell." A comparison of some of these methods is afforded in table 19, and shown on figure 27. These data show that regardless of the scales of measure used, the slopes of lines connecting plotted values are remarkably similar.

In the subsequent discussions it is assumed that the movement of cations and anions is within a network of close-packed oxygen ions rather than an unordered movement of oxides in space. The movement of individual ions, including oxygen, probably is variable; and it is recognized that Si and Al undoubtedly have strong structural expression (bonding) as oxide molecules and are therefore less mobile than the alkali, alkaline earth, and metal ions present. A further fundamental assumption that there is no significant volume change caused by the alteration process seems warranted from thin section and other observations.

TABLE 18.—*Chemical analyses, in weight percent, of fresh and altered wallrocks, Gilpin and Clear Creek Counties, Colo.*

[The location and rock type of all samples are indicated in table 1. Analysts: J. Theobold, 1-8; 14-18; and 31. L. N. Tarrant, 19-26; 32-34. P. L. D. Elmore, K. E. White, and S. D. Botts, 9-11. P. L. D. Elmore, K. E. White, S. D. Botts, and A, Sherwood, 12-13. H. F. Phillips, P. L. D. Elmore, P. W. Scott, K. E. White, 27-30; 35-40]

Analysis factor	Sample									
	1	2	3	4	5	6	7	8	9	10
Constituent:										
SiO_2	71.86	67.56	58.66	60.20	55.47	57.19	52.05	49.77	76.3	75.4
Al_2O_3	13.19	14.38	16.59	14.51	16.87	15.55	16.37	16.80	12.2	12.7
Fe_2O_3	1.24	.78	.96	3.56	4.11	4.47	3.28	1.62	.9	1.0
FeO	[1]1.80	[1]1.86	[1]3.28	[1]5.26	[1]5.03	[1]5.34	[1]7.58	[1]8.16	2.0	1.6
MgO	.55	1.12	1.56	2.33	2.36	2.23	2.71	1.68	.48	.50
CaO	1.29	.98	2.75	4.84	2.77	1.40	1.53	1.62	1.7	1.6
Na_2O	2.22	1.93	.22	2.93	1.57	.16	.08	.08	3.7	3.4
K_2O	5.85	7.05	6.86	2.71	2.58	2.66	5.71	6.92	1.1	1.2
H_2O-	.04	.10	.38	.02	1.95	1.75	.57	.52	.83	1.8
H_2O+	.57	.79	2.38	.70	2.92	3.88	3.34	2.02		
TiO_2	.41	.36	.51	1.48	1.37	1.66	1.69	1.78	.22	.18
CO_2	.34	2.78	4.76	.20	1.61	2.16	3.12	6.48	.20	.17
P_2O_5	.24	.11	.39	.63	.51	.69	.99	.99	.06	.05
MnO	.03	.04	.16	.14	.14	.16	.24	.70	.02	.02
BaO	.05	.07	.16	.12	.07	.09	.14	.12	------	------
F	.07	.06	.14	.21	.30	.34	.35	.39	------	------
Cl	(2)	------	------	------	------	------	------	------	------	------
S	.01	.00	.28	.15	.32	.17	.18	.47	------	------
Total	99.76	99.97	100.04	99.99	99.95	99.90	99.93	100.12	100	99.62
(1)	.03	.02	.13	.13	.21	.18	.20	.28	------	------
Total	99.73	99.95	99.91	99.86	99.74	99.72	99.73	99.84	100	99.62
Bulk density	2.66	2.64	2.52	2.78	2.57	2.46	2.70	2.65	[3]2.62	[3]2.56
Powder density	2.66	2.67	2.71	2.81	2.73	2.74	2.80	2.85	------	------
Porosity percent	.0	1.1	7.0	1.1	5.8	10.2	3.5	7.0	------	------

Analysis factor	Sample									
	11	12	13	14	15	16	17	18	19	20
Constituent:										
SiO_2	71.2	75.5	65.3	64.09	54.29	52.61	51.98	48.06	71.76	64.69
Al_2O_3	14.1	12.1	13.8	14.03	13.53	15.63	16.88	15.00	14.37	13.74
Fe_2O_3	2.5	2.3	6.7	3.44	5.28	2.71	4.69	8.77	1.13	3.56
FeO	1.4	.68	.82	[1]2.71	[1]5.10	[1]6.46	[1]4.77	[1]3.68	1.80	3.60
MgO	.55	.44	.74	1.14	2.42	5.67	2.90	2.08	.58	1.11
CaO	1.4	.12	.37	3.91	7.22	6.16	3.41	5.87	2.05	3.09
Na_2O	2.7	.11	.11	2.81	2.64	2.13	.47	.99	4.95	4.16
K_2O	1.2	4.6	4.7	4.18	2.40	2.85	2.66	1.43	2.02	1.61
H_2O-	3.4	1.6	1.5	.16	.22	.12	1.30	2.94	.03	.16
H_2O+				.38	.72	1.61	4.01	3.45	.33	.85
TiO_2	.20	.16	.43	1.31	2.37	.91	1.33	1.91	.27	.59
CO_2	.54	<.05	<.05	.12	.44	2.08	4.25	4.72	.54	2.37
P_2O_5	.04	.06	.20	.56	2.02	.33	.62	.46	.04	.16
MnO	.06	.01	.02	.09	.13	.15	.18	.18	.06	.17
BaO	------	------	------	.24	.17	.04	.06	.02	------	------
F	------	[4].61	[4].22	.59	.43	.32	.30	.24	------	------
Cl	------	------	------	.05	------	------	------	------	------	------
S	------	[5]2.0	[5]5.2	.03	.36	.22	.16	.32	------	------
Total	99.29	100.3	100.2	99.84	99.74	100.00	99.97	100.12	99.93	99.86
(1)	------	------	------	.25	.27	.19	.17	.18	------	------
Total	99.29	100.3	100.2	99.59	99.47	99.81	99.80	99.94	99.93	99.86
Bulk density	[3]2.47	[3]2.15	[3]2.58	2.70	2.80	2.79	2.47	2.32	2.67	2.67
Powder density	------	------	------	2.73	2.85	2.85	2.77	2.74	2.65	2.69
Porosity percent	------	------	------	1.1	1.8	2.1	10.8	15.3	------	.7

See footnotes at end of table.

TABLE 18.—*Chemical analyses, in weight percent, of fresh and altered wallrocks, Gilpin and Clear Creek Counties, Colo.*—Continued

Analysis factor	Sample									
	21	22	23	24	25	26	27	28	29	30
Constituent:										
SiO_2	65.61	68.61	72.12	76.33	75.10	53.68	72.0	65.5	58.1	59.5
Al_2O_3	13.05	14.10	9.83	12.30	13.47	17.35	11.9	17.7	14.3	14.9
Fe_2O_3	3.40	2.33	.39	1.08	1.11	5.06	1.9	1.3	4.5	4.9
FeO	4.05	1.72	[1] 6.99	1.75	1.32	.54	3.2	2.5	5.4	4.2
MgO	.77	.64	.52	.59	.65	2.43	1.5	.92	3.0	3.2
CaO	1.88	2.51	.03	2.41	2.24	2.65	1.7	.50	5.2	1.5
Na_2O	3.17	3.60	.08	3.13	2.57	.05	2.6	.04	1.2	.07
K_2O	1.50	1.15	3.69	1.02	.98	1.63	2.2	4.6	2.6	2.6
H_2O-	1.19	1.71	.15	.17	.49	8.35	1.0	4.0	1.8	5.6
H_2O+	1.31	1.22	1.29	.42	1.32	4.63				
TiO_2	.66	.54	.20	.27	.29	1.41	.58	.85	1.4	1.3
CO_2	2.78	1.49	.82	.01	.01	.01	.64	2.0	.70	1.2
P_2O_5	.17	.11	.03	.03	.04	1.43	.15	.06	.62	.53
MnO	.30	.14	.11	.03	.02	.02	.12	.17	.14	.14
BaO	----	----	----	----	----	----	----	----	----	----
F	----	----	----	----	----	.44	----	----	----	----
Cl	----	----	----	----	----	----	----	----	----	----
S	----	----	4.96	----	----	----	[5] .25	[5] .09	[5] .19	[5] .29
Total	99.84	99.87	101.21	99.54	99.61	99.68	99.74	100.23	99.15	99.93
(1)	----	----	1.24	----	----	.19	----	----	----	----
Total	99.84	99.87	99.97	99.54	99.61	99.49	99.74	100.23	99.15	99.93
Bulk density	2.30	2.52	2.52	2.65	2.26	1.91	----	[3] 2.71	----	----
Powder density	2.70	2.64	2.83	2.69	2.67	2.49	----	----	----	----
Porosity percent	14.8	4.6	10.9	1.4	15.5	23.1	----	----	----	----

Analysis factor	Sample									
	31	32	33	34	35	36	37	38	39	40
Constituent:										
SiO_2	55.27	48.19	51.39	54.62	52.4	51.4	49.7	43.1	52.8	49.7
Al_2O_3	16.17	15.66	21.80	18.97	8.2	9.4	7.8	9.7	11.1	10.5
Fe_2O_3	2.27	3.86	5.87	5.41	11.7	10.9	11.0	7.6	3.4	4.9
FeO	[1] 6.30	9.08	.45	.16	13.9	15.0	13.9	17.9	16.5	16.8
MgO	3.42	7.17	1.62	1.54	2.8	3.8	1.9	2.4	2.0	2.0
CaO	7.49	8.68	1.54	1.28	5.4	3.0	2.4	2.4	2.4	2.3
Na_2O	4.65	.92	.60	.22	.10	.09	.06	.06	.05	.09
K_2O	1.31	1.07	.62	.54	.49	1.8	1.2	.91	.21	.31
H_2O-	.03	.67	7.96	10.21	.74	.87	1.3	1.3	1.2	1.4
H_2O+	.99	3.17	6.77	5.68						
TiO_2	.98	.75	.94	.79	.32	.52	.30	.42	.42	.38
CO_2	.48	.01	.01	.01	.25	.13	6.4	9.6	5.6	8.0
P_2O_5	.24	.12	.18	.16	.70	.59	.94	.76	.89	.52
MnO	.19	.26	.01	.01	3.5	3.2	3.0	3.7	3.5	3.5
BaO	.02	----	----	----	----	----	----	----	----	----
F	.17	----	----	----	----	----	----	----	----	----
Cl	----	----	----	----	----	----	----	----	----	----
S	.01	----	----	----	[5] 1.2	[5] .94	[5] 1.2	[5] .86	[5] 1.3	[5] 1.2
Total	99.99	99.61	99.76	99.60	101.70	101.64	100.10	100.71	100.37	101.60
(1)	[1] .07	----	----	----	----	----	----	----	----	----
Total	99.92	99.61	99.76	99.60	101.70	101.64	100.10	100.71	100.37	101.60
Bulk density	2.92	2.80	1.90	1.93	[3] 3.31	[3] 3.28	[3] 3.10	[3] 3.24	[3] 3.18	[3] 2.63
Powder density	2.87	3.02	2.46	2.28	----	----	----	----	----	----
Porosity percent	----	7.3	22.6	15.3	----	----	----	----	----	----

[1] A calculated correction was made for FeO present as pyrite based on percent S. In making correction it is assumed all sulfur is present as pyrite.

[2] Not determined.

[3] Determined by author from representative samples.

[4] Special standard chemical method.

[5] Total S as S.

To represent the relative movement of ions or elements as a result of rock alteration a simple calculation is employed to determine the weight of the ions in an equivalent volume.[4] It is obvious that these calculations from analytical data give weights of the element. These data recalculated into percent (table 19) resemble Barth's values for the number of ions in a rock standard cell. In a very loose sense the term "ion" is used here to represent these elements. The gravimetric calculation is widely accepted and these data are therefore readily compared with similar data elsewhere

From these data, ratios of ions lost to those gained may be calculated (fig. 28). In these diagrams a value of less than 1 represents a loss of the ion relative to the amount of that ion in the original host rock; a value greater than 1 represents a gain of the ion. It is found empirically that oxygen varies least percentagewise in normally altered wallrocks (fig. 29). Thus the ratio of gain or loss of an ion may be plotted against that of oxygen to show the large variations in cation population gains or losses to that of a nearly constant anion population.

The writer realizes that the calculated values used herein are simple approximations that are mainly useful in describing a very complex chemical environment

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[4] Ions in an equivalent volume are computed from the chemical analysis by multiplying that portion of the oxide weight percent due to the ion by the bulk density of the rock to obtain small whole numbers in (g per cm³) × 10².

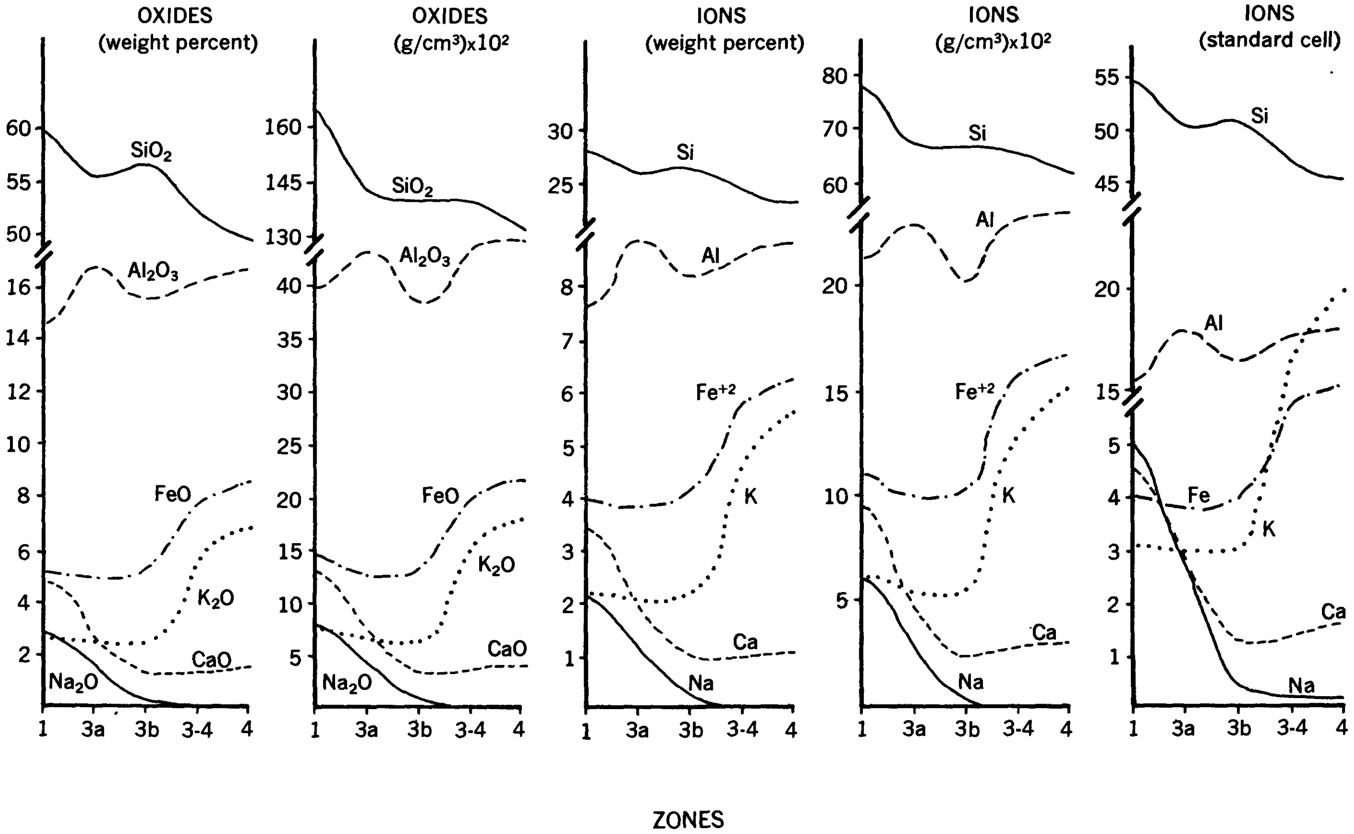

FIGURE 27.—A graphic comparison of methods of plotting chemical analysis data. Samples 4 through 8 are from fresh and altered granodiorite.

TABLE 19.—*A comparison of alternative methods of expressing chemical analysis data, granodiorite*

[Bulk density, 2.78; Powder density, 2.81; Percent pore space, 1.1]

Constituent	Oxide (weight percent)[1]	Oxide molecular proportion in equivalent volume	Oxide (g per cu cm)	Ion (weight percent)	Cation (percent)	Ions (g per cu cm) × 10[2]	Ions in rock standard cell[2]
SiO_2	60.20	61.1	1.6736	28.1	52.4	78.1	55.1
Al_2O_3	14.51	8.7	.4034	7.7	14.3	21.3	15.6
Fe_2O_3	3.56	2.5	.0990	2.5	4.7	6.9	2.5
FeO	[3] 5.26	9.2	.1462	4.1	7.7	11.3	4.1
MgO	2.33	3.5	.0648	1.4	2.6	3.9	3.1
CaO	4.84	5.3	.1345	3.5	6.5	9.6	4.7
Na_2O	2.93	2.9	.0814	2.2	4.1	6.1	5.1
K_2O	2.71	1.7	.0753	2.2	4.1	6.3	3.1
H_2O-	.02	.1	.0006	.12	.2	.2	--------
H_2O+	.70	2.4	.0195				
TiO_2	1.48	1.1	.0411	.9	1.7	2.5	1.0
CO_2	.20	.3	.0055	.05	.1	.1	.2
P_2O_5	.63	.3	.0175	.3	.6	.7	.4
MnO	.14	.1	.0039	.1	.2	.3	.1
BaO	.12	.05	.0033	.1	.2	.3	.04
F	.21	.7	.0058	.2	.4	.6	.5
S	.15	.03	.0042	.1	.2	.4	.2
	[3] .13						
Total	99.86	100.0	2.7796	53.57	100.0	148.6	[4] 95.0
(OH)	--------	--------	--------	--------	--------	2.0	4.0
O	--------	--------	--------	46.43	--------	129.7	155.3
Total	--------	--------	--------	100.0	--------	280.3	[5] 160.0

[1] Table 18, sample 4, M. and M.-Dixie mine, Chicago Creek area.

[2] Barth (1955) calculations.

[3] Calculated correction made for FeO present as pyrite, based on percent S. In making correction it is assumed that all sulfur is present as pyrite.

[4] Total of cations (minus F and S).

[5] Total of anions F+S+(OH)+O.

found in the alteration of a diverse and unhomogeneous rock assemblage. McKinstry (1957, p. 751) points out that the representativeness of the sample itself may be questioned more strongly than the accuracy of the analyses.

Chemical analyses of altered wallrocks viewed as ion distributions indicate that the major textural and structural differences between fresh and strongly altered rocks may be represented by a very small cation-anion redistribution. The number of ions gained or lost does not always balance (table 20), but there is generally a charge and size balance. While these summations are interesting they fail to indicate the contribution and effect of each zone to the chemistry of all the altered rock.

SILICON

Silicon constitutes 30 to 35 percent (calculated from grams of ion per volume) of wallrocks; however, the silicon ions are small (0.42A) and occupy a very limited volume (table 21). Silicon occurs in tetrahedral coordination with oxygen and is part of the fundamental silicate "building block" in quartz, feldspars, micas, amphibole, and clay minerals. Silicon generally is

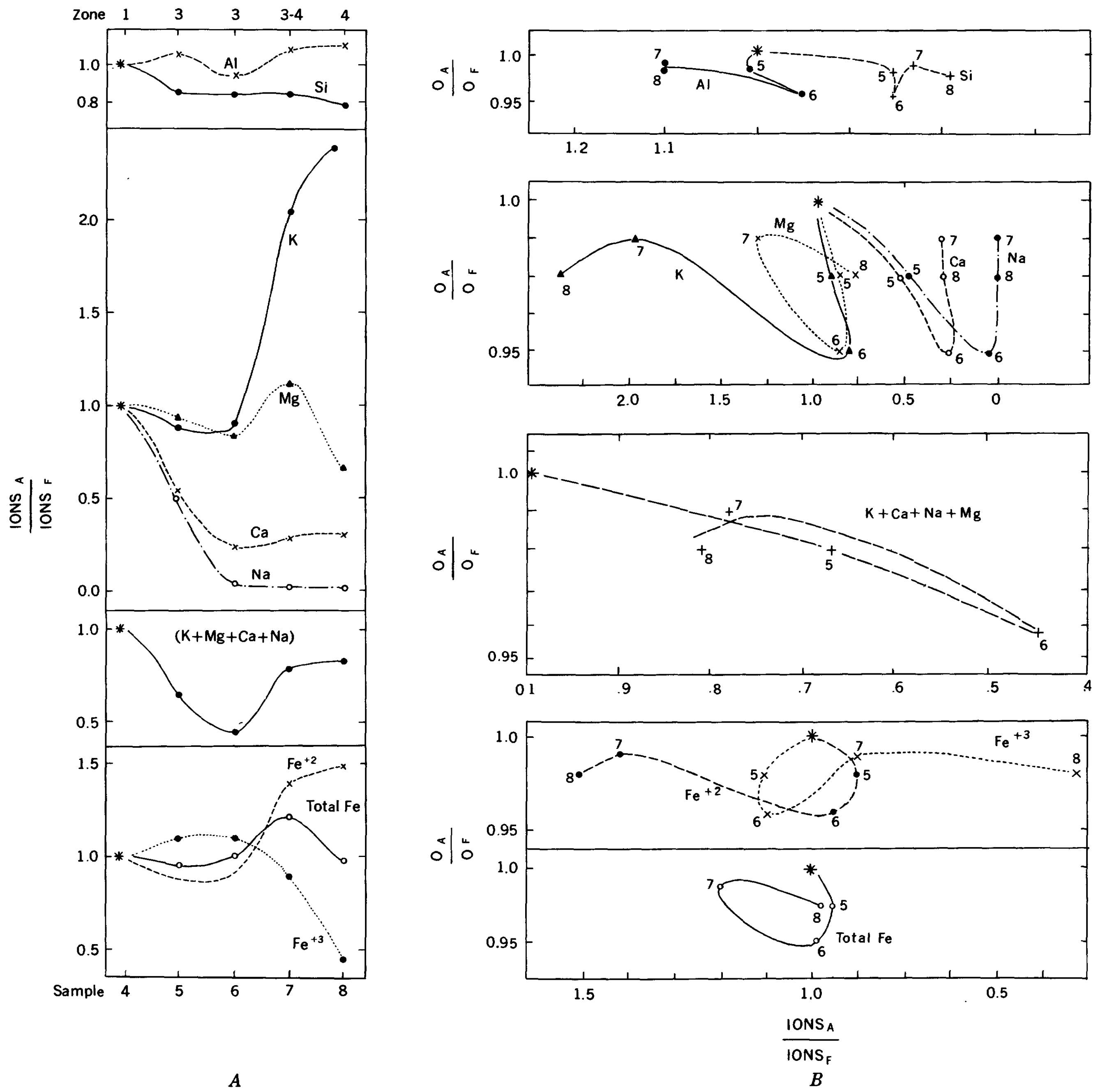

FIGURE 28.—Variations in the ratio of altered (A) to unaltered (F) ions in terms of (A) their zonal position and (B) the ratio of altered to unaltered oxygen ions in the sequence of fresh to most altered granodiorite. Asterisk denotes original host rock. Note exaggeration in vertical scales in (B).

lost in altered rock zones 1 through 4 (fig. 30), however, modal analyses show that the amount of quartz is not markedly changed. As quartz is recrystallized some silicon may be moved veinward; however, a source of abundant silicon ions results from the transformation of three-dimensional framework silicate structures into layer silicate structures. Barth (1955, p. 357) points out that the number of silicon tetrahedra drops from 12 to 8 as plagioclase structure is supplanted by sericite structure. Wallrocks undergoing alteration at depth therefore were a possible source of silica for vein quartz. The loss of silicon in altered granodiorite (fig. 28) and some other rocks (fig. 31) are shown in terms of oxygen content. Most silicon is lost as a consequence of the breakdown of plagioclase and hornblende to clay minerals. The change from clay minerals to sericite (muscovite in table 21) accounts for a very small loss of silicon.

TABLE 20.—*Ion redistribution resulting from the alteration of igneous and metamorphosed rocks*

[Data from Barth (1955) standard cell calculations]

a. Grandiorite (gneiss), M. and M.-Dixie mine			b. Quartz diorite (gneiss), Nabob and Jo Reynolds mines				c. Microcline-quartz-plagioclase-biotite gneiss, Essex mine		
Cations		Anions	Cations		Anions		Cations		Anions
Gained	Lost	Gained	Gained	Lost	Gained	Lost	Gained	Lost	Gained
Al 2.3 Ti .2 Fe^{+2} 2.1 K 4.8 Mn .4 C 7.8 P .4 18.0 (Represents 50.7 valences)	Si 10.1 Fe^{+3} 1.4 Ca 3.1 Na 4.8 Mg .8 20.2 (Represents 57.2 valences)	(OH) 6.5 F .5 S .5 7.5 (Represents 8.0 valences)	Ti 0.7 Fe^{+3} 4.2 C 3.3 P .2 8.4 (Represents 29.6 valences)	Si 4.1 Al .6 Fe^{+2} 2.1 Ca .2 Na .2 Mg 5.0 K 1.8 14.0 (Represents 34.8 valences)	(OH) 8.4 S .1 8.5 (Represents 8.6 valences)	F 0.3 .3 (Represents 0.3 valences)	Al 2.5 Ti .2 Fe^{+3} 4.0 Mg .4 K 4.2 P .1 11.4 (Represents 25.4 valences)	Si 6.2 Fe^{+2} .8 Ca 1.2 Na 6.8 C .1 15.1 (Represents 36.0 valences)	(OH) 3.1 S 7.2 10.3 (Represents 17.5 valences)

TABLE 21.—*Constituents of some silicate minerals*

[a, in percent by volume, after Rankama and Sahama (1950, p. 114); and b, in percent of ions]

Mineral	O		Si		Al		K		Na		Ca		Fe		Mg	
	a	b [1]	a	b	a	b	a	b	a	b	a	b	a	b	a	b
K-feldspar [2]	87.1	46.1	0.9	30.2	0.9	9.7	11.1	14.0	--------	--------	--------	--------	--------	--------	--------	--------
Albite [2]	93.3	48.7	.9	32.2	1.0	10.3	--------	--------	4.8	8.8	--------	--------	--------	--------	--------	--------
Anorthite [2]	91.6	44.0	.6	20.2	1.9	19.4	--------	--------	--------	--------	5.9	15.6	--------	--------	--------	--------
Quartz [3]	98.7	53.3	1.3	46.7	--------	--------	--------	--------	--------	--------	--------	--------	--------	--------	--------	--------
Pyroxene [4]	--------	41.2	--------	24.1	--------	1.2	--------	--------	--------	6.2	--------	4.9	--------	19.5	--------	1.6
Hornblende [5]	--------	[6] 38.6	--------	17.9	--------	5.4	--------	1.3	--------	1.1	--------	7.9	--------	25.6	--------	.7
Biotite [7]	--------	42.3	--------	17.8	--------	9.5	--------	7.1	--------	--------	--------	--------	--------	17.7	--------	5.4
Muscovite [7]	89.9	41.6	.6	21.0	1.9	17.7	7.6	8.7	--------	--------	--------	--------	--------	--------	--------	--------
Ca-montmorillonite [8]	--------	[6] 56.4	--------	25.6	--------	10.0	--------	--------	--------	--------	--------	1.9	--------	1.9	--------	2.3
Na-montmorillonite [8]	--------	[6] 54.6	--------	26.0	--------	10.6	--------	.5	--------	2.0	--------	.4	--------	2.7	--------	1.5
Illite [8]	--------	47.1	--------	25.9	--------	14.9	--------	5.5	--------	--------	--------	--------	--------	4.8	--------	1.8
Kaolinite [8]	--------	50.5	--------	25.8	--------	23.7	--------	--------	--------	--------	--------	--------	--------	--------	--------	--------

[1] Oxygen associated with amount of ions reported above, not from complete analyses.

[2] Feldspars, average, Winchell and Winchell (1951, p. 263).

[3] Quartz, assumed 100 percent SiO_2.

[4] Alkali pyroxene from alkali granite, Rankama and Sahama (1950, p. 146).

[5] Hornblende from granite, Rankama and Sahama (1950, p. 150).

[6] Oxygen also included from H_2O.

[7] Micas, average analyses, Rankama and Sahama (1950, p. 150).

[8] Clay minerals, representative analyses, Grim (1953, p. 370–373).

ALUMINUM

Aluminum occurs in fresh rocks in amounts averaging 7 percent; its ion has a small size (0.51A) and requires little volume in the crystal structure. Aluminum occurs in potassium and plagioclase feldspars, micas, amphiboles, pyroxenes, and their secondary clay-mineral alteration products (table 21). The ionic size and +3 valence of aluminum permit partial but not complete substitution of aluminum for silicon in silicate tetrahedra in feldspar structures as well as for iron and magnesium in octahedral coordination outside the silicon-oxygen framework in amphibole and mica structures. In clay mineral structures, however, aluminum replaces silicon with difficulty in the two dimensional SiO_2 sheets, and most aluminum occurs in $Al(O,OH)_6$ sheets.

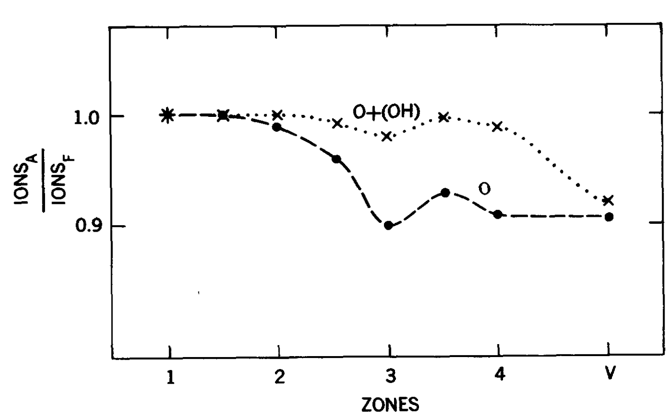

FIGURE 29.—Average variations of ratios of altered (A) to unaltered (F) oxygen and oxygen plus hydroxyl ions in altered rock zones. Oxygen replaced by sulfur in vein rock (V). Asterisk denotes original host rock.

Rocks containing abundant plagioclase are less stable in the hydrothermal environment than those contain-K-feldspar, and the plagioclase readily alters to three-layer clay mineral structures that contain less aluminum. Kaolinite formation directly from plagioclase is not common here. Thus plagioclase-rich rocks exhibit loss of Al ions while plagioclase-deficient rocks consistently show slight gains in Al ions (figs. 30, 31). The freed Al ions diffuse veinward where some may be fixed in an acid environment (powder pH of 4–5) in the Al-rich kaolinite and sericite structures (table 21).

SODIUM AND POTASSIUM

Sodium, which averages 2.4 percent in most rocks here, is generally more abundant than is potassium,

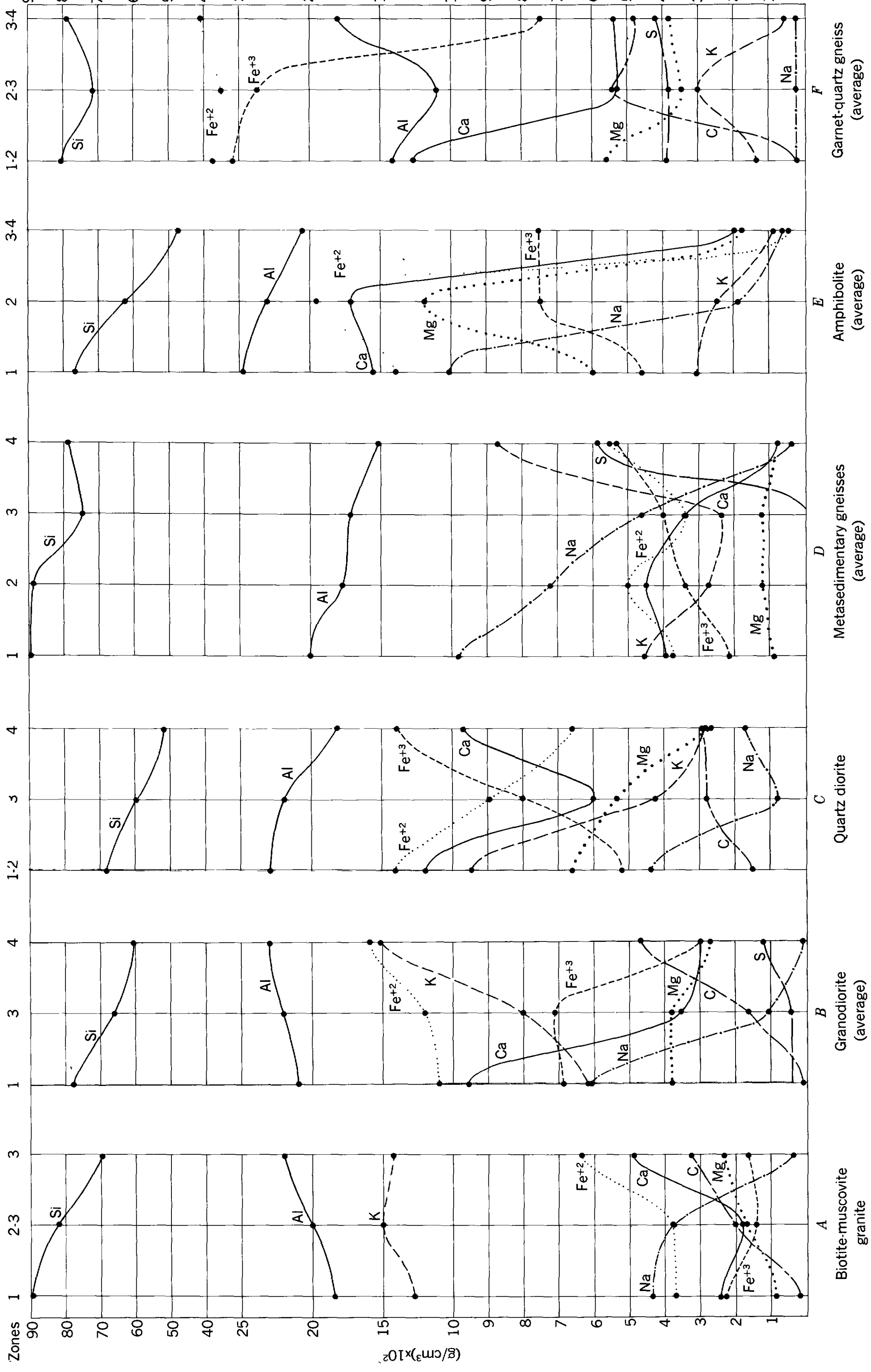

FIGURE 30.—Summary of ion distributions in altered wallrocks.

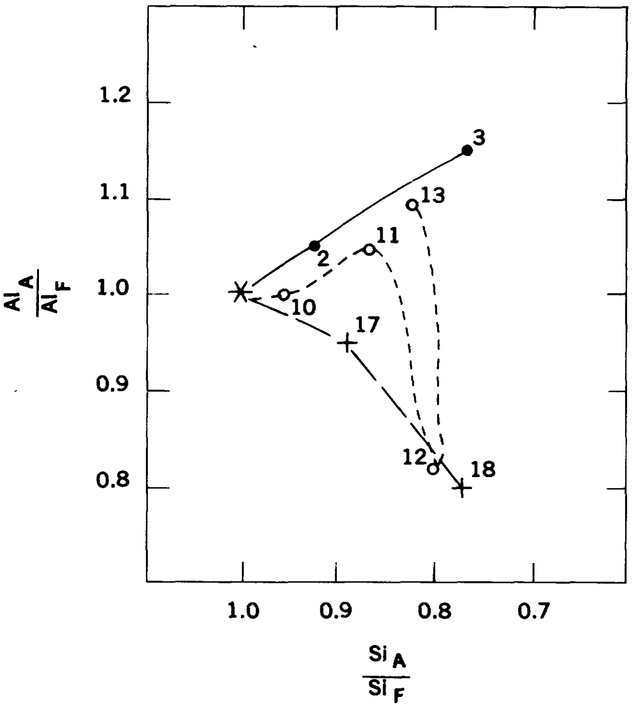

FIGURE 31.—Variations in the ratio of altered (A) to unaltered (F) aluminum ions in terms of the ratio of altered to unaltered silicon ions in granite (2–3), microcline-quartz-plagioclase-biotite gneiss (10–13), and quartz diorite (17–18). Asterisk denotes original host rock.

which averages 1.6 percent; however, potassium is more abundant than sodium in granite. These large ions (Na, 0.97A, and K, 1.33A) occur mostly in K-feldspar and sodic plagioclase (table 21) where they fill lattice sites in the octagonal chains of the three-dimensional silicon-oxygen framework. The micas and three-layer type clay minerals also may contain sodium and potassium (paragonite was not identified in the region), but these large ions fit only in interlayer positions. Normally sodium is eliminated from strongly argillized rock minerals (fig. 30); however, sodium is not displaced in certain metasedimentary gneisses until zone 4. The increase of sodium in intensely altered quartz diorite gneiss may be related to the presence of sodium-bearing mica or clay minerals. The loss of sodium in strongly altered rock follows the pattern of loss of sodium observed in weathered rocks and soils; the low ionic potential of sodium (Mason, 1952, p. 138) facilitates solution. Sodium, as expected, is a predominant constituent in hot-spring waters.

Potassium is enriched in K-feldspar-rich rocks (fig. 30) adjacent to the vein; the increase seems to be rerelated not only to the apparent stability of K-feldspar and biotite structures in the argillic zone, and of K-feldspar in zone 4, but also to the introduction of K to form sericite in zone 4. Thus very small loss of potassium (fig. 30) reflects the metastability of K-bearing minerals in zones 2 and 3, and the conversion of clay minerals to sericite coincident with introduction of potassium accounts for the marked increase of potassium in zone 4.

Potassium is removed from the argillized less abundant plagioclase-rich K-feldspar-poor rocks (fig. 30) in the area; introduction of K in zone 4 is noted in certain metasedimentary rocks whose mineralogical composition lies between the K-rich and K-poor rock types. This introduction of potassium is in accord with the observations of Lovering and others (1949, p. 45–47) at Tintic, Utah. The alteration of biotite in zone 3 in some of these rocks and the consequent loss of potassium before sericite could form in altered amphibolite, quartz diorite, and garnet-quartz gneiss (rocks which possess only a little sericite in a narrow zone 4) constitutes an extreme variation in potassium distribution. K^+ ions have a low ionic potential (Mason, 1952, p. 138) and, like Na^+ ions, should be lost by solution unless preferentially incorporated in illite and sericite. The ion size and charge distribution in K ions more closely satisfies the geometric space and charge requirements of the interlayer mica structure than those for Na or Ca ions (Grim, 1953, p. 146–147). Quartz diorite and garnet-quartz gneiss occur principally in the peripheral ore zone where a narrow sericitized zone is not unexpected; however, amphibolite occurs in all ore zones and consistently lacks a sericitized zone. Fournier (written communication, 1960) considers that the formation of sericite (fixation of K^+) is controlled in large part by the K^+/H^+ activity ratio. In rocks relatively devoid of K-bearing minerals the K^+/H^+ ratio perhaps was too low (even after K^+ introduction from without) to form large amounts of sericite. In contrast, in rocks containing large amounts of stable K-feldspar (having a finite solubility that may permit contribution of K ions into solution locally), the K^+/H^+ activity ratio was such that introduced K ions became fixed in sericite. This is the simplest view. Perhaps more significant, however, is the fact, suggested by Hemley (written communication, 1960), that the formation of the alteration assemblage depends on the relative activities of Ca, Na, Mg, and other ions as well as K in a complex system. If these activities are high enough, as they may well be in alteration of more basic rock, the stability field of the mica may be displaced by that of clay mineral or chlorite; and K might even show a net decrease in the rocks.

MAGNESIUM

Magnesium is present in fresh rocks in amounts generally less than 1 percent. Its ions are of medium size (0.66A) and are combined with oxygen in garnet, amphibole, and layer silicate structures. It occurs commonly in three-dimensional framework structures. The magnesium occurs in garnet in interstices of linked silicon tetrahedra and aluminum octahedra, in amphiboles as cation links between the double-chain structures, in the micas and certain three-layered clay minerals in octahedral coordination replacing aluminum, and in chlorite in the micalike portion as well as in brucitelike octahedral interlayer sheets.

The distribution of magnesium in altered wallrocks is variable, but in general parallels the distribution of ferrous iron (fig. 30). The exception to this general Mg-Fe^{2+} relation in granodiorite and garnet-quartz gneiss is caused by the formation of pyrite in zone 4 rock. In general there is an increase in magnesium as K-feldspar-rich rocks are argillized, and a decrease as plagioclase-rich rocks are argillized (fig. 32); quartz diorite and amphibolite also contain more Mg than granite or granodiorite, and thus contain more Mg ions in excess of those required in the secondary clay-mineral structures.

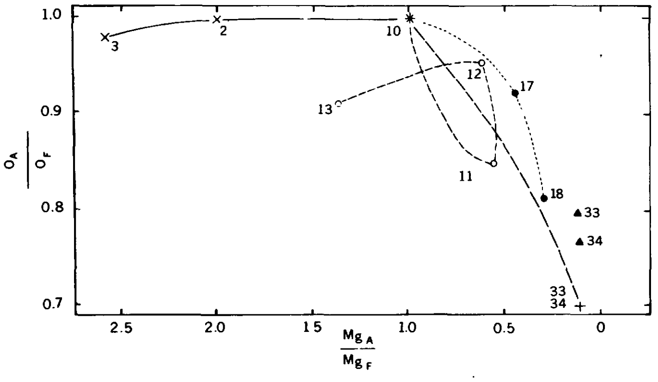

FIGURE 32.—Variations in the ratio of altered (A) to unaltered (F) magnesium ions in terms of the ratio of altered to unaltered oxygen ions in granite (2–3), microcline-quartz-plagioclase-biotite gneiss (10–13), quartz diorite (17–18), and amphibolite (33–34). Asterisk denotes original host rock; ▲ 33, 34 plots consider OH as O ions. Note exaggeration in vertical scale.

Magnesium distribution is most closely correlated with biotite and clay-mineral occurrences; magnesium is separated slowly from biotite in the outer zones of altered rock and remains in the resulting secondary clay-mineral lattices in chlorite and (or) illite. Biotite is altered along cleavage planes coincident with but at a slower rate than is the plagioclase in plagioclase-rich rocks. Closer to the vein, where clays and the biotite are altered, magnesium is released and migrates veinward. Magnesium is less soluble in aqueous solutions than Fe^{+2}, Ca, Ba, Na, and K, and Mg^{+2} ions are more strongly bonded in silicate structures than Fe^{+2}.

CALCIUM AND BARIUM

Calcium and barium are closely related in chemical properties, but their abundance and manner of occurrence in these altered rocks differ markedly. Calcium makes up from less than 1 to 4.5 percent of the fresh wallrocks, whereas barium is present in amounts less than 1 percent and was determined (by chemical analysis) in only a few of the rocks studied. Calcium, ionic radius 0.99A, occurs in many silicate structures; it is most prominent in plagioclase in interstices of octagonal chains of silica and alumina tetrahedra. Calcium also occurs in garnet, pyroxene, amphibole, and more rarely, in mica structures in positions similar to those described for magnesium ions. Montmorillonite most commonly contains calcium as an adsorbed interlayer ion, and calcium also is found in wallrocks in the structures of calcite, fluorite, and apatite. Barium may replace potassium in the feldspar structure and also in muscovite and biotite. Barite generally occurs in the vein and perhaps in the altered rocks near it, but has not been identified by the writer in thin sections of altered rock.

The distribution of calcium is closely related to the breakdown of plagioclase and hornblende into calcium-montmorillonite, and, following the alteration of the montmorillonite into kaolinite and sericite near the vein, calcium is available to form calcite. Apatite is stable throughout most altered rock. As most rocks are argillized they lose calcium (fig. 30). In general the amount of calcium retained in the montmorillonite structure (table 21) is less than that permitted in the plagioclase varieties commonly found in these rocks. Calcium is somewhat soluble in water, and the ions are free to migrate veinward to form calcite if fixed by CO_2 (as shown in fig. 30), or they are lost from the local environment and fixed elsewhere in the wallrocks or in the vein fissure as was observed in veins cutting garnet-quartz gneiss in the Golconda mine.

Barium content in these rocks is small (table 18), its ionic size is large (1.34A), and it is consistently more common in rocks with K-feldspar and mica structures and is less common in argillized rock. This seemingly reflects the stability of K-feldspar and biotite through zone 3–4, and suggests that these minerals are the host structures for barium disseminated in these rocks.

IRON

Iron is most abundant in garnet-quartz gneiss (19 percent), amphibolite (9 percent), and granodiorite and quartz diorite (6–7 percent), and occurs in the ferrous and ferric states (Fe^{+2}, 0.74A, and Fe^{+3}, 0.64A). The bulk of iron in wallrocks occurs in the ferrous state as a replacement of octahedral Mg or Al ions in layer and chain silicate structures, and as links between double chain structures. Iron occurs also in magnetite structure in chains of tetrahedral and octahedral Fe^{+2}-O and Fe^{+3}-O groups, in the hematite and ilmenite structure as ferric iron octahedrally coordinated within

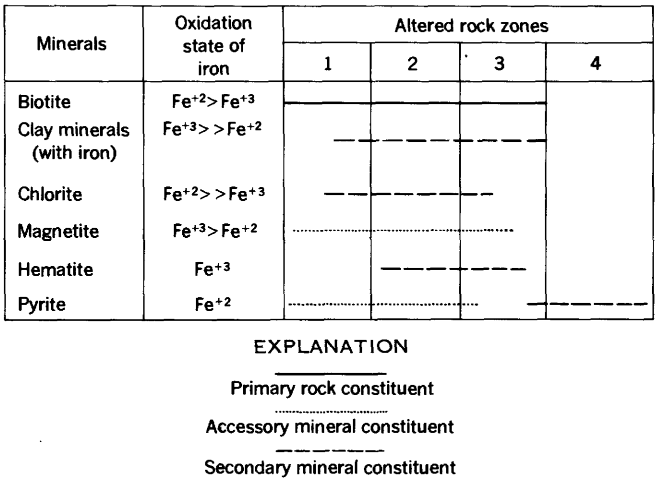

FIGURE 33.—Diagrammatic sketch of stability ranges and oxidation states of iron-bearing minerals in fresh and altered biotite-quartz-plagioclase gneiss, Essex mine.

hexagonally close-packed oxygen layers, and in pyrite as ferrous iron combined with S in a sodium chloride-type structure.

The relative proportion of iron in varied oxidation states in several minerals common to these rocks (fig. 33) illustrates one facet of the complex problem of accounting for iron distribution. The proportion of iron in equal volumes of iron minerals varies (table 22)—pyrite requires less iron per unit volume than iron oxide minerals. Graphical plots (fig. 34) reveal a nearly regular pattern of iron distribution in altered rocks, but these diagrams fail to account for actual amounts of iron in each zone. These data do not provide information on relative areas (or volumes) of altered rock zones that are needed to demonstrate whether a balance exists between the iron lost in zone 2 and that deposited in zones 3 and 4. We can generalize from laboratory

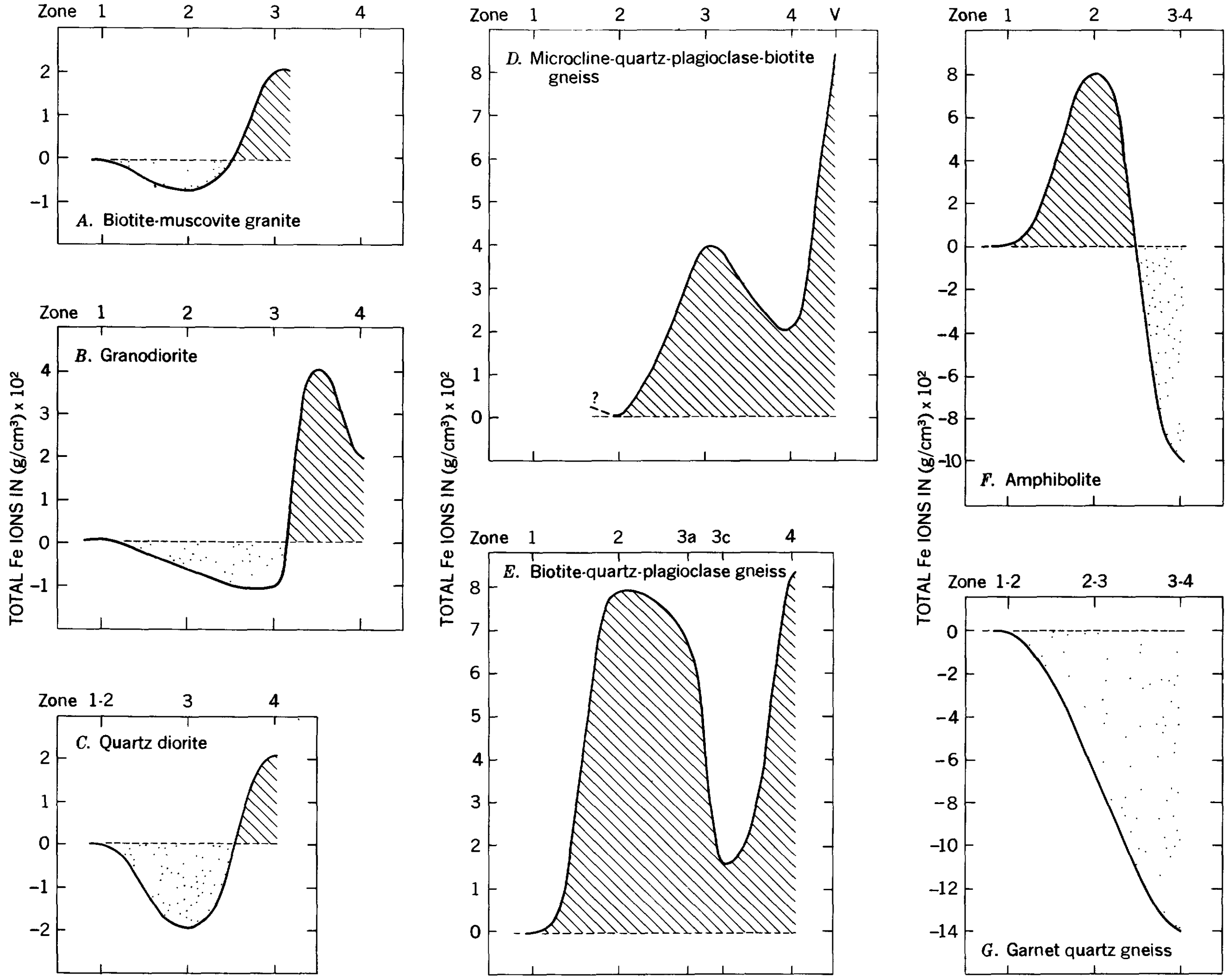

FIGURE 34.—Distribution plots of total iron in altered wallrocks. Values greater than zero, the starting material, indicate gain of Fe ions, values less than zero indicate loss of Fe ions.

TABLE 22.—*Distribution of constituent ions in common iron-bearing minerals*

[Data from Palache, Berman, and Frondel (1944)]

Mineral	Fe^{+2}	Fe^{+3}	Total Fe (g per cu cm) $\times 10^2$	O	S
Magnetite	1.24	2.50	3.74	1.45	----------
Hematite	----------	3.64	3.64	1.56	----------
Pyrite	2.33	----------	2.33	----------	2.67

and field observations only that a relatively small amount of iron is removed from zone 2 rock in comparison with the greater amount of iron deposited in zones 3 and 4. The area (and therefore volume) of zone 2 rock bordering a normal vein, however, is proportionally greater than that of zones 3 and 4 rock. Biotite is believed to be the major source of movable iron in most rocks in the region, but hornblende is a significant source in a few rocks. Clay minerals formed from relatively iron-free plagioclase adsorb or incorporate iron readily, and near the vein the iron combines with S to form pyrite.

In K-feldspar-rich rocks such as granite, granodiorite, and possibly in microcline-quartz-plagioclase-biotite gneiss there is a very small initial loss of iron in zone 2, a moderate concentration in zone 3, and a slight loss in zone 4. Biotite is unaltered through zone 3, thus losses of iron in zone 2 are attributed to accessory minerals, and gains in zone 3 may indicate movement of iron from zone 4.

In plagioclase-rich rocks, such a amphibolite, quartz diorite, and biotite-quartz-plagioclase gneiss, a loss of iron occurs in zones 1–2, a substantial gain occurs in zone 3, and a slight loss is noted in zone 3–4. Biotite and hornblende are more abundant in many plagioclase-rich rocks than in K-feldspar-rich rock in the area, and the argillic zone is broader in these rocks. Iron is freed from hornblende (zone 2–3) and biotite (zone 3) and is held in clay-mineral structures; however, additional iron must be moved from zones 2 and 4 or introduced from the vein(?) to account for the marked gain of iron in the argillic zone. The distribution of iron in biotite-quartz-plagioclase gneiss (fig. 34) is complicated by the variability of the biotite content in this layered rock; and the total iron in zone 1 may be low owing to inclusion in the sample of rock with less than average biotite content.

The increase in ferrous and loss of ferric iron in strongly altered garnet-quartz gneiss reflects reduction of iron to form abundant disseminated pyrite; however, there is a loss of total iron close to the vein which when added to vein solutions was available for use elsewhere. Amphibolite also contributed excess iron to the vein environment. The sequential oxidation and rearrangement of ferrous to ferric iron in the replacement of magnetite by hematite, and the subsequent reduction and rearrangement of ferric iron with S to form pyrite, involves a decrease of iron per unit volume (table 22). Thus the replacement of a magnetite grain by pyrite involves a decrease in iron per unit volume that becomes available for combination elsewhere.

MANGANESE

Manganese (Mn^{+2}, 0.80A) is present in most rocks of the area in amounts of less than 1 percent, but is more abundant in garnet-quartz gneiss (2.7 percent). The element occurs mostly as a replacement of iron, magnesium, and calcium in mica, amphibole, clay minerals, garnet, magnetite-ilmenite, and apatite; in the veins it occurs in sphalerite as well as in manganese oxides or carbonates. The uncommonly high amount of manganese in altered garnet-quartz gneiss is contained mostly in the garnet structure. Garnet is gradually altered in zone 3, and analyses (table 18) show that manganese is not removed abruptly from the rock environment in zone 3–4.

CARBON

Carbon is present in minerals with CO_3 groups in certain silicate and phosphate minerals, but mostly as independent carbonate minerals in these altered rocks. As shown in figure 30 the carbon content of these rocks increases toward the vein. Chemical as well as petrographic relations indicate that CO_2 must be introduced by the altering solutions. These CO_2-bearing solutions and released calcium, iron, and manganese ions combine to form the secondary carbonate minerals sparsely disseminated in the wallrocks and more abundantly filling the veins.

PHOSPHORUS AND TITANIUM

Phosphorus and titanium occur in amounts of 1 percent or less; the concentration of phosphorus is below spectrographic sensitivity. P ions (0.35A) occur mostly in apatite in tetrahedral arrangement with oxygen, but also may substitute for Si ions in garnet and zircon. P. K. Sims (oral communication, 1957) reports that biotite-quartz-plagioclase gneiss locally contains several percent of monazite and xenotime, minerals having zircon-type structures that are possible sites for P in these rocks. Titanium is found mostly in ilmenite, sphene, rutile, and leucoxene, and is incorporated also in magnetite, replacing iron, and in biotite and amphibole replacing Si, Al, Fe, and Mg. The distribution of phosphorus in altered rocks is not well known but generally remains consistent with the distribution of apatite, which persists unchanged even in the zones of intense alteration. Titanium remains constant across altered rock zones until lost in the zone of most intense alteration; thus the distribution pattern is consistent with the stability of magnetite-ilmenite, sphene, and leucoxene.

ASSOCIATED ANIONS

The associated anions include oxygen, sulfur, fluorine, and hydrogen. Oxygen (1.40A) is most abundant in weight percent as well as by volume (table 21). The weight distribution of oxygen in fresh and altered rocks is reasonably constant if one omits consideration of water lost below 100° C (H_2O-); however, if all water is considered (table 23) there is a slight gain of oxygen in argillized rock and a slight loss in sericitized rock. If equivalent volumes of rock are considered, the progressive alteration of wallrocks results in only small gains and losses of oxygen. The addition of H^+ ions to oxygen to form water involves no significant volume increase, and in essence there is no real loss of oxygen in the addition of H^+ to form structural hydroxyl groups shown in figure 30. There is, however, the possibility that variations in the O^{16}/O^{18} ratio occur in progressively more altered wallrock. Oxygen is present in excess in the hydrothermal environment and substitutions are possible. These factors do not negate the space relations that are of prime concern here.

TABLE 23.—*Oxygen in fresh and altered wallrocks by weight percent*

Sample	Altered rock zone	Weight percent
Biotite-muscovite granite		
1	1	48. 55
2	2–3	48. 89
3	3b	49. 14
Granodiorite		
4	1	45. 65
5	3a	47. 29
6	3b	48. 74
7	3–4	47. 75
8	4	47. 56
Microcline-quartz-plagioclase-biotite gneiss		
9	2	54. 0
10	3	50. 4
11	4	50. 5
12	4–V	49. 3
13	V	45. 6
Quartz-diorite		
16	1–2	47. 28
17	3	49. 59
18	4(?)	49. 31
Biotite-quartz-plagioclase gneiss		
19	1	49. 05
20	2	48. 51
21	3	49. 55
22	3–4	50. 38
23	V	47. 69
Biotite-quartz-plagioclase gneiss		
24	1–2	49. 77
25	3a	50. 50
26	3b	53. 29
Amphibolite		
32	2	45. 90
33	3	54. 43
34	3–4	46. 29
Garnet-quartz gneiss		
35	1–2	43. 36
37	2–3	45. 32
39	3–4	42. 31

Very few sulfide minerals are disseminated in the wallrocks, and only in intensely altered rock is the amount of S sufficient to be noted in analyses. Pyrite is the chief sulfur-bearing mineral and generally is texturally related to altered biotite or magnetite (fig. 7).

Fluorine occurs mostly in apatite and fluorite where it is retained locally even in intensely altered wallrock in the Banta Hill mine area. F^- (1.36A) diadochically replaces $(OH)^-$ (about 1.40A), CO_3, and Cl^- in mica, amphibole, and clay-mineral structures. Fluorite and apatite resist solution, but the solubility of CaF_2 increases with the content of CO_2 in water (Rankama and Sahama, 1950, p. 763).

TRACE ELEMENT CONSTITUENTS

Trace elements [5] occur as the primary constituents in accessory minerals or in very small amounts in essential minerals by replacement (camouflage, capture, or admission) of more common elements. Thus while the mode of occurrence of trace elements in minerals is not well known, the affinity of certain elements for one another (Rankama and Sahama, 1950, p. 125; Higazy, 1954), and therefore for certain minerals is marked. The distribution problem is twofold: Trace elements in altered rocks may be either redistributed from the fresh rock or may be introduced by the altering solution.

The semiquantitative spectrographic data are not statistically rigorous and are used only to indicate general trends of distibution of trace elements in altered wallrocks. The data were not obtained from statistically homogeneous samples, and neither the precision of the analyses nor the small numbers of samples analyzed permit a statistically small confidence interval for the values cited. It is assumed that the general and consistent distribution patterns of elements indicated by these data are real. Comparisons of trace elements in these rocks with similar constituents in related rocks are made but are subject to the limitations of these analytical data.

METHOD FOR PRESENTING DATA

Spectrographic analyses were made by the semiquantitative method developed in the Denver and Washington laboratories of the Geological Survey; [6] these analyses are presented graphically to illustrate the relations between them more easily. The value, plotted as X—looked for, but not detected—is plotted

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[5] Trace elements here include all elements in the rock determined by spectrographic methods except those reported by standard chemical analysis.

[6] The results of spectrographic analysis are reported in ranges, each power of ten being divided into three ranges of about 1, 2, 5, 10, etc. The approximate mid-points of the ranges, 1.5, 3, and 7 are the figures used in reporting results. These figures represent the amount of the element in the sample that may be expected to be between 1.0 and 2.1, between 2.1 and 4.6, etc. In such a series of reported results at least 60 percent are expected to lie within the correct range. It is believed that about 20 percent of the results will be one range too high and 20 percent one range too low (A. T. Meyers, written communication, 1958).

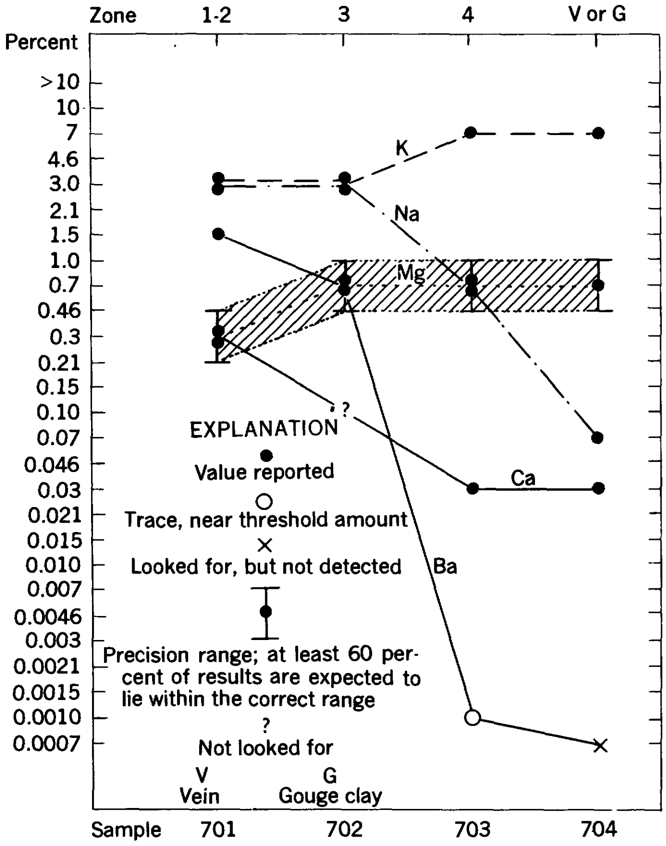

FIGURE 35.—Sketch illustrating conventions and notations used in plotting the semiquantitative spectrographic data.

TABLE 24.—*Standard sensitivities, in percent, of elements determined by semiquantitative spectrographic methods at U.S. Geological Survey laboratories*

[Denver, sensitivity as of August 1954, amended: Washington, sensitivity as of March 1956. Dashes indicate element not reported. A second exposure was required for the higher sensitivities shown in parentheses]

Element	Denver	Washington	Element	Denver	Washington
Ag	0.0001	0.00001	Na	0.05 (0.0005)	0.01 (0.0003)
Al	.001	.0001	Nb	.001	.001
As	.05	.01	Nd	.01	.006
Au	.003	.001	Ni	.0005	.001
B	.002	.005	Os	.005	.1
Ba	.0001	.0005	P	.3	.07
Be	.0001	.00005	Pb	.001	.001
Bi	.001	.005	Pd	.0005	.003
Ca	.001	.01	Pr	------------	.01
Cd	.005	.005	Pt	.003	.003
Ce	.01	.03	Rb	---- (0.01)	7. (0.007)
Co	.0005	.001	Re	.005	.04
Cr	.0001	.0006	Rh	.005	.004
Cs	------------	.8 (0.01)	Ru	.005	.008
Cu	.00005	.00005	Sb	.01	.01
Dy	.05	.006	Sc	.001	.0005
Er	.005	.003	Si	.001	.005
Eu	------------	.003	Sm	.01	.008
F	------------	[1].08	Sn	.001	.001
Fe	.001	.0008	Sr	.0001	.001
Ga	.0005	.001	Ta	.05	.1
Gd	.005	.006	Tb	------------	.01
Ge	.0005	.001	Te	.08	.08
Hf	.05	.007	Th	.05	.05
Hg	.01-0.1	.08	Ti	.0005	.0005
Ho	------------	.001	Tl	.01	.04
In	.001	.0004	Tm	------------	.001
Ir	.005	.03	U	.05	.08
K	.5 (0.001)	.1 (0.005)	V	.001	.001
La	.002	.003	W	.01	.05
Li	.02 (0.0001)	.01 (0.00003)	Y	.001	.001
Lu	------------	.005	Yb	.0001	.0001
Mg	.001	.00003	Zn	.02	.008
Mo	.001	.0005	Zr	.001	.0008
Mn	.0005	.0007			

[1] A separate exposure is required for the F estimation.

at the nearest unit below its standard sensitivity. The elements listed in boxes in diagrams also were looked for but not found in the sample. Other notations used in the graphical presentation of the data are given in figure 35. The standard sensitivities for elements determined by the semiquantitative method are presented in table 24. The sensitivities obtained in the Washington and Denver laboratories vary slightly, and the appropriate sensitivity has been used in the graphic-data plots. In some individual analyses greater sensitivities than those in table 24 were possible and and the values reported have been plotted.

To facilitate graphing these data, the elements have been arbitrarily grouped into four sections based on the arrangement illustrated by Barbor (1944, p. 25): *a*, The alkali and alkaline earth (regular) elements; *b*, regular elements; *c*, transitional metals, first group; and *d*, transitional metals of the second and third groups and secondary transitional metals.[7] Semiquantitative spectrographic data for 112 samples of fresh and altered wallrocks are reported in figures 36–62; elements which were looked for, but not detected, also are listed. The resulting graphs show the distribution of elements from fresh rock to the vein; on some graphs the distance between the samples can be measured by reference to the scale shown on the figure; on others, no scale is shown as the distance was not measured, and the samples have been plotted without reference to true relative distance between them.

Average values for some Front Range rocks (granite, microcline-quartz-plagioclase-biotite gneiss, granodiorite, biotite-quartz-plagioclase gneiss, metasedimentary and migmatite gneisses, and quartz diorite) are shown in figures 63, 64, and 65. A comparison of these data with the average abundance of certain elements in igneous rocks is afforded by table 25. Trace elements of groups *b* and *d* in rocks of the Front Range mineral belt are slightly higher than the average. In fact the average values given by Rankama and Sahama (1950) mostly are below the analytical sensitivity obtained in these analyses. Trace elements in group *c*, however, are equal to or lower than average values. Elements of group *a* are mostly major constituents.

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[7] Regular elements are those with the distinguishing electron in the outermost electron shell; "transitional" metals have the distinguishing electron in the next outermost shell; "secondary transitional" metals (including rare earths) have the distinguishing electron in the third from outermost electron shell.

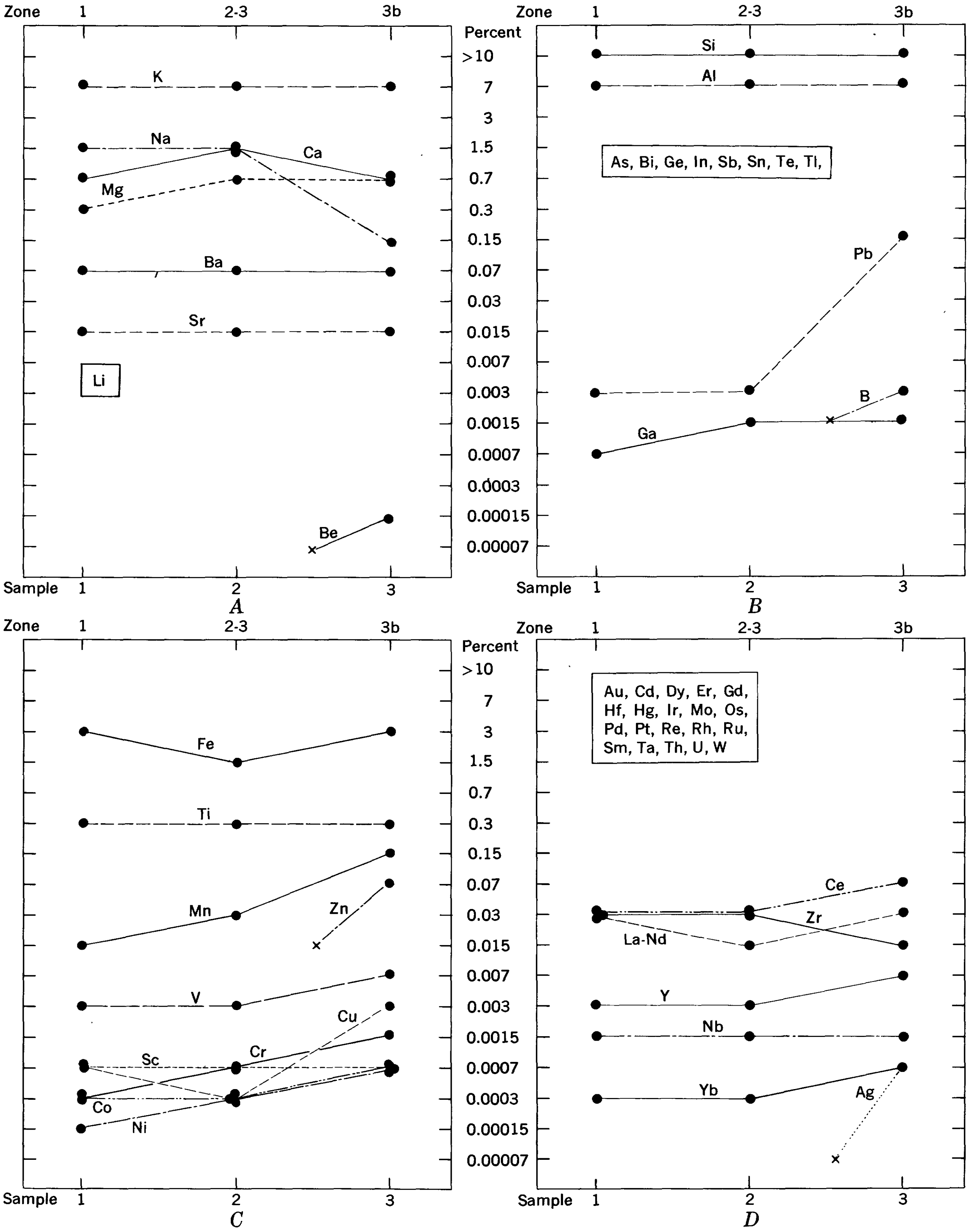
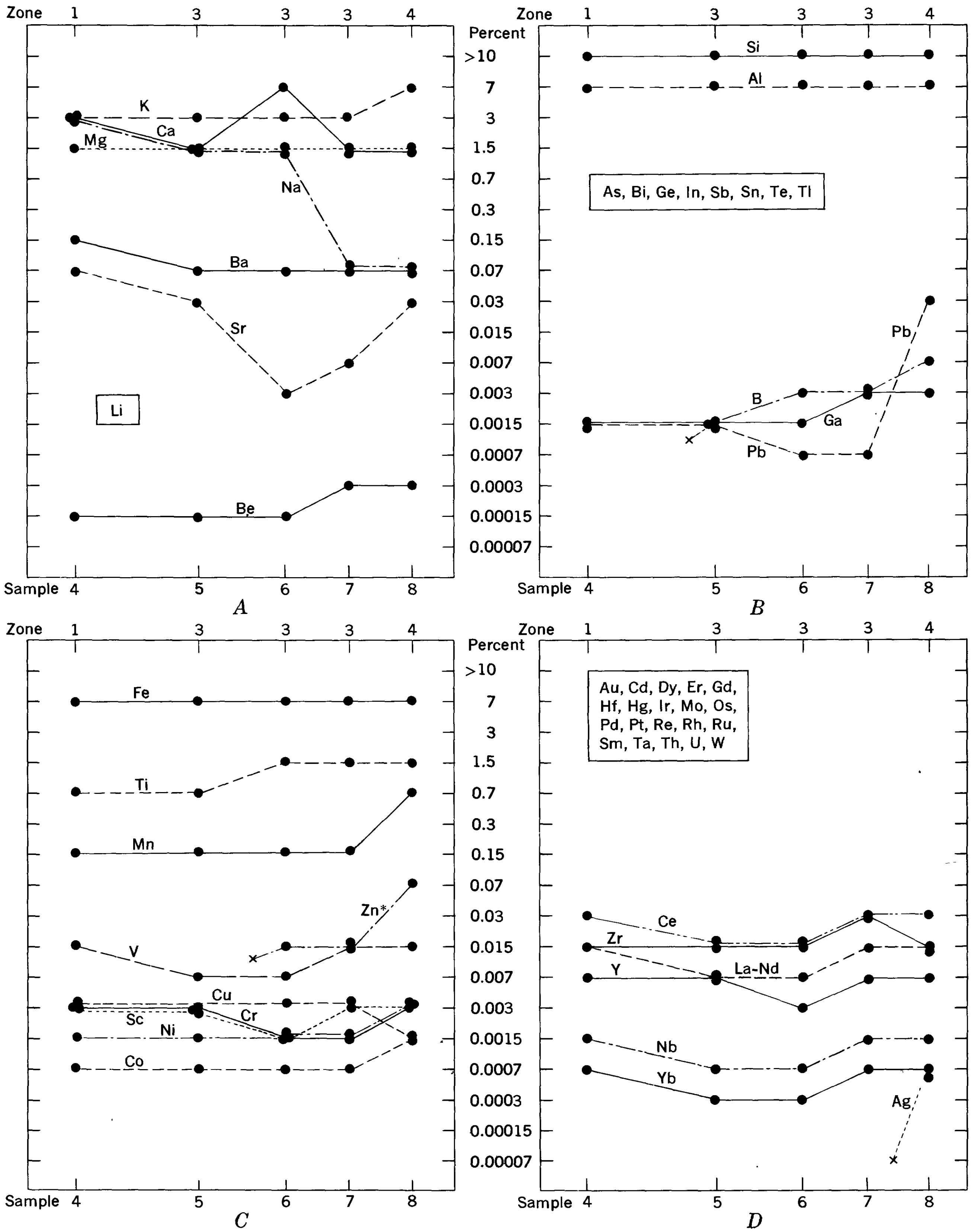


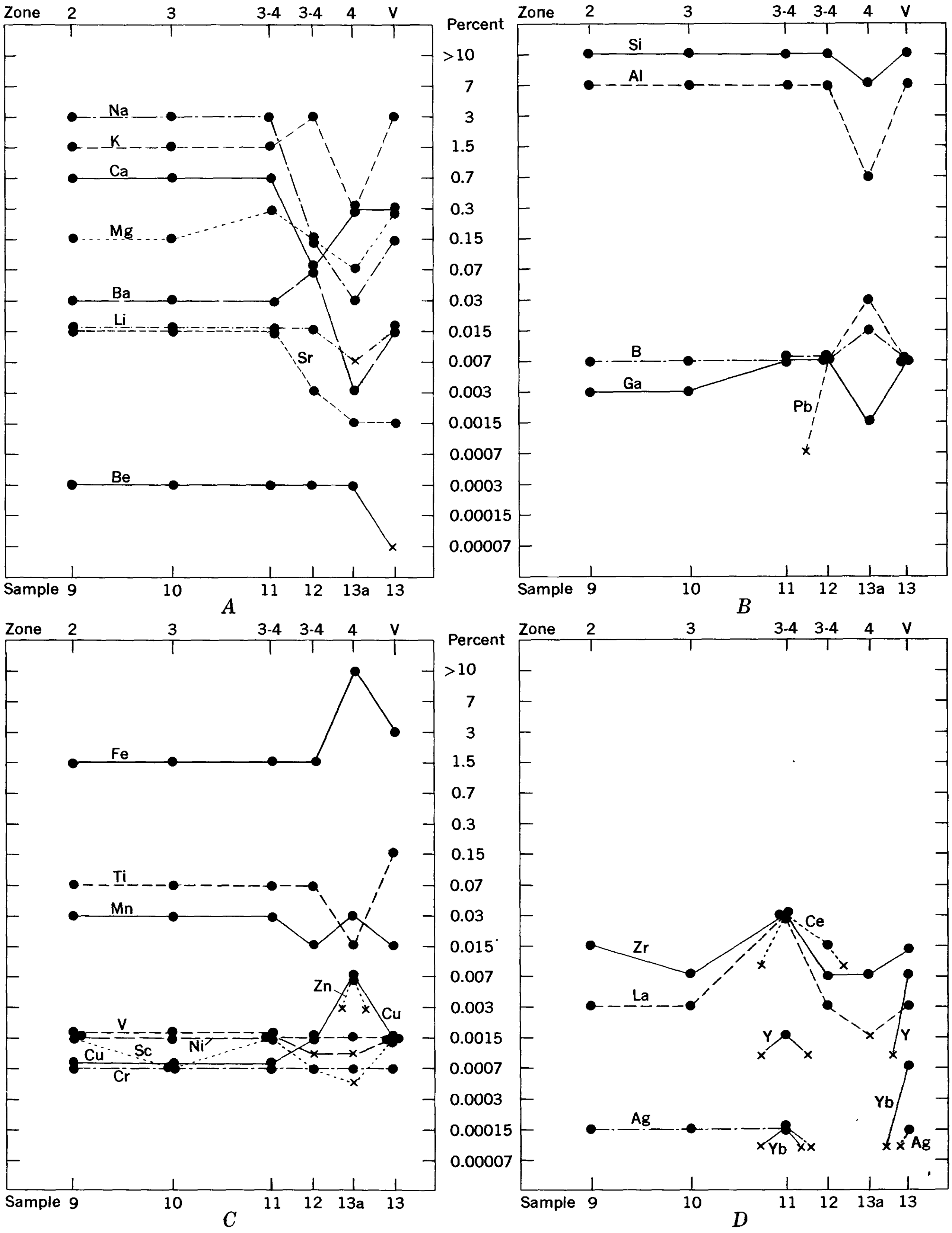
FIGURE 36.—Semiquantitative spectrographic analyses of fresh and progressively more altered biotite-muscovite granite, Nabob and Jo Reynolds mines, Lawson-Dumont-Fall River district, Clear Creek County, Colorado. Analyst: P.R. Barnett.



*Special sensitivity for Zn in these samples

Analyst: P. R. Barnett, Denver

FIGURE 37.—Semiquantitative spectrographic analyses of fresh and progressively more altered granodiorite, M. and M.-Dixie mine, Chicago Creek area, Clear Creek County, Colorado. Analyst: P. R. Barnett.



Analyst: K. E. Valentine, Washington, D. C.

FIGURE 38.—Semiquantitative spectrographic analyses of progressively more altered microcline-quartz-plagioclase-biotite gneiss, E. Calhoun mine, Central City district, Gilpin County, Colorado. Analyst: K. E. Valentine.

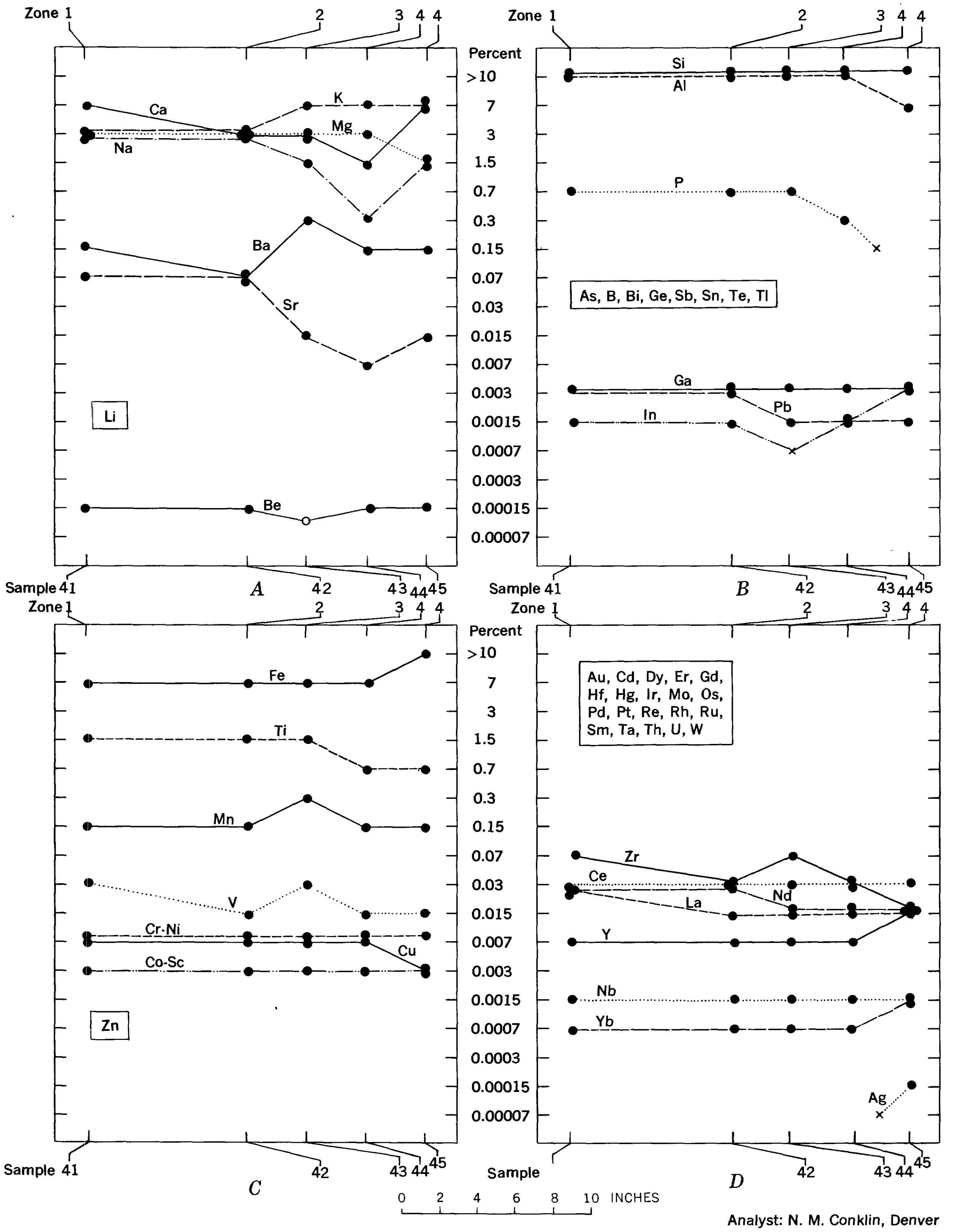


FIGURE 47.—Semiquantitative spectrographic analyses of fresh and progressively more altered granodiorite, Hayes and Wheeler tunnel, Central City district, Gilpin County Colorado. Analyst: N. M. Conklin.

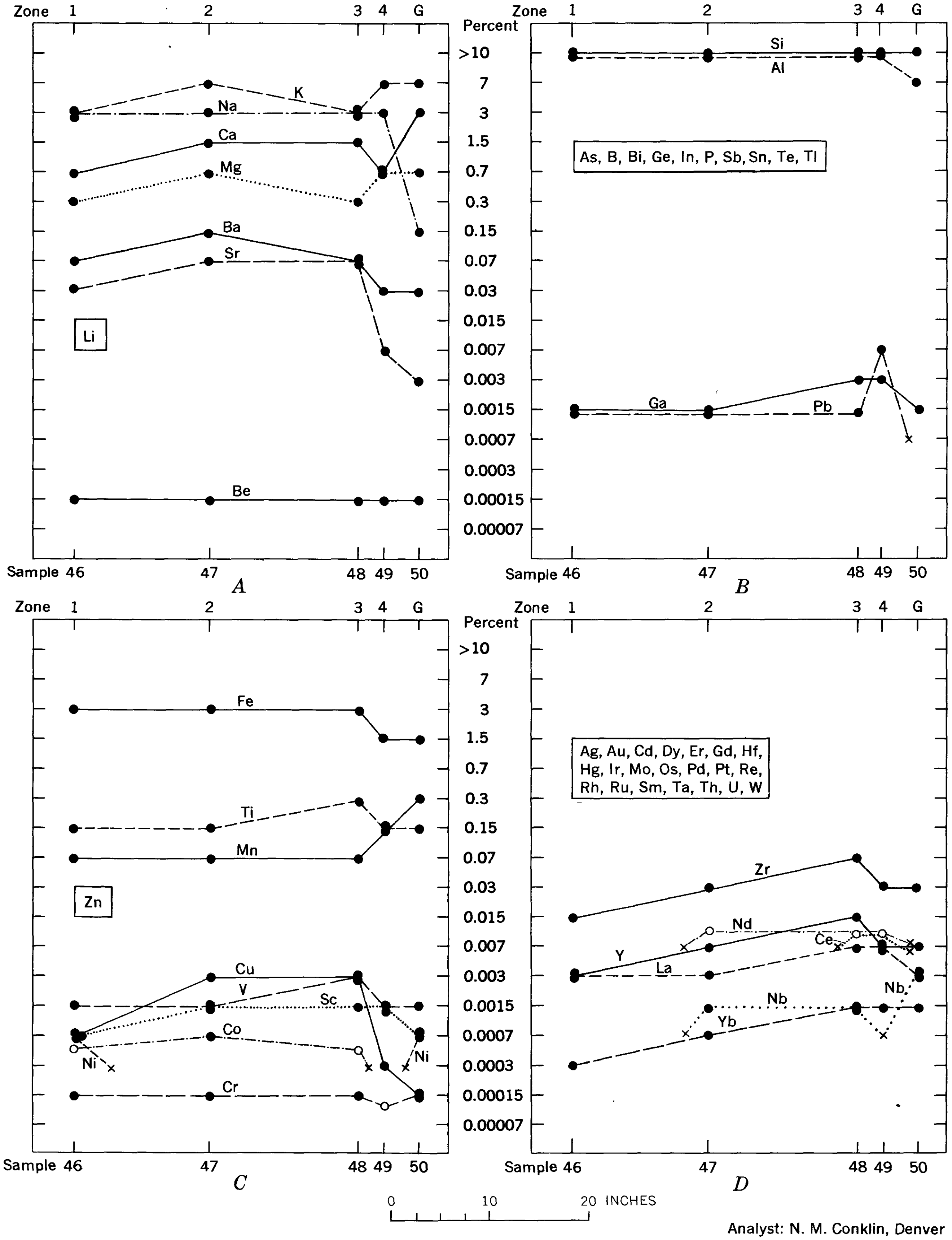


FIGURE 48.—Semiquantitative spectrographic analyses of fresh and progressively more altered microcline-quartz-plagioclase-biotite gneiss, Essex mine, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.

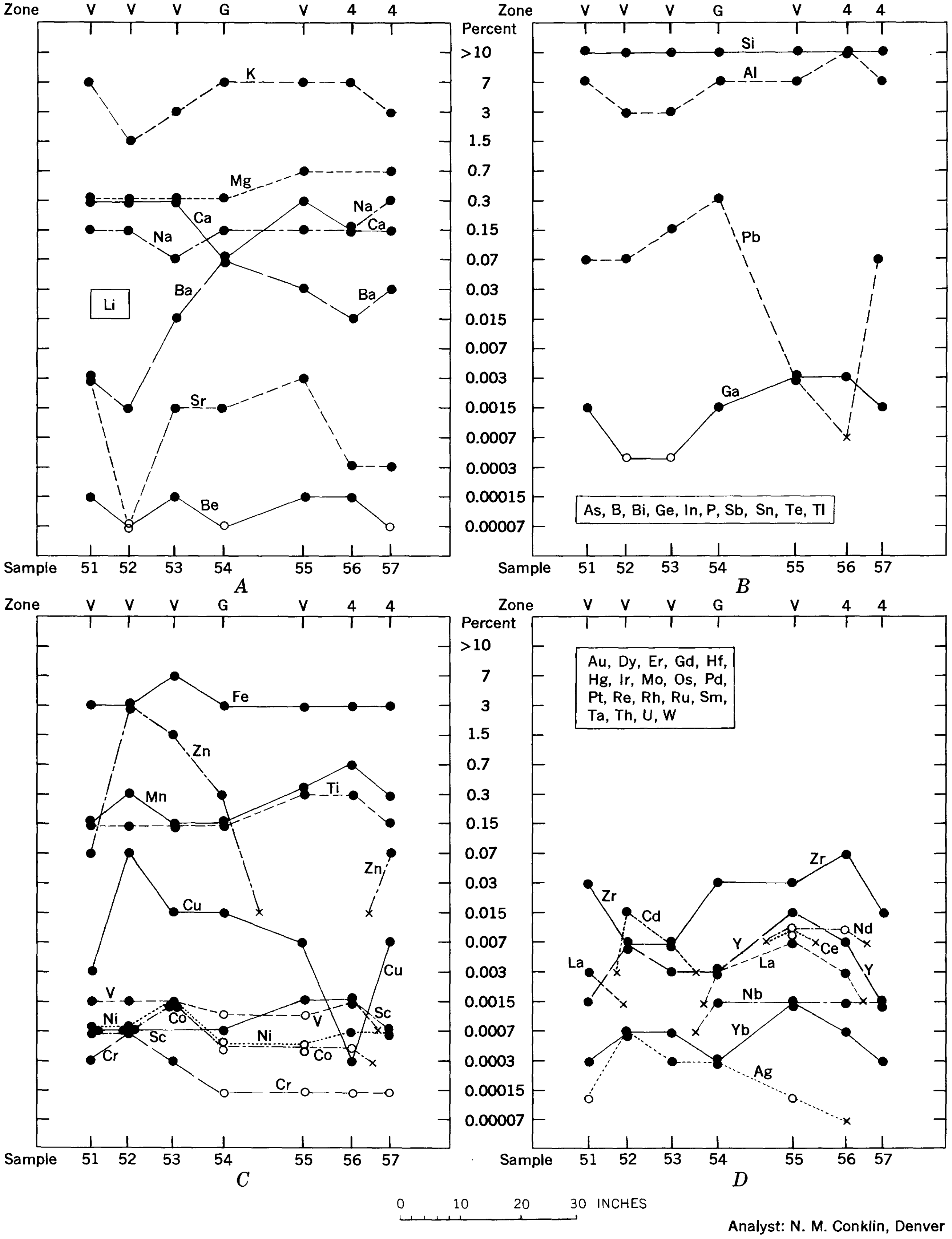


FIGURE 49.—Semiquantitative spectrographic analyses of progressively more altered microcline-quartz-plagioclase-biotite gneiss, Essex mine, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.

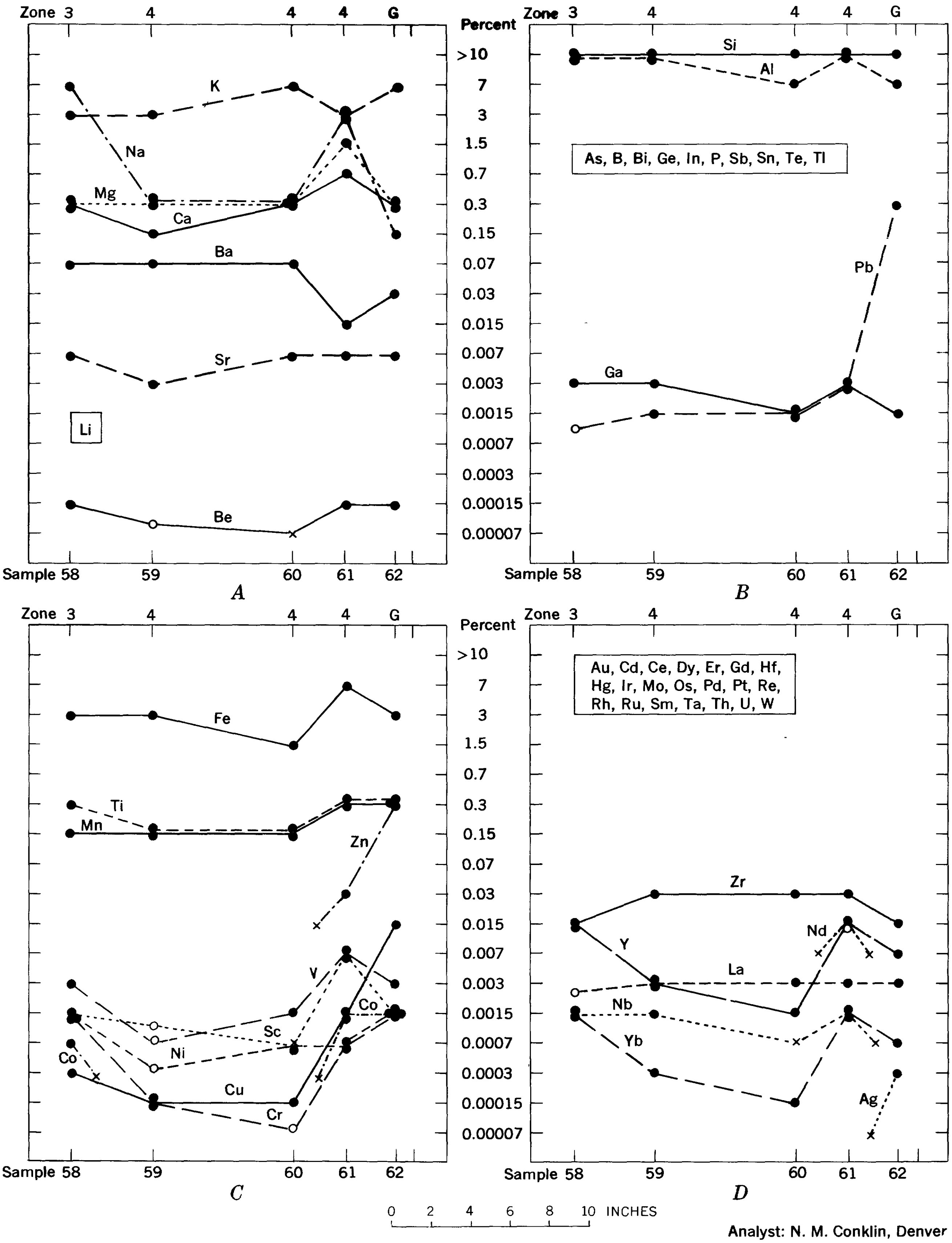


FIGURE 50.—Semiquantitative spectrographic analyses of progressively more altered microcline-quartz-plagioclase-biotite gneiss, Essex mine, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.

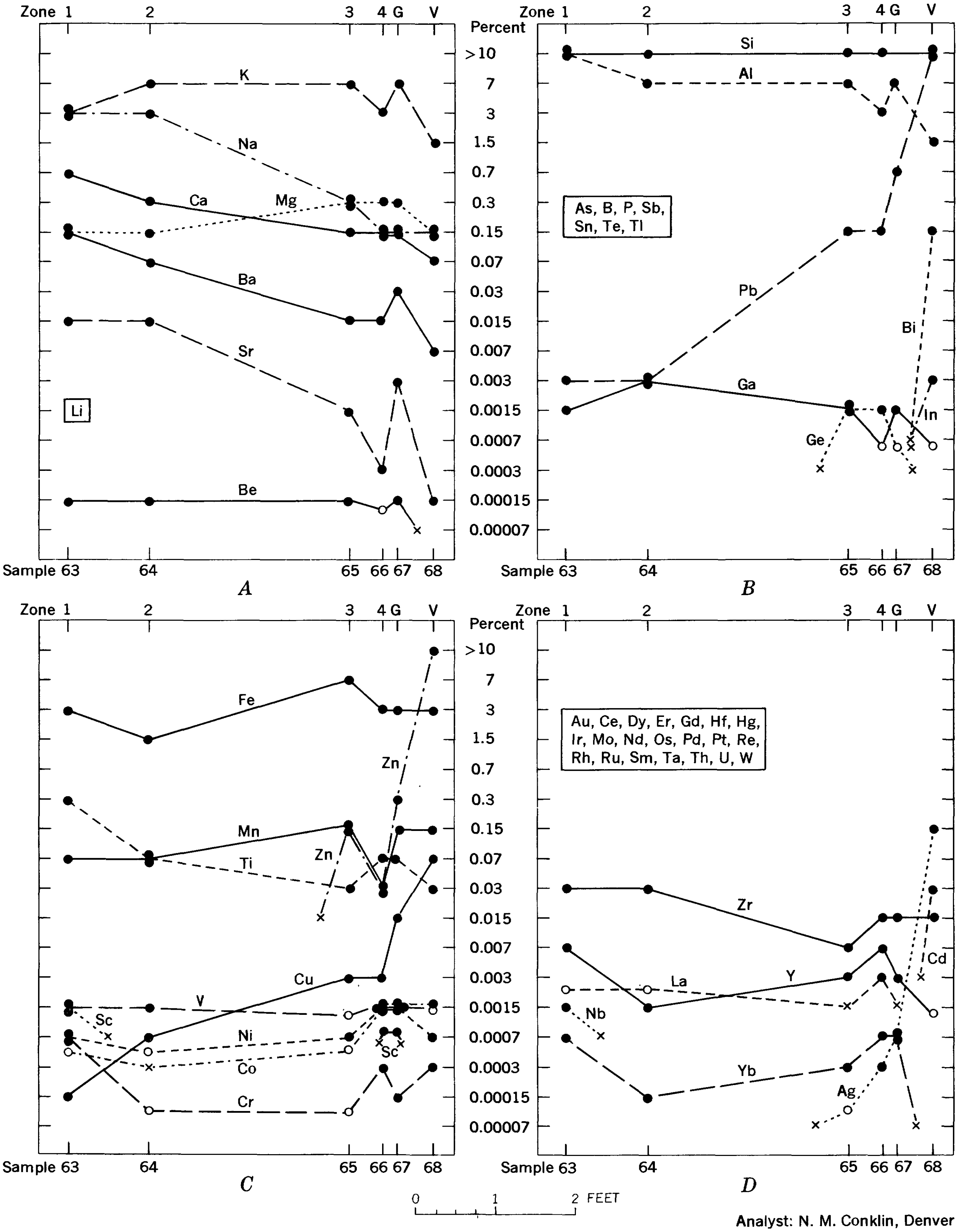


FIGURE 51.—Semiquantitative spectrographic analyses of fresh and progressively more altered microcline-quartz-plagioclase-biotite gneiss, Essex mine, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.

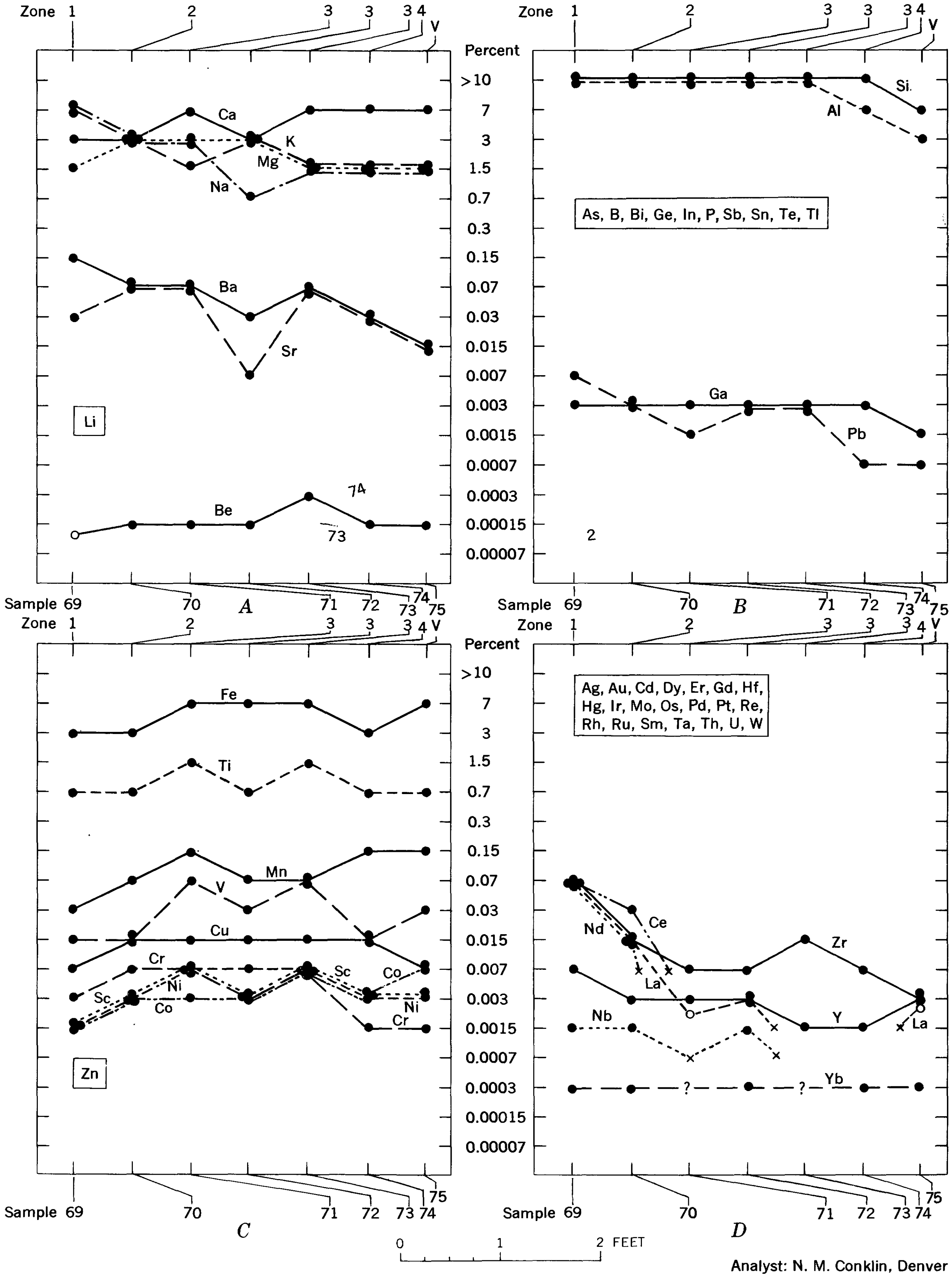
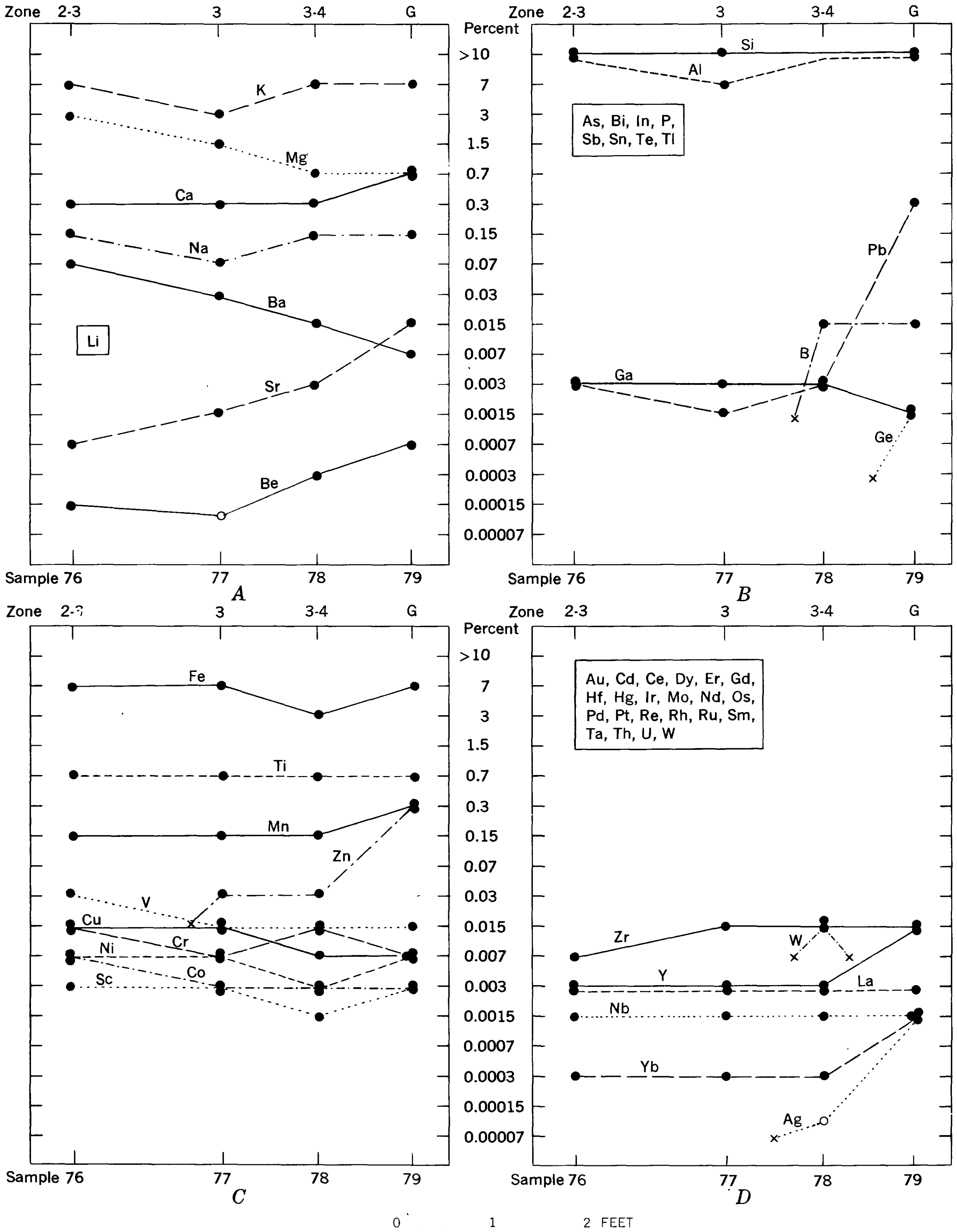
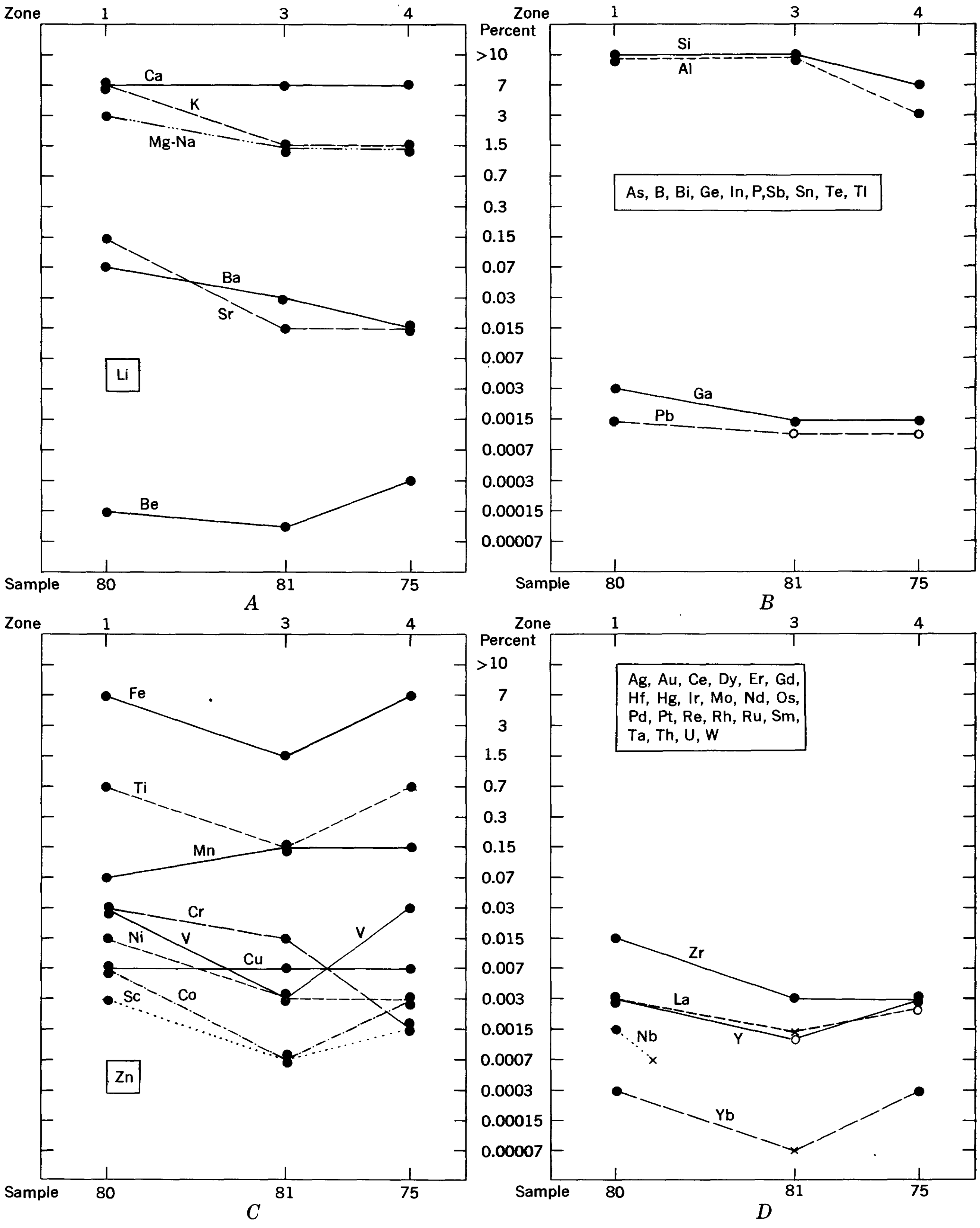


FIGURE 52.—Semiquantitative spectrographic analyses of fresh and progressively more altered quartz diorite, Jo Reynolds mine, Lawson-Dumont-Fall River district, Clear Creek County, Colorado. Analyst: N. M. Conklin.



Analyst: N. M. Conklin, Denver

FIGURE 53.—Semiquantitative spectrographic analyses of progressively more altered quartz diorite, Nabob mine, Lawson-Dumont-Fall River district, Clear Creek County, Colorado. Analyst: N. M. Conklin.



Analyst: N. M. Conklin, Denver

FIGURE 54.—Semiquantitative spectrographic analyses of fresh and progressively more altered quartz diorite, Jo Reynolds mine, Lawson-Dumont-Fall River district, Clear Creek County, Colorado. Analyst: N. M. Conklin.

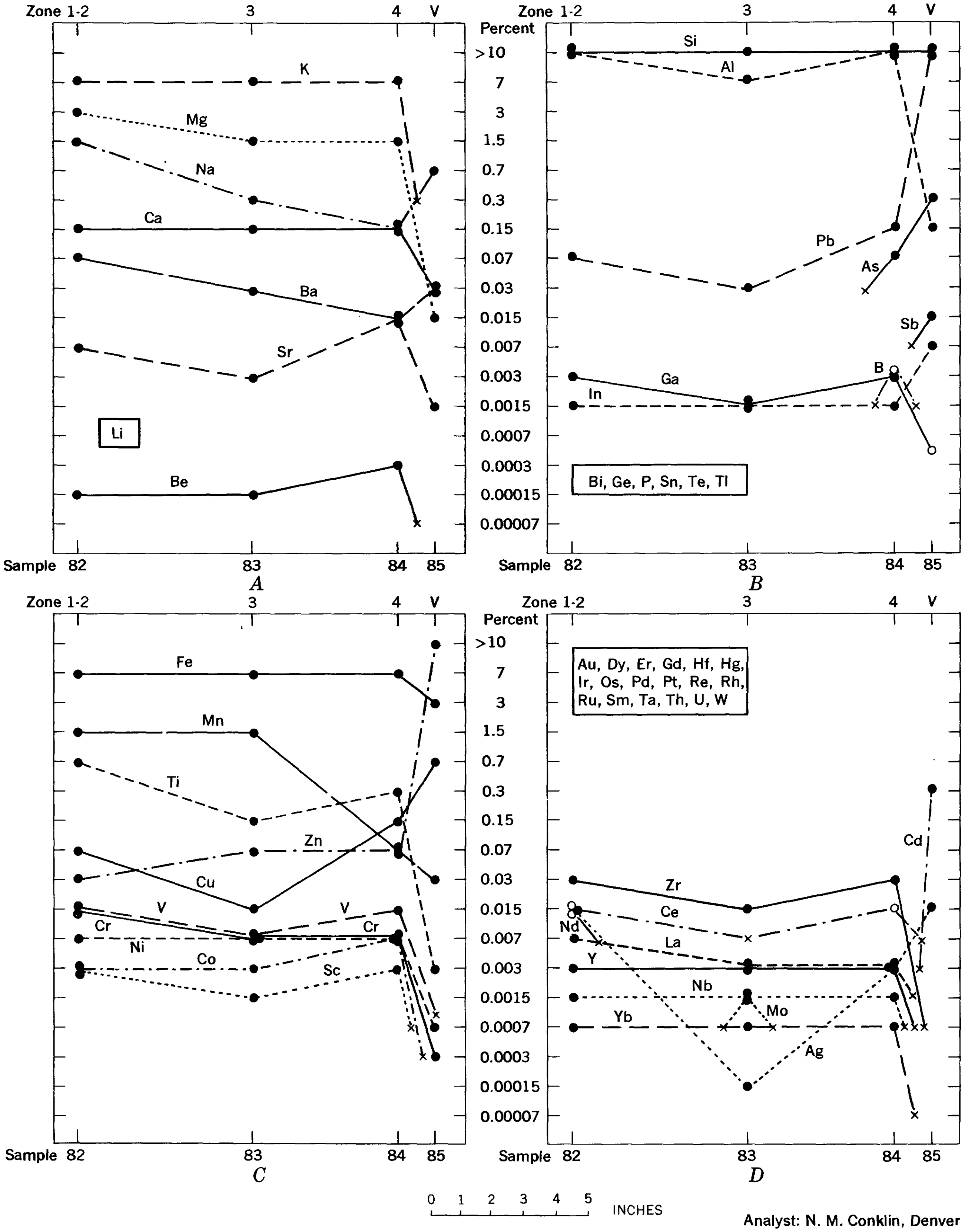
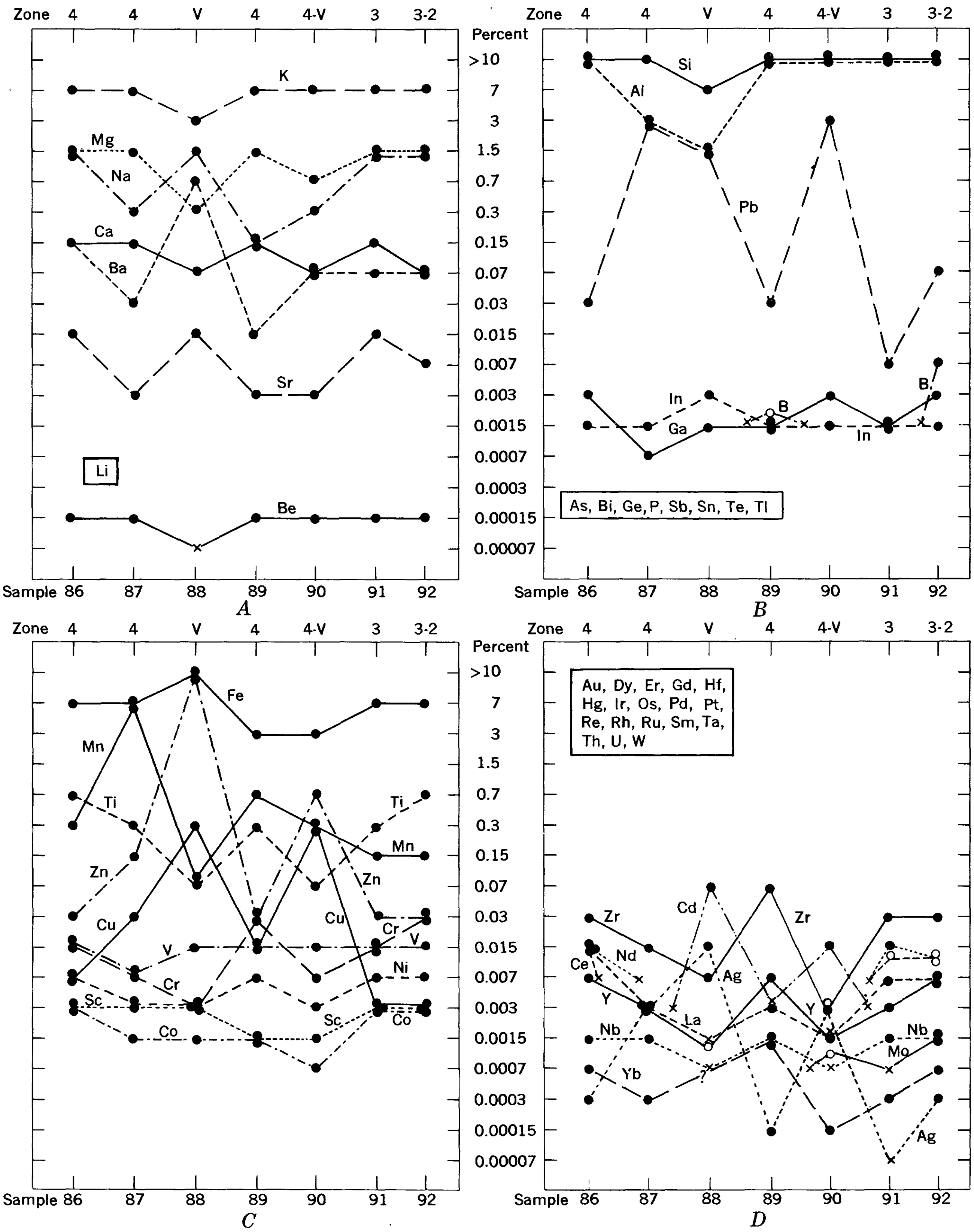


FIGURE 55.—Semiquantitative spectrographic analyses of progressively more altered biotite (metasedimentary) gneiss, Widow Woman-Cherokee mines, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.



Analyst: N. M. Conklin, Denver

FIGURE 56.—Semiquantitative spectrographic analyses of progressively more altered biotite-quartz gneiss, Cherokee mine, Russell Gulch area, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.

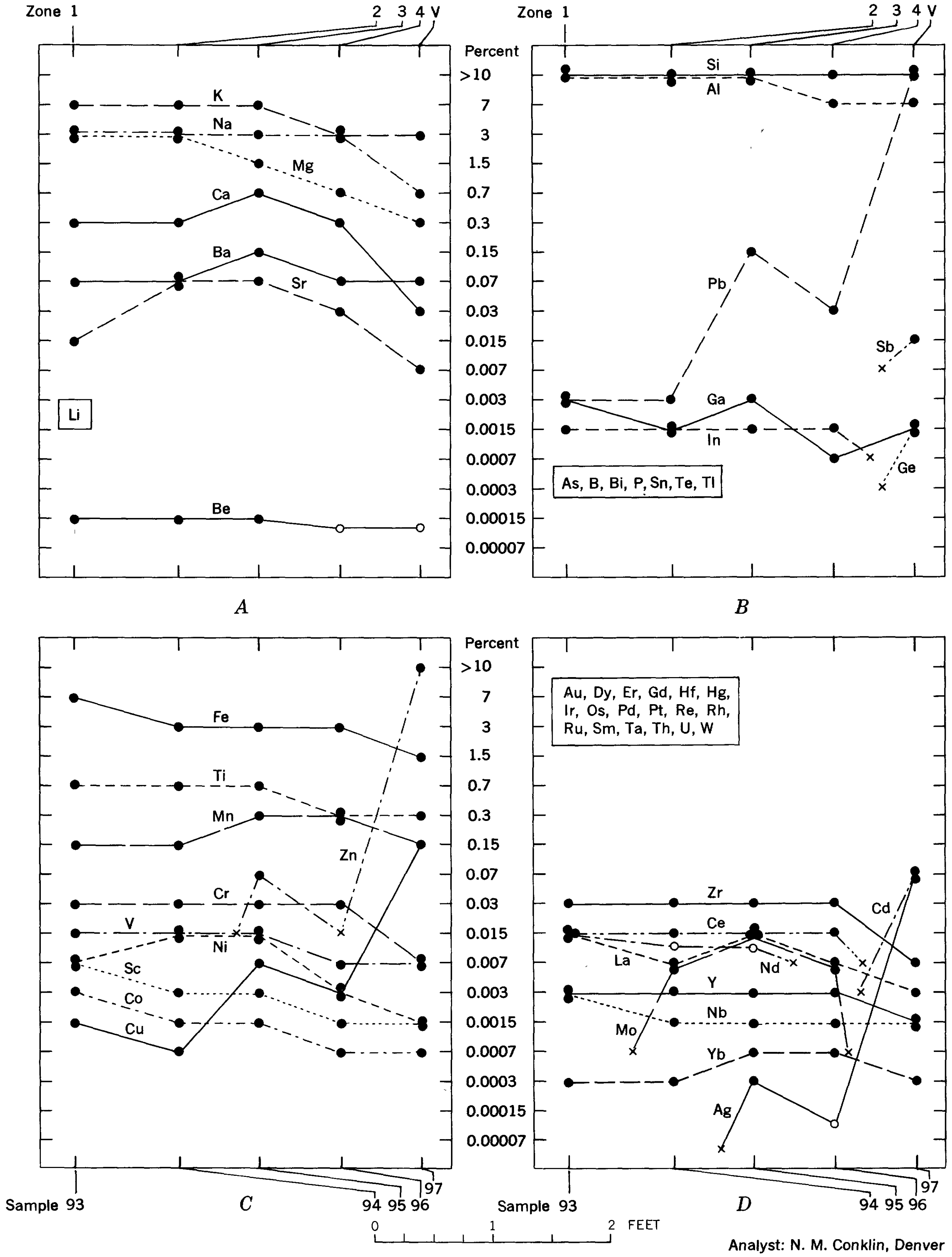


FIGURE 57.—Semiquantitative spectrographic analyses of fresh and progressively more altered biotite (metasedimentary) gneiss, Banta Hill mine, Russell Gulch area, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.

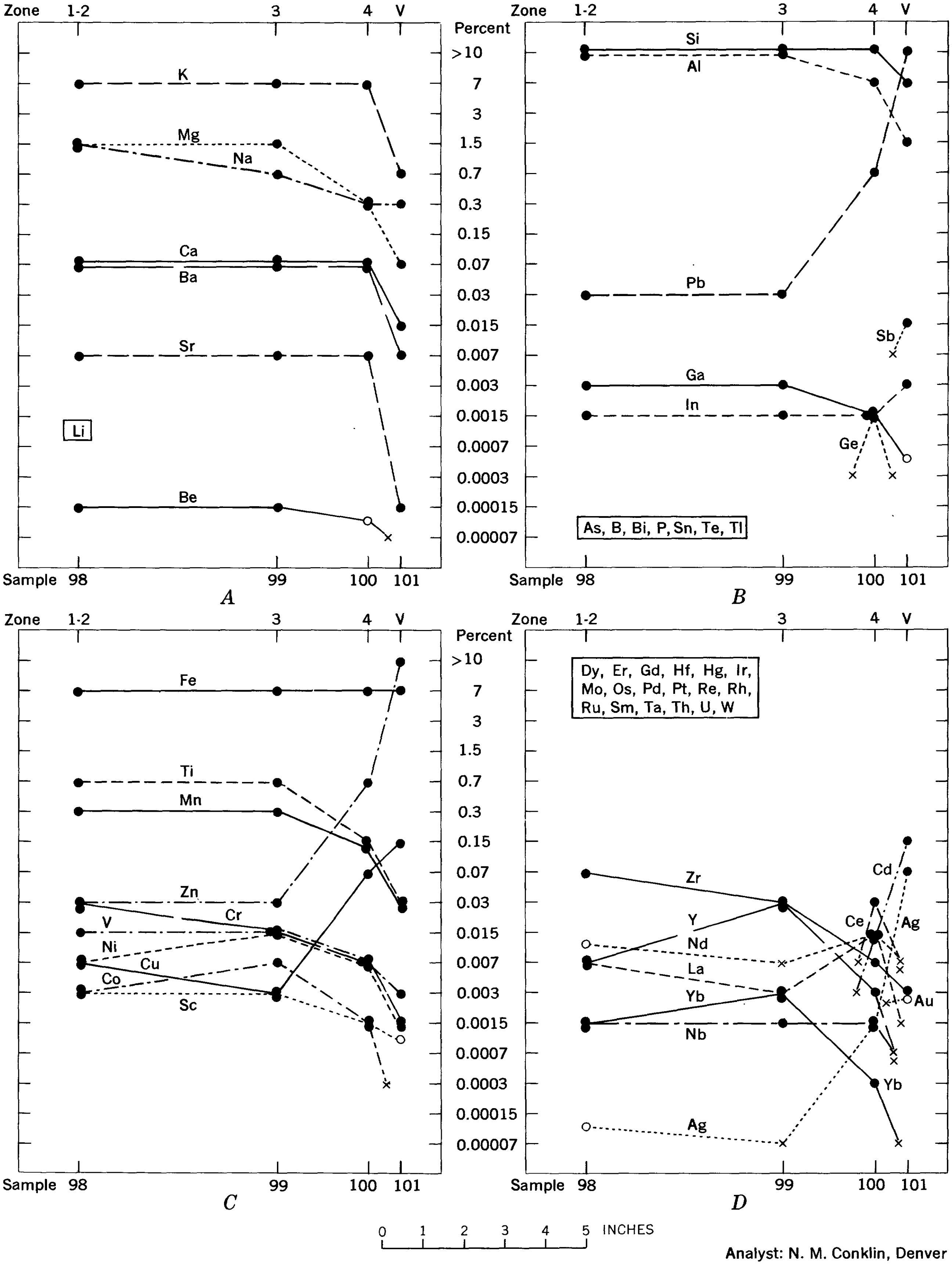
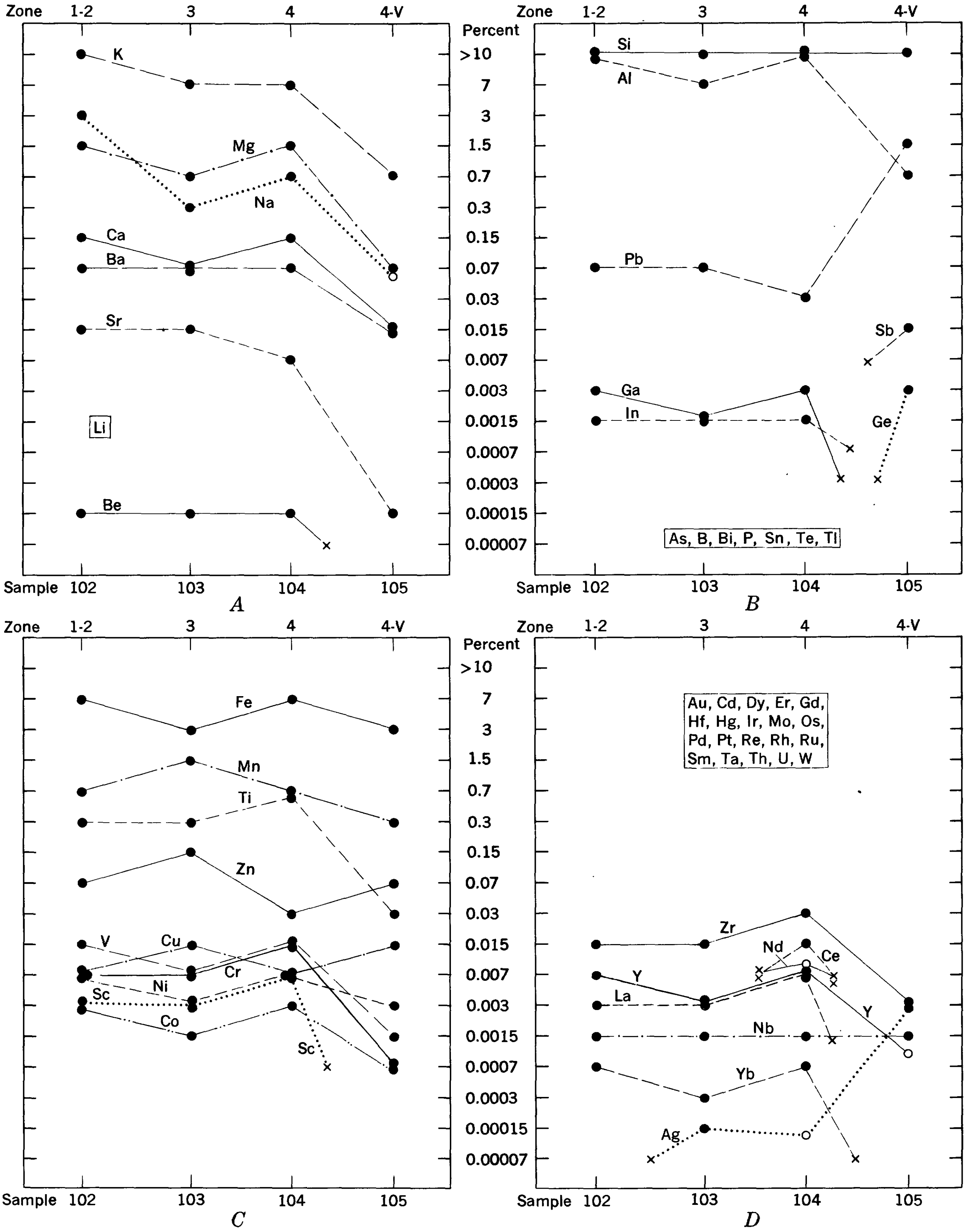
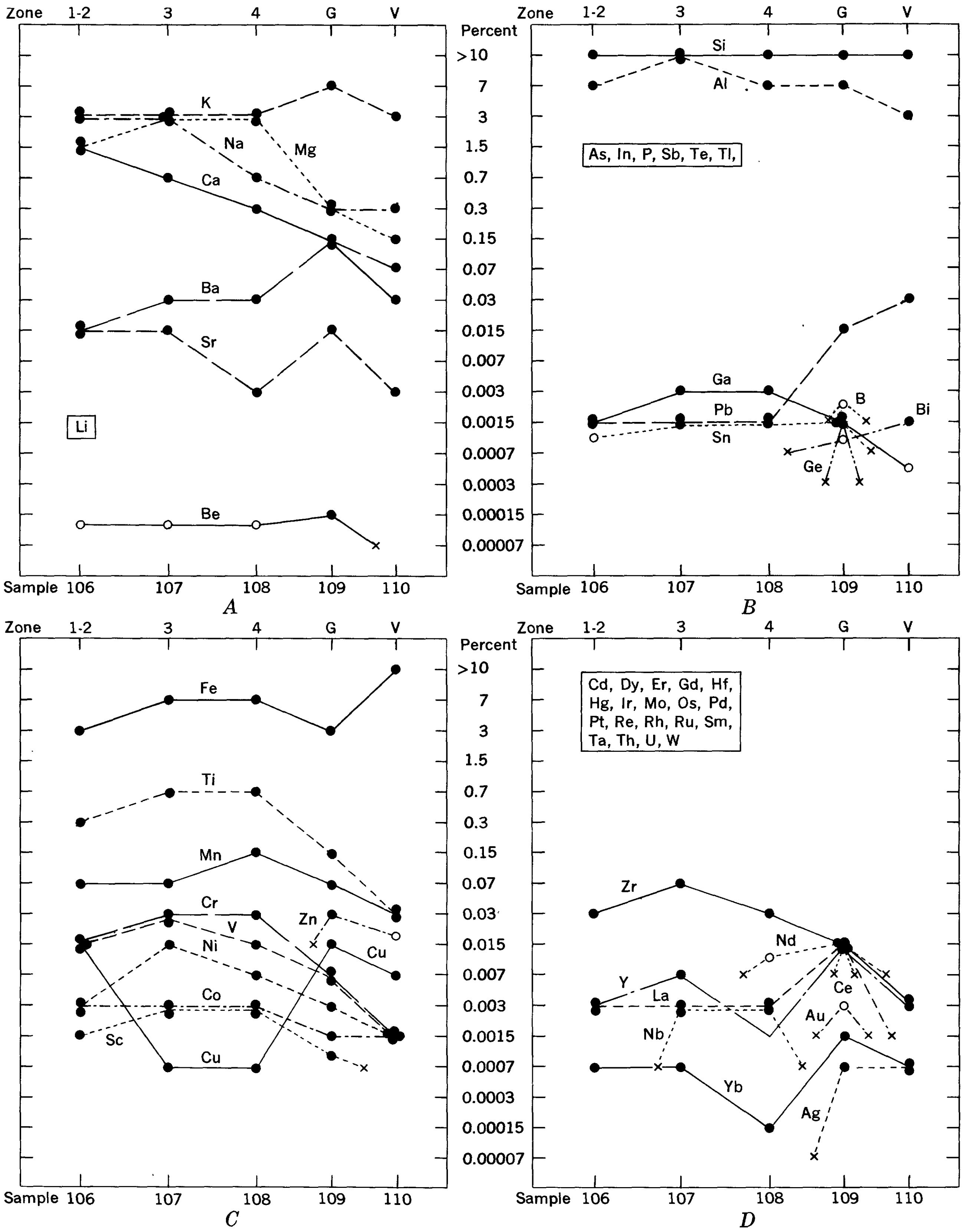


Figure 58.—Semiquantitative spectrographic analyses of progressively more altered biotite-quartz gneiss, Cherokee mine, Russell Gulch area, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.



Analyst: N. M. Conklin, Denver

Figure 59.—Semiquantitative spectrographic analyses of progressively more altered biotite-quartz gneiss, Cherokee mine, Russell Gulch area, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.



Analyst: N. M. Conklin, Denver

FIGURE 60.—Semiquantitative spectrographic analyses of progressively more altered metasedimentary gneiss, Phoenix mine, Idaho Springs district, Clear Creek County, Colorado. Analyst: N. M. Conklin.

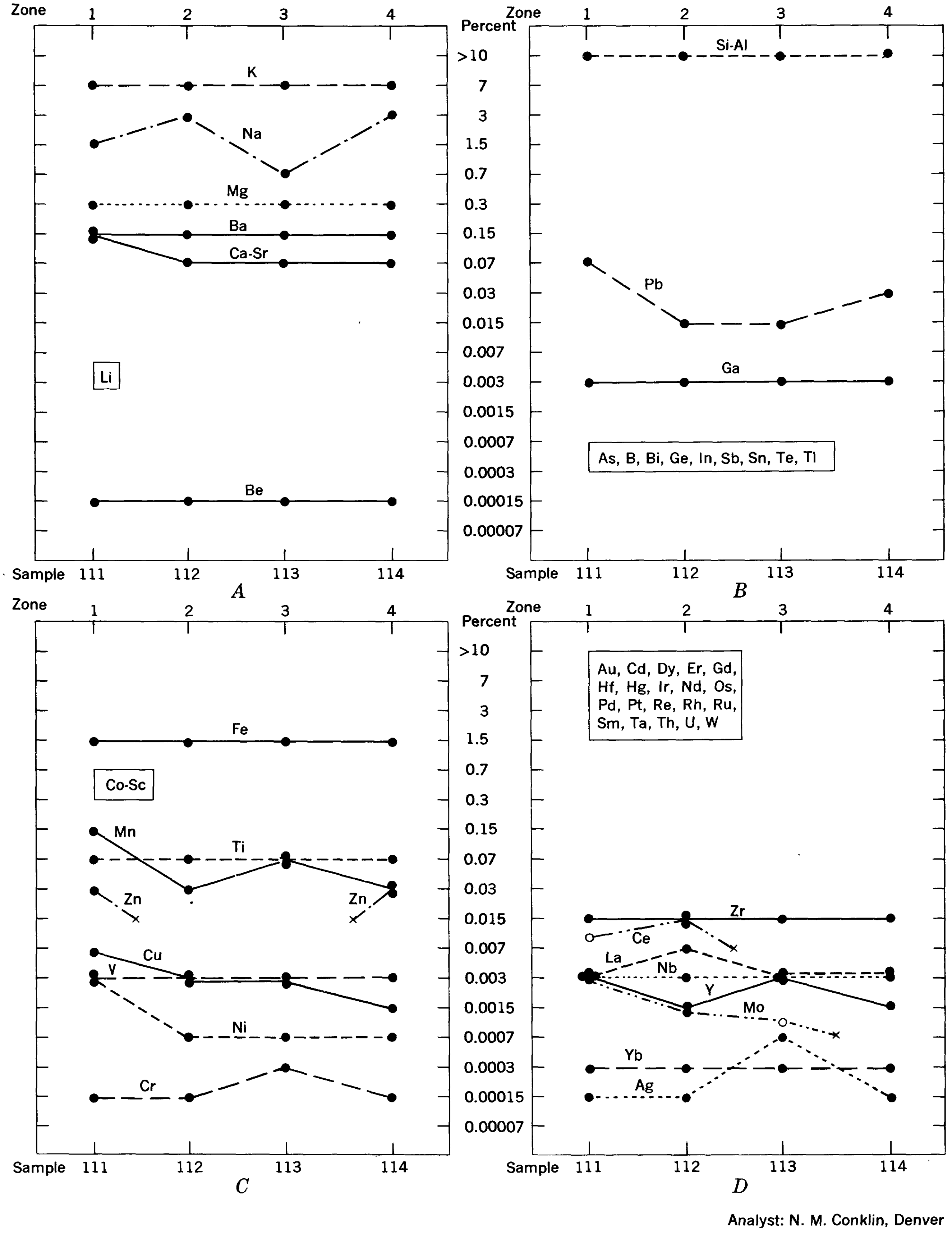
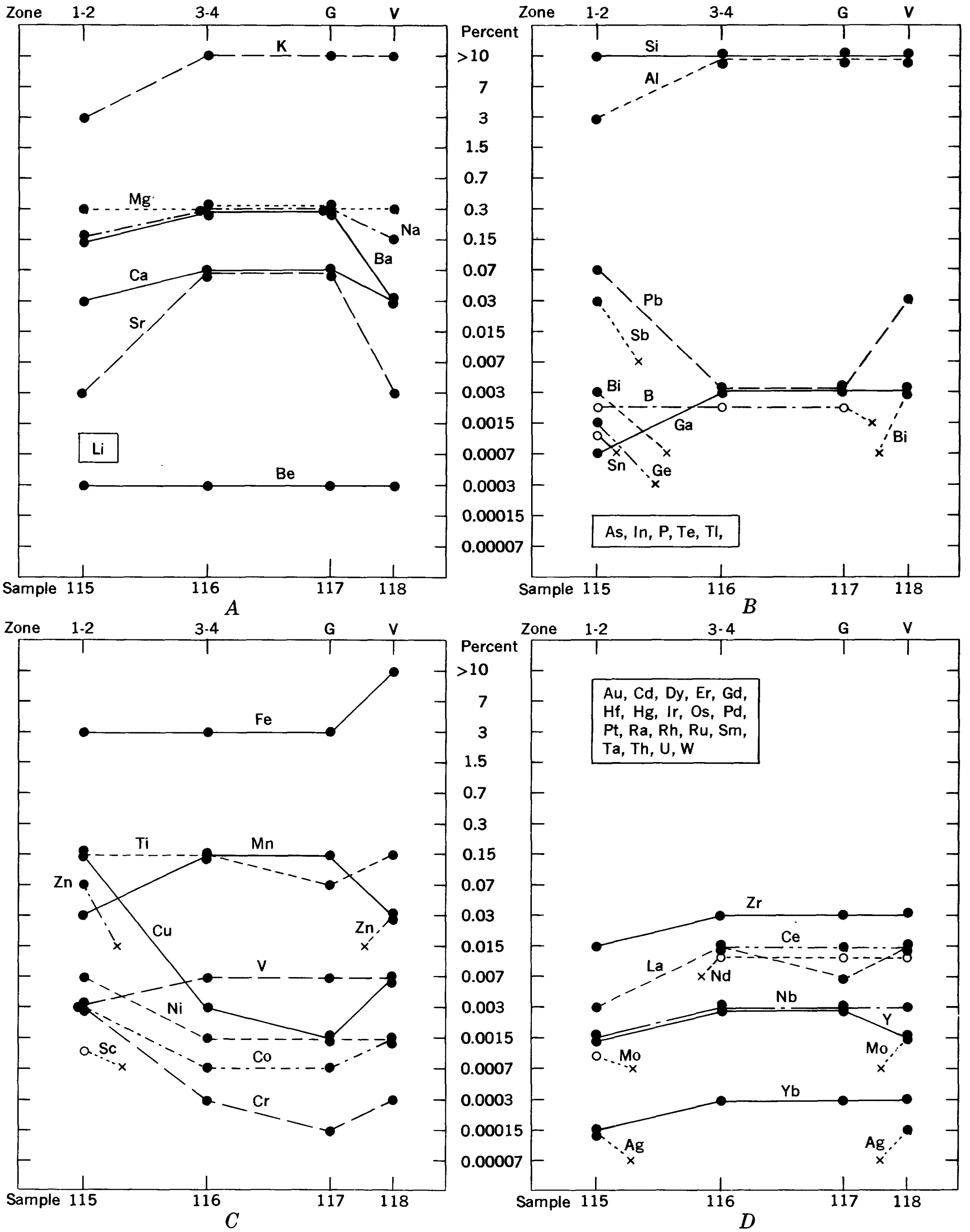
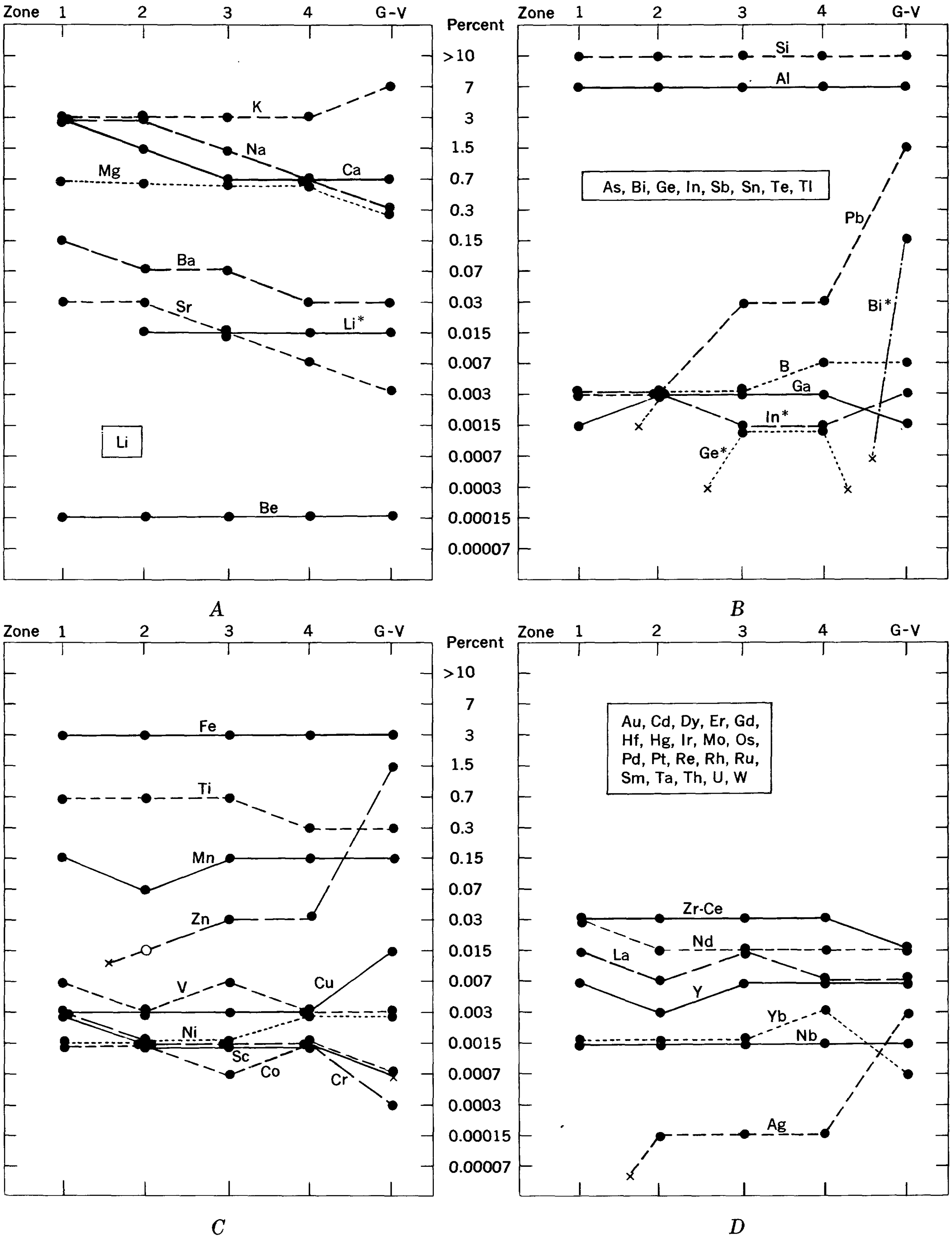


FIGURE 61.—Semiquantitative spectrographic analyses of fresh and progressively more altered quartz monzonite prophyry, Banta Hill mine, Russell Gulch area, Central City district, Gilpin County, Colorado. Analyst: N. M. Conklin.



Analyst: N. M. Conklin, Denver

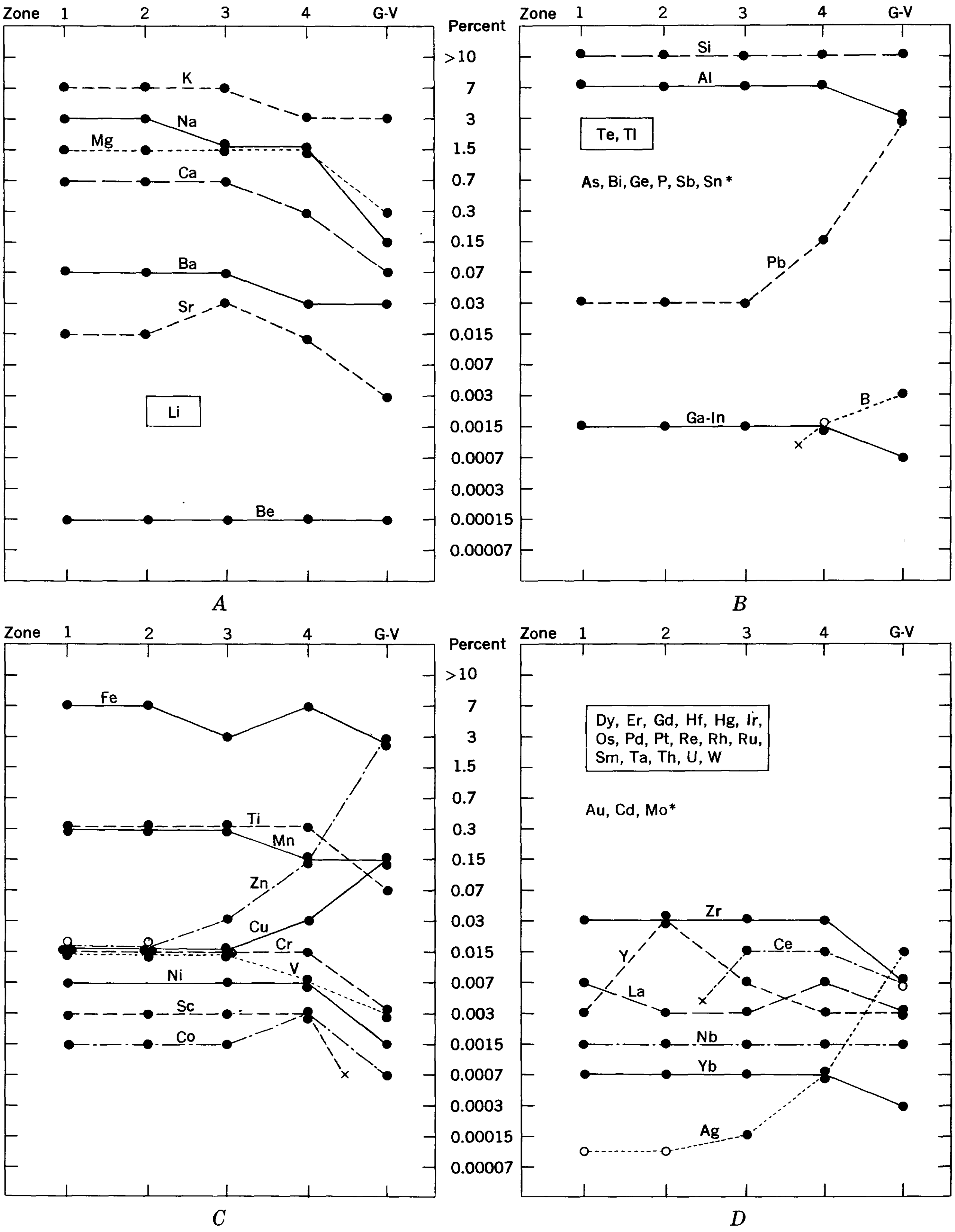
FIGURE 62.—Semiquantitative spectrographic analyses of progressively more altered bostonite, Phoenix mine, Idaho Springs district, Clear Creek County, Colorado. Analyst: N. M. Conklin.



*Averages of six samples or less

Analysts: P. R. Barnett and N. M. Conklin, Denver
and K. E. Valentine, Washington, D. C.

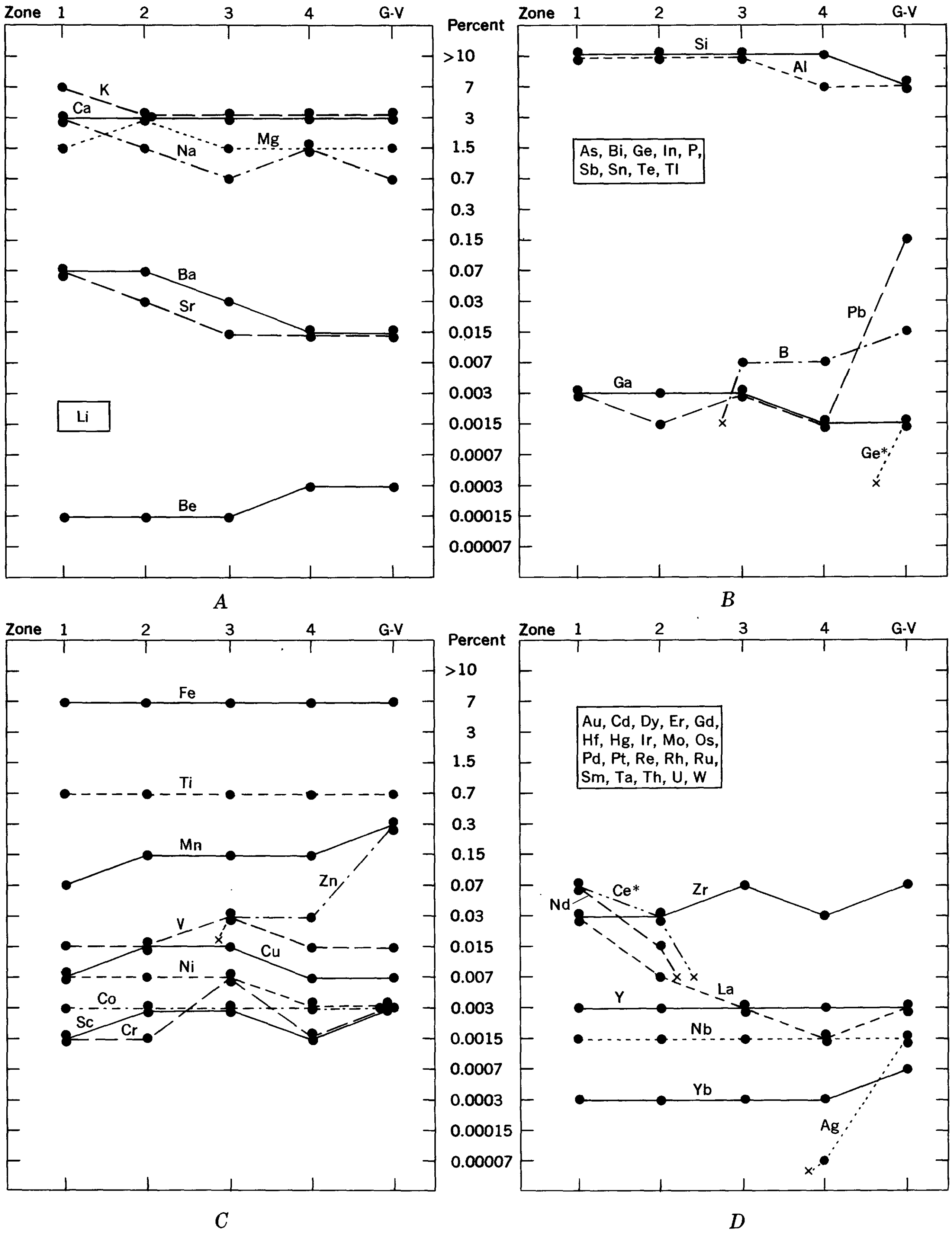
FIGURE 63.—Average values for elements determined by semiquantitative spectrographic analysis of 44 samples of fresh and progressively more altered granite, granodiorite, and microcline-quartz-plagioclase-biotite gneiss. Analysts: P. R. Barnett, N. M. Conklin, and K. E. Valentine.



*Sparse amounts in 1 to 4 samples only

Analysts: P. R. Barnett and N. M. Conklin, Denver

FIGURE 64.—Average values for elements determined by semiquantitative spectrographic analysis of 31 samples of fresh and progressively more altered biotite-quartz-plagioclase gneiss, biotite gneiss, and metasedimentary and migmatitic gneisses. Analysts: P. R. Barnett and N. M. Conklin.



*Average of two samples

Analysts: P. R. Barnett and N. M. Conklin, Denver

FIGURE 65.—Average values for elements determined by semiquantitative spectrographic analysis of 17 samples of fresh and progressively more altered quartz diorite. Analysts: P. R. Barnett and N. M. Conklin.

TABLE 25.—*Amount, in percent, of some regular elements in rocks*

Element	1	2	3	4	5	6	7
Li	--------	--------	--------	--------	0.00065	0.01792	0.007
Be	0.0001	0.0001	0.0001	0.0002	.0002	.00036	.00055
Sr	.051	.02	.07	.076	.0300	.0090 .0120	.03
Ba	.134	.075	.08	.15	.0250	.0670	.083
B	--------	--------	--------	.002	.0003	.0003	.0015
Ga	.0016	.002	.003	.0018	.0015	.00148	.003
Pb	.002	.013	.003	.07	.0016	.0030	.002
Ge	--------	--------	--------	.007	--------	.00035	.0003
As	--------	--------	--------	--------	.0005	--------	.00015
In	.0003	.0014	--------	--------	.0011	.00148	.000012
Sn	--------	--------	--------	.001	.0040	.0080	.0045
Sb	--------	--------	--------	.01	.0001	.0003	.00004
Bi	--------	--------	--------	.001	.00002	.0002	[1] (.0002)

[1] Requires further checking.

1. Biotite-muscovite granite and granodiorite, average of 5 samples (this report).
2. Metasediments, average of 14 samples (this report).
3. Quartz diorite, average of 3 samples (this report).
4. Dike rocks, average of 2 samples (this report).
5. Igneous rocks, average, Green (1953).
6. Granite, average, Rankama and Sahama (1950).
7. Acidic rocks, average, Vinogradov (1956).

ALKALI AND ALKALINE EARTH (REGULAR) ELEMENTS

Lithium, beryllium, strontium, and barium occur in trace amounts that generally decrease toward the vein. The distribution of sodium, potassium, magnesium, and calcium ions is noted earlier. Lithium is rarely detected in these wall rocks, but where present in microcline-quartz-plagioclase-biotite gneiss (fig. 38) no marked redistribution is noted. Lithium (Li^{+}, 0.68A) can be admitted into the lattices of silicate minerals in 6-fold coordination as replacement for Mg^{+2} and Al^{+3} [8]; biotite, muscovite, and their clay-mineral alteration products are the persistent potential lithium-bearing mineral structures in the rock. Lithium is readily leached and is not concentrated in zone 4.

Beryllium is present in amounts close to its analytical sensitivity in all of the wallrocks; and in all rocks except quartz diorite there is little variation across the zone of alteration. Beryllium is concentrated slightly toward the vein in quartz diorite. The beryllium ion is small and occurs in most types of silicate structures, where Be^{+2} (0.35A) most generally replaces tetrahedrally coordinated Si^{+4}, as well as in some oxide and phosphate mineral structures. Beryllium is readily leached and removed from zone 4 rocks by thermal waters.

Strontium and barium are moderately abundant trace elements in this group, and barium is detected in some rocks by standard analysis. Spectrographic analyses indicate that both elements are depleted in rock adjacent to the vein. Barium (Ba^{+2}, 1.34A) most commonly replaces potassium in mineral structures, and strontium (Sr^{+2}, 1.12A) has an affinity for calcium; their relative distributions here correlate directly with the abundance and distribution of potassium- and calcium-bearing minerals, especially feldspar and mica. Bray (1942) has shown that strontium is more abundant in feldspar than in mica in granodiorite from Boulder County, Colorado. Strontium also replaces calcium in carbonate and fluoride structures. A loss of strontium, shown especially well in figures 37 and 47, occurs in rock in which the clay minerals are being enriched in potassium at the expense of calcium. Strontium is more soluble than barium in thermal water and is more likely to be removed from the system; barite is a common gangue mineral in the vein and may be disseminated in adjacent walls along fracture and foliate surfaces.

REGULAR ELEMENTS

Boron, gallium, and lead are important trace constituents in this group (aluminum, silicon, and phosphorus are discussed earlier), but germanium, arsenic, indium, tin, antimony, and bismuth were identified in sparse amount in some rocks; thallium and tellurium were looked for but not detected.

Boron is sparse, occurs mostly in rocks containing abundant biotite, and is concentrated veinward. The element is small (B^{+3}, 0.23A) and is believed to be a replacement for silicon in hydroxyl-bearing minerals such as biotite and amphibole. The boron data are too sparse to permit assignment of boron entirely to veinward concentration by leaching from host rock minerals or by direct deposition from vein solutions.

Gallium occurs in most Front Range rocks in amounts comparable to those elsewhere, and well above its spectrographic sensitivity. Ga^{+3} ions (0.62A) may replace Al^{+3} in minerals having a mica structure as well as Fe^{+3} and Zn^{+2} in magnetite and sphalerite. In moderately argillized wallrocks gallium may be concentrated slightly, but it is generally leached from sericitized rock. This suggests that gallium may be concentrated in clay minerals as a result of rock alteration.

Indium is sparse in most Front Range rocks, in part because it occurs in amounts close to the spectrographic sensitivity limit for the element. Metasediments that contain abundant biotite, however, commonly contain indium in amounts comparable with averages in other rocks. The distribution of indium in general follows that of gallium, but In^{+3} (0.81A) may replace Fe^{+2}, Mn^{+2}, and Zn^{+2} in sulfide minerals as well as Al^{+3} in micas. Slight increase in indium occurs in vein gouge clay and in wallrocks caught up in some veins (figs. 55 and 56).

Germanium is of sparse occurrence in fresh bostonite and in sericitized metasedimentary rock in and bordering veins. The analytical sensitivity for germanium is higher than average values in comparable rocks else-

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[8] Here and in the following discussion of sizes and position of ions in mineral lattices the data from Bragg (1937), Green (1953), and Rankama and Sahama (1950) are the principal sources of reference.

where. Ge^{+4} (0.53A), which resembles gallium and is found most commonly in quartz replacing Si^{+4}, also is in simple silicates, topaz, garnet, and ilmenite. The element is readily weathered at the surface and is leached from argillized rock.

Tin is present in small amount in bostonite and metasediments from the Phoenix mine. The rocks in this area contain less tin than averages of other igneous rocks. Sn^{+4} (0.71A) can occur in plagioclase, clay minerals, fluorite, and sphene as a replacement of Fe^{+2}, Ca^{+2}, Sc^{+3}, Si^{+4}, and may also occur in some sulfide minerals. No pattern of tin distribution could be detected from these data.

Lead is a common trace constituent in most rocks in this area and occurs in amounts comparable with averages of analyses in table 25. The amount of lead in igneous and metasedimentary rocks of Precambrian age increases veinward whereas the amount of lead in dike rocks of Tertiary age decreases veinward. In addition to a site in galena, Pb^{+4} (0.84A) can occur as a replacement of Ca^{+2} in apatite, aragonite, epidote, and biotite, or of K^{+} in feldspars and micas. Most of the lead in vein rock and gouge clay is believed to occur in galena. Small fractures into sericitized rock may also contain galena.

Arsenic, antimony, and bismuth are very rarely found in the altered wallrocks in the central part of the Front Range mineral belt, and all are rare in the structures of silicate minerals. These elements more commonly combine with sulfur as sulfo-salts of the heavy metals (Cu, Fe, Ni, Co). Arsenic (As^{+5}, 0.46A) is most common in pyrite and arsenopyrite but may replace phosphorus in apatite. Bismuth (Bi^{+5}, 0.74A) may replace calcium in apatite and with Sb^{+5} (0.62A) occurs in an unknown granite mineral. Antimony, arsenic, and bismuth also can occur in galena structure. These elements occur in or close to the vein in wallrocks such as bostonite (Sb, Bi) and metasedimentary gneiss (Bi) at the Phoenix mine, biotite-quartz-plagioclase gneiss (As, Sb) at mines in the Russell Gulch area, and in microcline-quartz-plagioclase-biotite gneiss (Bi) in the Essex mine.

TRANSITIONAL METALS

The first group of transitional metals, characterized by progressive filling of electrons in the next outermost shell, comprises those from scandium through zinc (atomic numbers 21 through 30); the distribution of iron, titanium, and manganese was noted previously. Zinc, and less markedly copper, are concentrated veinward, but the heavy metal group, V, Cr, Co, Ni, and Sc, are closely related in their distribution and concentrations within these rocks; they tend to be removed from rocks adjacent to the vein.

Scandium is present in most rocks of Precambrian age in amounts greater than average. It is sparse or absent in dike rocks of Tertiary age. Sc^{+3} (0.81A) can replace Mg^{+2}, Fe^{+2}, or Al^{+3} in ferromagnesian minerals. In general scandium is lost as rocks are altered, and in quartz diorite it is lost gradually, but is concentrated in and adjacent to the vein. However, argillized metasedimentary gneisses (zone 3) often show a concentration of scandium whereas the adjacent sericitized rock in and bordering veins appears to have lost scandium. Vanadium is most abundant in igneous and metasedimentary rocks of Precambrian age, but is less abundant in later dike rocks. The average abundances of vanadium in these and other rocks in table 26 are comparable. Va^{+4} (0.63A) is most common in igneous rock minerals replacing Ti^{+4}, Fe^{+3}, and Al^{+3}; V^{+5} (0.59A) commonly replaces Al^{+3} in clay minerals and P^{+5} in apatite. Vanadium has been detected in amphiboles, micas, sphene, rutile, and magnetite-ilmenite.

TABLE 26.—*Amount, in percent, of some transitional metals in rocks*

	1	2	3	4	5	6
V	0.014	0.010	0.017	0.003	0.015	0.004
Cr	.002	.009	.021	.0015	.020	.0025
Co	.002	.002	.004	.0015	.0023	.0005
Ni	.002	.003	.008	.005	.008	.0008
Cu	.004	.015	.007	.078	.007	.003
Zn	--------	--------	--------	.05	.0051	.006

1. Biotite-muscovite granite and granodiorite, average of 5 samples (this report).
2. Metasediments, average of 14 samples (this report).
3. Quartz diorite, average of 3 samples (this report).
4. Dike rocks, average of 2 samples (this report).
5. Igneous rocks, average, Green (1953).
6. Acidic rocks, average, Vinogradov (1956).

A trend toward concentration of vanadium in argillized zones of some biotite-rich rocks is observed, but the general distribution trend is a loss of vanadium toward the vein. The sericitized zone of most rocks is leached, and vanadium also is removed during weathering as evidenced by the significant content of the element in water emerging from the Argo tunnel (table 34).

In the central part of the Front Range mineral belt, granite, granodiorite, some biotite-quartz-plagioclase gneisses (all biotite-rich), and amphibolites have a higher content of vanadium than microcline-quartz-plagioclase-biotite gneiss and porphyry dike rocks. Bray (1942) noted that biotite in monzonite from Jamestown, Colorado, areas contained 0.0670 percent V; plagioclase from the same rock contained 0.0007 percent V. It is inferred from these data that biotite and amphibole are the prime host minerals for vanadium in these rocks. As amphibole is altered in zone 3 and biotite is altered in zone 4, vanadium is released.

Chromium is present in most rocks in amounts well above the spectroscopic sensitivity limit but in amounts comparable with other rocks. It is probably concealed in oxide and silicate structures where Cr^{+3} (0.63A) replaces Al^{+3} and Fe^{+3} outside of the silicate framework. Chromium in magnetite-ilmenite is not released as rapidly by weathering as chromium in hornblende, mica, and garnet; Bray (1942) reports 0.11 percent Cr in biotite and 0.05 percent Cr in muscovite. The abundance of chromium generally decreases veinward but locally it increases slightly in the clay-mineral zone and in the vein.

Cobalt and nickel, present in most rocks, are most abundant in quartz diorite; nickel is more abundant than cobalt. In general, these data suggest that the elements are slightly less abundant in rocks of the Front Range than in comparable rocks elsewhere. Hornblende and biotite are the host mineral structures unless sulfide-mineral structures are present. Ni^{+2} (0.69A) is liberated from host minerals readily during weathering and is stable in aqueous solution, but can be fixed in clay-mineral structures. Co^{+2} (0.72A), which is soluble also, commonly does not remain in hydrated silicates.

These elements generally are removed from rocks during alteration as hornblende and biotite are altered. Rocks that produce a prominent montmorillonite alteration phase appear to localize nickel slightly, and nickel often is concentrated in vein-gouge clay—perhaps in a sulfide mineral.

Copper is an abundant trace constituent in these rocks in amounts generally higher than the average from rocks elsewhere. The porphyry dike rocks contain the most copper, and metasediments contain more than the igneous rocks. Cu^{+2} (0.72A) may replace Fe^{+2} and Mg^{+2} in silicates but more often occurs in accessory amounts in pyrite disseminated in zone 4 rock. Copper is readily soluble but reprecipitated in the vein in the presence of S.

In general copper is concentrated veinward in all rocks except porphyry dikes, where there appears to be a decrease veinward.

Zinc is not detected in amounts less than 0.02 percent, thus is reported only in the dike rocks whose content of zinc is considerably above average. Zn^{+2} (0.74A) can replace Fe and Mg in magnetite, amphibole, and biotite. These data suggest that zinc is leached from the argillized rocks and concentrated in the sulfide phase disseminated in rock caught up in the vein.

TRANSITIONAL AND SECONDARY TRANSITIONAL METALS

Of the large group of transitional and secondary transitional metals, yttrium, zirconium, niobium, neodymium, silver, lanthanum, and ytterbium are found in very small amounts; molybdenum, cadmium, gold, and cerium occur in some rocks in small amounts very close to analytical sensitivity limits; and ruthenium, rhodium, palladium, osmium, iridium, platinum, hafnium, mercury, tantalum, tungsten, samarium, gadolinium, dysprosium, erbium, thorium, and uranium were looked for but were not detected.

Zirconium is most abundant in granite, granodiorite, and quartz diorite, less abundant in the metasediments, and least common in the porphyry dike rocks. Values for Front Range rocks in table 27 are higher than average values for the earth's crust but the values for granitic rocks are comparable with average values given for granite. Most of the zirconium is assumed to reside in zircon, but minor amounts may be incorporated in mafic rock minerals. Zr^{+4} ion (0.79A) has coordinations of 6 and 8; however, this ion is not incorporated in mineral structures during early stages of crystallization or after alteration. Once formed, zircon is very stable under mechanical and chemical weathering and mild hydrothermal rock alteration conditions. In general there is a decrease in Zr toward the vein; the loss is small, and the variation across zones of altered rock is not great. Zirconium is not reported in available analyses of mineral water.

TABLE 27.—*Amount, in percent, of zirconium and niobium in rocks*

	1	2	3	4	5	6	7
Zr	0.043	0.025	0.033	0.015	0.028	0.046	0.02
Nb	.002	.001	.0015	.0022	.0024	.003	.002

1. Biotite-muscovite granite and granodiorite, average of 5 samples (this report).
2. Metasediments, average of 14 samples (this report).
3. Quartz diorite, average of 3 samples (this report).
4. Dike rocks, average of 2 samples (this report).
5. Earth's crust, Sisler, Van der Werf, and Davidson (1949, p. 706).
6. Granite, Rankama and Sahama (1950, p. 566).
7. Acidic rocks, average, Vinogradov (1956, p. 49–51).

Niobium was detected in about equal amounts in most rocks, but in amounts only slightly above analytical sensitivity. These numbers are comparable with averages in other rocks. The Nb^{+5} ion is of medium size (0.69A) and has a coordination of 6. The element is rare in sulfides, but is common in silicates as replacement for Ti in sphene, rutile, biotite, and clay minerals, and for Zr in zircon. Ta^{+5} (0.68A), which commonly occurs with niobium, was looked for but not detected in these rocks. Niobium content decreases slightly in more altered rock along veins in the Front Range and parallels the distribution of zirconium in the same rocks.

Trace quantities of silver were reported in altered wallrock and vein rock in about 70 percent of the localities sampled; gold was reported in vein rock in only two mines. Ag^{+} (1.26A) and Au^{+} (1.37A) are large ions not easily taken into mineral structures, and therefore are most commonly found in sulfides, sulfosalts, tellurides, or in the native state. In fresh rock, silver was not found in amounts above the 0.0001 percent analytical sensitivity limit. Veinward, in the argillic and sericitic zones, silver content increases on the average to 0.0018, and in the vein rock and gouge clay sampled up to 0.07 percent. Two samples yield 0.003 percent gold in vein and gouge material. In igneous rocks the average amount of silver is 1×10^{-5} percent, and of gold is 5×10^{-7} percent (Green, 1953).

Silver may be dissolved and moved as a sulfate, but gold generally is inert and remains in the native state. Silver may be reprecipitated in the vein environment as a secondary mineral.

Molybdenum, a rare constituent in most rocks, occurs in altered rocks in the peripheral ore zone in the Russell Gulch, Fall River, and the Idaho Springs (Phoenix mine) areas. Mo^{+4} (0.70A) possibly is concealed in sulfide structure of accessory minerals, and only where concentrated from 0.001 to 0.007 percent in the argillic zone is it detected. Molybdenum is dissolved readily, and when released in the absence of S it complexes with O. Thus no secondary deposits are formed, and the element is depleted in zone 4 and in the vein.

Cadmium is rare in most wallrocks of the mineral belt; Rankama and Sahama (1950, p. 712) report an average content of 0.00002 percent Cd in granite—well below the analytical sensitivity for spectrographic analyses. Biotite and other ferromagnesian minerals, plagioclase, and apatite are possible carriers of cadmium, but in an unknown manner. Cd^{+2} (0.97A) more commonly occurs concealed in sphalerite. Cadmium is dissolved readily as rocks are altered, but has a great affinity for S and may be enriched in the vein as greenockite (CdS). However, intensely altered metasedimentary wallrock and ore from veins in the peripheral ore zone in the Russell Gulch area commonly contain abnormal amounts of Cd. The maximum amount reported was 0.3 percent, the average content is 0.10 percent. Sphalerite is judged to be the host-mineral structure.

Yttrium, lanthanum, cerium, neodymium, and ytterbium are more abundant in rocks in the Front Range than in the rocks listed in table 28. The sizes of these ions are large (Y^{+3}, 0.92A; La^{+3}, 1.14A; Ce^{+3}, 0.94A; Nd^{+3}, 1.04A; Yb^{+3}, 0.86A), their coordination ranges from 6 to 8, and their chemical properties are very similar. Yttrium and the lanthanides (La, Ce, Nd, and Yb) are common low-concentration constituents of minerals in igneous and sedimentary rocks. They have an affinity for phosphorus and fluorine in oxide- and silicate-mineral structures in apatite, xenotime, fluorite, zircon, monazite, garnet, and biotite. In many of these minerals yttrium and lanthanide ions replace Ca^{+2} (0.99A) ions as in calcite, sphene, apatite, amphibole, garnet, and plagioclase; La may replace K in K-feldspar.

The distribution of individual transitional metals varies: Y and La generally are lost in rocks adjacent to the vein; Ce and Nd occur in amounts close to analytical sensitivity in only 60 percent of rocks sampled and have no pronounced trend toward gain or loss across a zone of altered rock; Yb, which is more commonly captured by replacement of Ca, is frequently concentrated veinward. There is no marked concentration of these elements in clay-mineral structures.

TABLE 28.—*Amount, in percent, of some transitional and secondary transitional metals in rocks*

	1	2	3	4	5	6	7
Transitional metals							
Sc	0.003	0.0023	0.002	--------	0.0005?	0.00013	0.0007
Y	.0078	.004	.004	0.0022	.0017	--------	.002
La	.027	.005	.004	.003	.0019	.0043	.0046
Secondary transitional metals							
Ce	0.046	--------	0.023	--------	0.00247	--------	0.006
Nd	.027	--------	.02	--------	.0018	--------	.004
Yb	.0008	0.0011	.0003	0.0002	.00027	--------	.0002

1. Biotite-muscovite granite and granodiorite, average of 5 samples (this report).
2. Metasediments, average of 14 samples (this report).
3. Quartz diorite, average of 3 samples (this report).
4. Dike rocks, average of 2 samples (this report).
5. Igneous rocks, average, Green (1953).
6. Granites, Rankama and Sahama (1950).
7. Acidic rocks, average, Vinogradov (1956).

Yttrium and the lanthanides are readily removed by solution when the minerals containing them are altered. The ions are carried off and generally accumulated in carbonate sediments (Rankama and Sahama, 1950). The spectrographic data suggest, therefore, that these ions are more commonly concealed in structures of accessory minerals that are resistant to alteration even in the argillic zone.

POWDER pH OF ALTERED WALLROCKS

The relative acidity or alkalinity of water, an important property in chemical reactions, is expressed in terms of hydrogen-ion concentration, or pH. Rocks are composed of variable amounts and types of hydrous minerals and contain water-filled pore spaces. When pulverized and suspended in a standard water solution, the rocks impart a hydrogen-ion concentration to the suspension (Stevens and Carron, 1948); this hydrogen-ion concentration then is assumed to be characteristic for the rock environment. It is not this simple, however, and Hemley (written communication, 1960) suggests that pore water is the least contribution to the powder pH. It is mostly a measure of the extent of room temperature hydrolysis of the powdered minerals involved; fresh rocks give high values, argillic and sericitic minerals give low values. These low values may reflect low hydrolyzing character and (or) included traces of acid sulfates from oxidized sulfides. Nevertheless, powder pH data are a measure of these values and should serve as an aid in the evaluation of abnormal rock oxidation.

A qualitative measure of the hydrogen ion concentration of progressively more altered powdered wallrocks indicates that rock pH decreases toward the vein (table 29). Thus the pH of rocks known or inferred to have been exposed to extensive migration of oxidizing waters is significantly lower than the pH of rocks not so exposed. The pH values from rocks sampled in old, abandoned, commonly wet or water-filled mine workings are generally lower than pH values obtained from the same types of rocks in newer and (or) drier workings. The pH values obtained from granitic rocks on the fringes of the fractured and mineralized area are high (pH of 8), and the writer infers that the high pH value reflects the fact that these rocks have never been exposed to extensive migration of oxidizing waters. Intensely altered wallrock in the central ore zone has a pH of only 4.5, whereas "fresh" rock in the same zone has a pH of about 6, equivalent to the pH of the same rock where intensely altered in the peripheral ore zone. This may indicate also that rocks of the central ore zone had more intense hydrothermal permeation than those of the peripheral ore zone.

TABLE 29.—*Summary of qualitative powder pH determinations on several variously altered wallrocks*

Zone	pH[1]	Mine	Sample
		Granite	
4	6.7, 7.8	Nabob	154
4	8.4, 8.0	do	149
G	7.1, 8.0	do	151
G	8.9, 8.8	Jo Reynolds	144
V	7.1, 6.9	Nabob	155, 156
V	7.9, 8.2	do	146
		Granodiorite	
2	8.6, 8.5	Hayes and Wheeler	42
3	7.8, 8.4	do	43
3c	7.9, 8.5	do	44
4	8.8	do	45
		Quartz diorite	
3	8.4, 8.7	Nabob	77b
4	8.2, 8.4	do	78
G	7.6, 7.4	do	79
2	8.8, 9.0	Jo Reynolds	16
2	8.5, 8.3	do	71
3	8.4	do	73
3	8.4, 8.4	do	72
3–4	8.4, 8.3	do	74
V	8.5, 8.1	do	75
		Granite pegmatite	
3–4	3.6	Alma Lincoln	[2] 411
4	2.7	do	412
V	3.1	do	416
V	8.1, 8.7	Carroll	203, 206
4–V	5	Cherokee	334
		Bostonite	
3	4.5, 4.7	Mammoth	[3] 443
4	4.5, 5.2	do	444
4	4.5, 4.9	do	439
G–V	4.5, 5.7	do	440
V	4.5, 4.6	do	441
		Microcline-quartz-plagioclase-biotite gneiss	
3	8.6	Carroll, 3d Level	205
4	8.1	do	204
G	7.5	do	203
V	7.4	do	207
3–4	5	Carroll, 2d Level	[3] 195
4	5	do	194
G	4.5	do	191
V	4.5	do	197
3	4.8, 4.9	Mammoth	[3] 268
4	5.7, 4.4	do	267
V	5.6, 4.3	do	266
2	6.0, 8.3	Dumas-Kinney	261
2–3	6.0, 8.2	do	263
3	6.0, 8.0	do	259, 260
V	5.5, 7.6	do	258
2	5	Alma-Lincoln	[4] 234
3	4.6	do	240
4	4.2	do	236
V	3.5	do	246
		Biotite-quartz-plagioclase and related gneisses	
2–3	8.0	Dorit	323
3	7.9	do	325
4	7.0	do	322
2	4.5	Golconda	408
3	4.5	do	409
G	4.5, 5.0	do	378
4–V	4.5	do	382a
2	8.2	Almadin	372
2–3	8.2	do	373
3	8.2	do	370
3	7.9	do	369
3–4	8.6	do	368
1–2	4	Old Stagg	[5] 301
V	4	do	302
3	6	Phoenix	[3] 107
4	4	do	110
G–V	3	do	109
		Amphibolite	
1–2	5	Foxhall	403
3–4	5.6	do	404
		Garnet-quartz gneiss	
2–3	6	Golconda	386
3	6	do	385
3–4	6	do	384
4–V	6	do	390
G	4.5	do	383
V	5	do	389

[1] The second value represents determination in the same sample after it was allowed to stand for 1 month.
[2] Near adit.
[3] Near surface.
[4] Very strong ground-water movement, near surface.
[5] Ground-water movement strong.

GENERAL CONCLUSIONS

The alteration of a varied assemblage of metamorphic and igneous wallrocks by mild hydrothermal solutions in the central Front Range mineral belt left a more informative record than intense hydrothermal alteration of a single rock type might have done. The distribution of major and trace elements in these rocks may be related to (1) the stability of host minerals, (2) the geochemical nature and response of the elements to the altering hydrothermal solutions, and perhaps (3) the physical and chemical nature of the ore solution. Some significant conclusions concerning rock alteration may be deduced from these chemical data.

1. Oxygen ions comprise nearly 90 percent of the volume of fresh rocks, and the weight of oxygen in equivalent volumes of rock remains nearly constant across altered zones. Any apparent loss is adequately accounted for by the addition of H^+ to oxygen. The introduction of sparse F and S ions locally in zone 4 must displace some O ions if equal volume relations are maintained as assumed.
2. Most other ions are small and fit into the interstices of the close-packed oxygen network in a manner governed by geochemcial factors such as ion size, charge, concentration, solution pH, crystal structure, and temperature. Because a majority of the rocks are composed of minerals with sheet-type silicate structures, only minor shifts in interstitial ions need be postulated to occur when H^+ ions in acid waters invade the rock and convert O^{-2} ions to $(OH)^-$. In general K, Fe^{+2} (and total iron), C, H, S, and some Al were added to the altered rock while Si, Na, Ca, Fe^{+3}, and Mg are lost from the wallrock.
3. The structures of the host and newly formed minerals may influence ion distribution. Interstitial ions with low ionization potential are removed first, hence alkali and the alkaline earth ions (except for potassium which has a large ionic radius and almost certainly was introduced to zone 4 of certain rocks by altering solutions) are fewer toward the vein and are leached during rock alteration. K-feldspar is stable under all alteration conditions, and potassium is not released; biotite is altered in zones 3 and 4, and the clay-mineral structure that results retains much of the potassium that was released. Plagioclase is partly altered in zone 2, and all ions in excess of those required to form (or to be acceptable in) clay minerals—such as sodium and calcium—are freed. The anomalous loss of strontium in zone 3 and its reappearance in zone 4–V (figs. 37 and 47) is correlated with (a) its release from altered plagioclase, (b) its incompatibility in clay-mineral structures, and (c) its stabilization as a replacement for calcium in calcite. The alteration of biotite in zone 4 is correlated in part with the release of barium, lead, gallium, boron, all of which have an affinity for iron, and perhaps some rare earth elements. Montmorillonite and illite clay structures may retain trace elements such as vanadium, zinc, titanium, and chromium.
4. The decrease in K^+ fixation in vein walls of K-feldspar-deficient rocks is thought to relate to the K^+/H^+ activity ratio in the hydrothermal system. Thus local changes in the composition of the hydrothermal solution occasioned by a more rapid local alteration reaction in the host rock could effect large-scale local changes in the alteration environment. Solutions moving in K-feldspar-rich rocks should tend to adjust close to the K-feldspar-mica-quartz equilibrium curve and additions of K^+ tend to produce sericite in these rocks. A similar solution moving in rapidly altering K-feldspar-poor rocks would be farther from the equilibrium curve in the clay-mineral field and a similar addition of K^+ would produce montmorillonite or illite.
5. Most of the trace elements occur in amounts equal to or slightly greater than their average contents in similar rocks elsewhere. Therefore the evidence for geochemical partition of trace-element ions during rock alteration is not clear. In general the smaller ions (in each period) with larger ionization potential and higher valences occur or tend to be concentrated in rock near the vein (table 30). These elements coincidentally have a pronounced affinity for sulfur, and several are commonly major constituents or concealed trace elements in sulfide ore minerals. Elements with low ionization potential are relatively soluble and are more readily removed from zone 4 rock into the vein solutions.

TABLE 30.—*Generalized distribution of trace elements in rocks in the Front Range area*

[Asterisk indicates major rock-constituent ions]

Ions concentrated mostly in unaltered rock (or leached from altered rocks)	Indeterminate or variable ion concentration	Ions concentrated mostly in altered rocks adjacent to veins
Li, *Na, Be, *Mg, *Ca, Sr, Ba		*K.
*Si, Ga, In	*Al, B, Sn, P	Ge, Pb, As, Sb, Bi.
V, Cr, Co, Ni, Sc	*Fe, Ti, *Mn	Zn, Cu.
Nb, Mo, Cd, Y, La	Zr, Ce, Nd	Ag, Au, Yb.

6. Powder pH measurements generally show a decrease of H ions in rocks away from the vein, but this is mainly the effect of differential rates of hydrolysis of the minerals involved. Lower than average pH values are most useful as indicators of more intense rock oxidation near the vein. The pH data suggest also that rocks in the central ore zone were subjected to a more intense and acid hydrothermal and (or) supergene alteration environment than were similar rocks in the peripheral ore zone.
7. Very sparse data indicate that ions in addition to K and S may be introduced (in an unknown form) into zone 4 rock by vein solutions; their identity

and proportions seem to be consistent with vein mineralogy; intensely altered metasedimentary and igneous rocks in the intermediate ore zone contain copper, lead, zinc, and silver in amounts above average. Quartz diorite, which crops out almost exclusively in the peripheral ore zone, contains average amounts of copper and lead and above average amounts of zinc and silver. There is no compelling chemical evidence to suggest that introduced ions (except H) penetrated appreciably into the wallrocks beyond the zone 3-zone 4 interface.

ORIGIN OF THE ALTERED WALLROCKS IN THE CENTRAL PART OF THE FRONT RANGE MINERAL BELT

The sequence of the hydrothermal activity that produced the altered wallrocks in the central part of the Front Range mineral belt was compiled from detailed observations of the altered-rock environment and from some reasonable inferences.

LOCALIZATION OF ALTERED ROCK

The location and character of the altered rocks studied is governed, at least in part, by several geological factors, including location within an area containing a favorable combination of rock structures, a sequence of mild hydrothermal solutions, and variable rock mineral-chemical relations. First, it is assumed that the source of the hydrothermal solutions is at depth. Second, it is inferred that the openings along the fissure veins were produced by dilatancy caused by regional shear stresses; these openings were loci of deposition of the vein-forming minerals during early Tertiary time (Eckelmann and Kulp, 1957, p. 1128).

Rock alteration took place outward from openings, which range in size from large vein faults a foot or more wide to pore spaces of microscopic width, at a depth of at least 1 or 2 miles below the existing land surface.[9] Thus some channelways were throughgoing whereas others permitted only slow percolation of solutions and ions. These openings provided access for primary hydrothermal solutions to the rocks and an outlet for altered hydrothermal solutions that resulted from the exchange of constituents between the rock and the solution. Finally, the vein faults locally provided a locus for deposition of the ore deposits.

The principal routes of access for rising hydrothermal solutions were first along deep-reaching faults, then along less pervasive faults that connected with the main channels, and finally out into the bedrock along minor openings such as foliation planes, joints, small fractures, and pore spaces. The principal channels resulted from repeated faulting that formed open spaces and allowed access to rising solutions. The continuity of these faults and their numbers depends to a large extent on the lithology and kind of folding present in local areas within the whole area. As examples, the Central City and Freeland-Lamartine districts (Sims, Drake, and Tooker, 1963; Harrison and Wells, 1956) are sharply contrasting. At Central City, broad open folds in an alternating sequence of thick competent and thin less competent gneisses contain continuous, wide, closely spaced vein faults. At Freeland-Lamartine, a tightly folded, in part overturned, and generally distorted series of thin competent and thicker incompetent metasedimentary gneisses produces fewer, less continuous, and narrow vein faults. At Central City, the width of altered walls and altered rock zones in the central zone are consistently greater than along veins with a similar ore-mineral assemblage in the Freeland-Lamarine district. The penetration of solutions into the walls away from the faults is either restricted or aided by foliation planes, joints, and similar structures. Prominent foliation parallel to the vein fissure restricted the ease of migration of solutions into the walls, whereas prominent foliation at a large angle to the vein fissure, cross fractures related to faulting, and joints aided the migration of solutions into the walls.

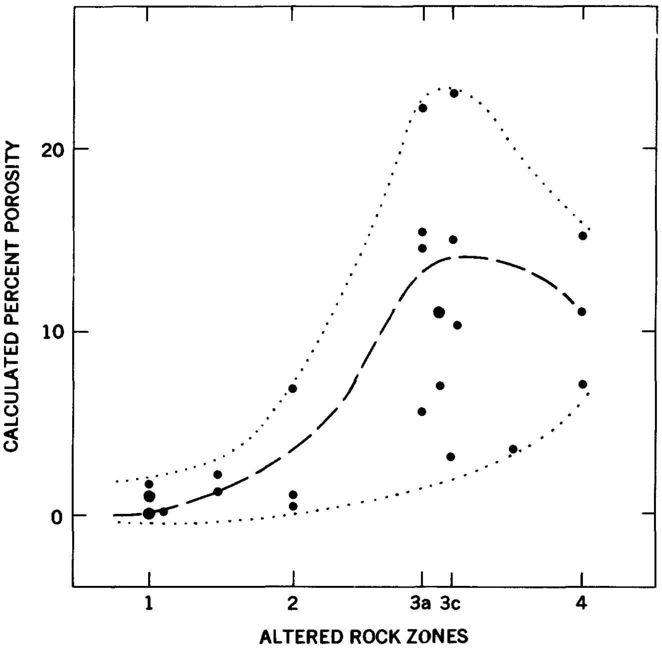

FIGURE 66.—Plot showing variation in apparent percent porosity calculated from bulk and powder density values (table 18). Dashed line indicates an average value, dotted lines indicate the range in values.

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[9] From geomorphic evidence Murray (1956) suggests that the Continental Divide in early Tertiary time extended northwestward from Pikes Peak (14,274 ft) through the Tarryall Mountains, Mount Evans (14,260 ft), Gray-Torreys (14,274 ft), and northeastward to James Peak (13,260 ft), passing across the mining district area. The present position of the Continental Divide is the result of the headward erosion westward of the Platte and Arkansas Rivers since Tertiary time. The ore deposits at Idaho Springs and at Central City, which range in altitude from 7,000 to 10,000 feet, underlie the Tertiary land surface by at least 1 to 2 miles.

It appears that significant pore space for the migration of fluids and deposition of vein minerals was not formed by leaching, as noted in particular at East Tintic district, Utah (Lovering and others, 1949, p. 27). First, the narrow alteration envelopes in the Front Range area studied, indicate either that hydrothermal solutions were not able to alter the rocks very far from the veins or that they did not freely move very far into the walls. There is no evidence that zone 4 rock has been extensively leached, and the formation of montmorillonite, the principal alteration product of plagioclase and hornblende in zones 2 and 3, may well have restricted solution movement away from the vein. Second, although pore-space measurements, determined from bulk and powder density data, indicate a veinward increase in pore space of as much as 20 percent in the clay zones (fig. 66), thin sections of these rocks fail to show any evidence of open space[10] of this order of magnitude. The absence of disseminated base-metal sulfides in altered rock zones also indicates that the apparent increase in porosity did not result in permeable rock; such open space likely would have been filled by sulfides during the ore-depositing stages. Bastin and Hill (1917, p. 103) were puzzled by the contrast between the localization of vein minerals and those formed by replacement of wallrock minerals. "One of the most striking exemplifications of these contrasts is the scarcity or entire absence of galena and sphalerite in the walls of many fissure fillings in which these minerals are very abundant; in such walls pyrite is the dominant sulfide. Carbonates may be present in a filled fissure but absent from its walls. Conversely, sericite the commonest alteration product in the walls, is never found in fissure fillings."

In a very general way the spectrographic data summarized in figures 67 and 68 show that some essential elements in ore minerals are only moderately concentrated, if at all, even in strongly altered wallrock included in the vein zone. The data also suggest, however, that lead and zinc are more abundant in wallrocks of galena-sphalerite veins. There is little or no abnormal increase in base-metal vein-mineral constituents in zones 1 to 3 of most rocks, and zinc and lead are, if present at all, in amounts below the sensitivity limits of detection. Copper, lead, zinc, and silver are concentrated locally in small amounts in zone 4 rock, but this may be partly a leakage outward from the vein along small fractures. The progressive concentration of iron is less marked because it is a major constituent in the rocks. These sparse data suggest that most elements which are in trace amounts in sulfides are not greatly concentrated or dissipated from their normal abundance in the wallrocks.

The sequence of rock alteration and metallization events is not wholly clear, but the bulk of the evidence indicates that most wallrocks were altered prior to base-metal ore deposition. Such a sequence of hydrothermal activity provides a possibility that some localization factors for the alteration phase may have differed from those of ore deposition. The observed ore-mineral occurrences, for example, generally are restricted to a vertical range of 2,000 feet along the vein fissures,[11] whereas the enclosing altered wallrocks essentially are unchanged where they are observed along veins above and below the ore horizon. In addition, most base-metal sulfide ore occurs in shoot structures along veins, and there is no indication of any difference between alteration of essentially barren wallrocks and those enclosing ore.

Crosscutting relations between veins, such as those reported at Butte (Sales and Meyer, 1948) which may demonstrate synchronous alteration and metallization stages, are not observed here, and the low-angle vein intersections observed in the central part of the Front Range mineral belt are not informative. Some small-scale vein structure relations also suggest different ages of alteration and metallization. One such example, sketched in figure 69, from the back of a drift in the Houston mine (Kitty Clyde area), shows a change in vein direction which is not reflected in the altered walls. This may be reasonably interpreted as alteration of rock along a vein structure, subsequent reopening along a slightly different path, and finally vein filling by an agent not capable of modifying the walls further.

A regional temperature gradient may be responsible for the ore-mineral zonation[12] and also for the greater amounts of altered rock in the central zone as con-

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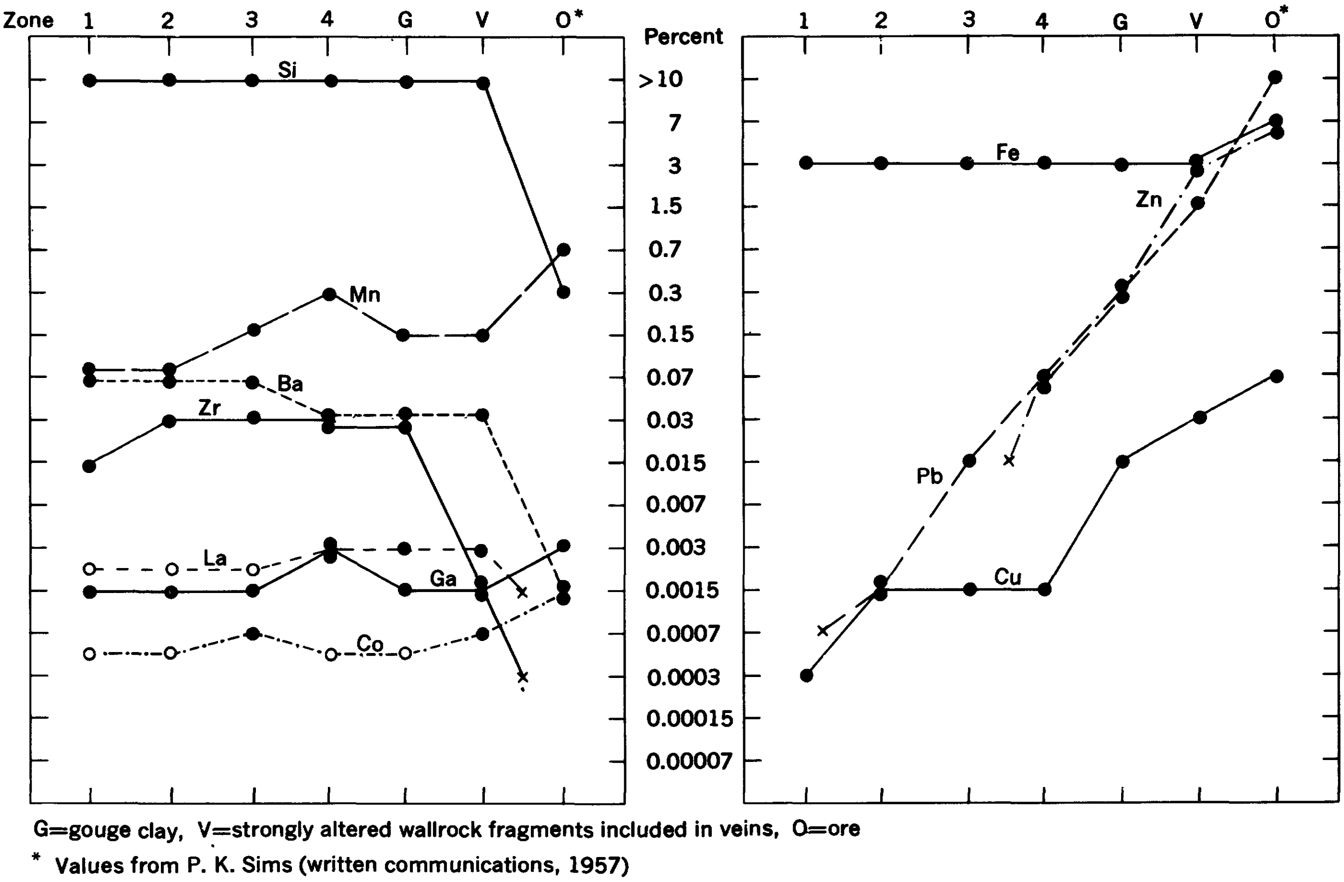
[10] The method for obtaining the powder density determination may in part explain this seeming discrepancy in argillized rock. The gravity of the powdered rock is usually determined in a pycnometer in water solution (L. C. Peck, written communication, 1957). Small errors are possible owing to air bubbles entrapped in a finely ground powder and the possibility that some colloidal material may be lost in the manipulation of the pycnometer. A fundamental error, however, may be caused by the clay minerals in the samples. The water associated with clay minerals is an integral part of the structure, is partially ordered, and is structural water (Grim, 1953). The amount of water adsorbed by a fluffed-out powder may not necessarily be directly proportional to that which can be adsorbed on the same surface in the confined-rock environment. (This problem also is of concern in the analytical determination of H_2O- in argillized rock; it is not possible to judge the proportions of pore water to structural water that may be lost below 100° C.) Because the ground and air-dried clays adsorb water, probably more than that originally held, their density per volume should be reduced or their volume increased. This cannot be evaluated in the computations, so the powder density value is effectively increased, and the calculated pore space is thereby increased. The more clay minerals that are present, especially montmorillonite, the greater this discrepancy may become. The unusually high pore-space values on figure 66 occur in the clay-mineral zones in altered wallrocks.

[11] Harrison and Wells (1956, p. 85) show that many ore shoots in the Freeland-Lamartine district pinch out before reaching the surface (10,450 feet altitude) and that values in the ore decrease with depth—none have been minable below an altitude of 7,900 feet. Similar decreases in ore values were discovered in veins which the Argo tunnel intersected under the Central City district at an altitude of 7,500 feet (Lovering and Goddard, 1950, p. 171).

[12] Sims (oral communication, 1960) has some evidence for a thermal gradient which accounts in part for the areal zonal distribution of ore minerals.



G=gouge clay, V=strongly altered wallrock fragments included in veins, O=ore

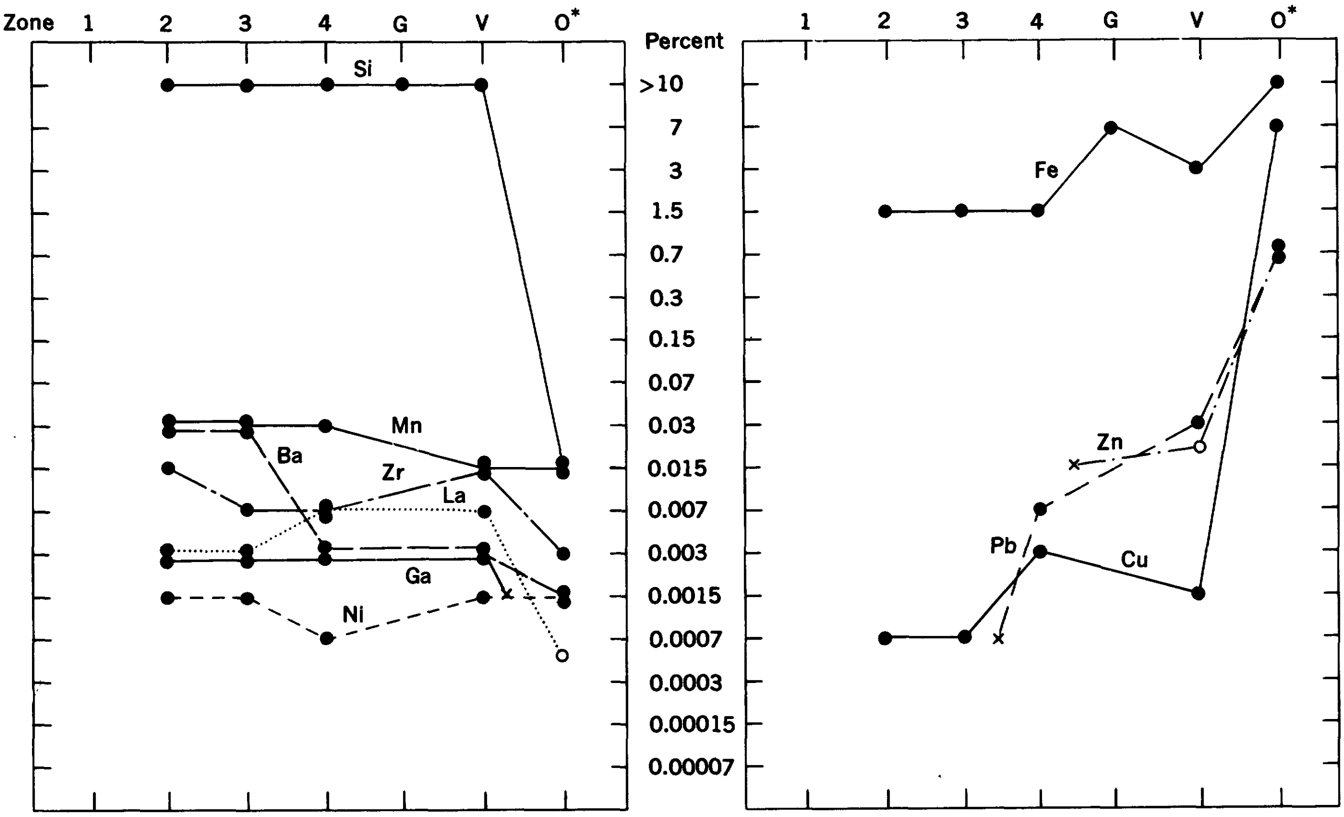
* Values from P. K. Sims (written communications, 1957)

FIGURE 67.—Average values form semiquantitative spectrographic analyses of some elements in microcline-quartz-plagioclase-biotite gneiss wallrocks and in galena-sphalerite ores from Essex mine.

trasted to peripheral areas; but this factor may be modified by structural movements that coincide with the period of alteration-metallization. At Central City, a rough correlation exists between the width of altered wallrock and the size of the vein. Pyrite-bearing veins in the central zone, presumably closer to the source of the solutions, are wider and are bounded by wider zones of altered rock than are galena-sphalerite-bearing veins in the peripheral zones. These relations are not as clear for altered rocks in the southwestern extension of the central zone through Idaho Springs, although the ore-mineral suites are comparable throughout. In addition, variations are evident in the general pattern of wide veins and associated wide altered-rock zones. For example, the galena-sphalerite-pyrite ore in the Bald Eagle vein in the Two Brothers mine (intermediate zone) is as much as 1 foot wide and enclosed by a narrow argillic and moderately wide silicified zone. However, portions of the Lincoln vein in the Alma-Lincoln mine and a subsidiary vein in the E. Calhoun mine (both close to the central-intermediate zone boundary and containing similar ore in comparable rock and rock structures), are 3 inches or less wide and are surrounded by wide silicified and argillized altered-rock halos. These variations in the width of altered zones suggest that the portion of the vein open during an early alteration phase was closed or restricted solution movement or deposition in the subsequent metallization phase. The reverse is also possible.

Cymoid-type zoned vein structures in the Freeland-Lamartine district were interpreted by Harrison (1955) to relate to periodic fracturing accompanied by stages of mineralization. The wallrocks here are altered uniformly where observed horizontally and vertically along the veins even though the veins contain a varied ore mineralogy. The possibility exists therefore, that these fractures are of preore age, their walls were altered prior to ore-mineral deposition, and that subsequent structural readjustments along curved vein-faults rendered different parts of the vein open during the different ore-forming stages.

An important factor in the localization of altered wallrock, and especially argillized rock, is the host-rock mineralogy. Plagioclase- and hornblende-rich rocks such as quartz diorite and amphibolite have a much wider envelope of altered wallrock than the K-feldspar-



G=gouge clay, V=strongly altered wallrock fragments included in veins, O=ore.

*Values from P. K. Sims (written communications, 1957)

FIGURE 68.—Average values from semiquantitative spectrographic analyses of some elements in microcline-quartz-plagioclase-biotite gneiss wallrocks and in pyrite ore from E. Calhoun mine.

rich rocks such as granite or microcline-quartz-plagioclase-biotite gneiss.

THE HYDROTHERMAL ENVIRONMENT

HOST ROCK

The chemistry and mineralogy of a changing host-rock environment in progressively more altered rock are conveniently summarized by means of phase diagrams; however, pressure and temperature relations are not directly observed and are based on less well-established data. Eskola's metamorphic facies principle is very useful to describe and illustrate the metamorphic-rock environment. Creasey (1959) proposes to extend the idea to include the action of hydrothermal solutions on rocks and thereby to describe a hydrothermal-mineral assemblage. Turner (1948, p. 54–59) describes a metamorphic facies as including rocks of any chemical composition that have reached chemical equilibrium during metamorphism under a particular set of physical conditions.[13] The altered-rock type that most closely approaches these requirements is the sericitized rock formed in zone 4 adjacent to the vein; most of the altered rocks we describe do not fully meet Turner's facies requirements.

Triangular composition diagrams used by Eskola relate rock chemistry and mineralogy. Eskola selected Al_2O_3, CaO, and (Mg, Fe)O as three components mainly responsible for the observed mineralogy, and

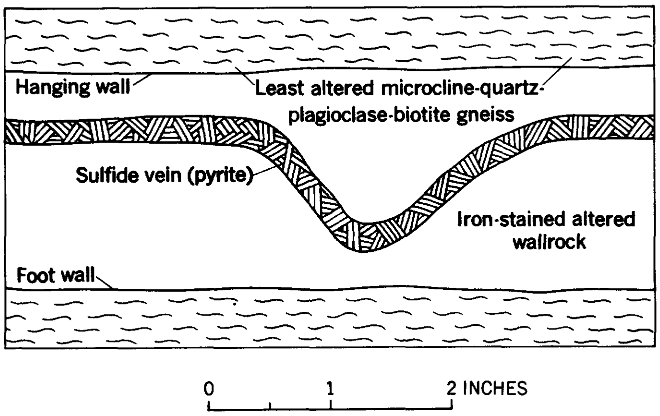


FIGURE 69.—Sketch of small-scale example of vein—altered wallrock relations, Houston mine, Idaho Springs district, Colorado.

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[13] Petrographic evidence for an approach to equilibrium is obtained if it can be observed (1) that the same mineral assemblage develops from different parent rocks under constant physical conditions that have been attained from different directions, (2) that a mineral assemblage is of wide distribution in metamorphic rocks of different ages, (3) that the mineral assemblage conforms to the requirements of the mineralogical phase rule, and (4) that disequilibrium, such as one mineral obviously in the process of replacement by another, is absent.

plotted them on an ACF diagram (Turner, 1948, p. 57). For rocks containing excess Al_2O_3 and SiO_2 (for example, rocks above the anorthite-biotite join on the ACF diagram) Eskola used the AKF diagram (Turner, 1948, p. 82). A plot of the Front Range fresh rock types on an ACF diagram (fig. 70) indicates that quartz diorite, amphibolite, and garnet-quartz gneiss are deficient in Al_2O_3, and that biotite-quartz-plagioclase gneiss and granodiorite are close to the borderline. Biotite-muscovite granite, microcline-quartz-plagioclase-biotite gneiss, biotite-quartz-plagioclase gneiss, and granodiorite, which are the predominant host rocks in the region, are plotted in the AKF diagram. Curiously muscovite is a megascopic mineral only in the granite, and it is assumed that the corresponding microscopic clay mineral (illite or deuteric sericite ?) proxies for muscovite in the other rocks.

As proposed by Creasey, the progressive alteration of rocks by a hydrothermal solution may be compared on an AKF diagram (fig. 71). Ideally field 1 contains a clay mineral-quartz equilibrium assemblage, and field 2 a K-feldspar-sericite-quartz assemblage. K-feldspar, muscovite, and biotite are unaltered in field 1, and K-feldspar and muscovite are recrystallized partly in field 2. Disequilibrium occurs in the Front Range altered rocks, and the triangular diagram is not rigorous; nevertheless, it is useful to show the chemical trends as an equilibrium assemblage is approached in zone 4.

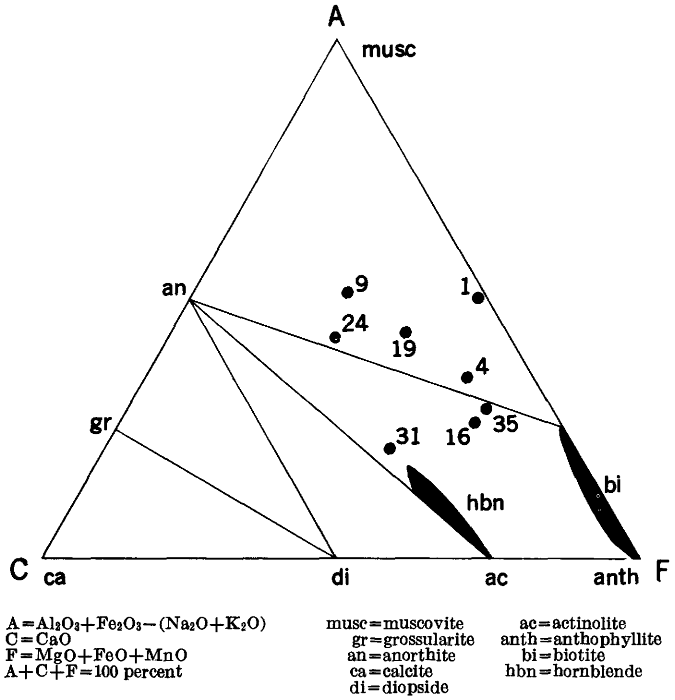


FIGURE 70.—ACF diagram of the amphibolite facies (low temperature subfacies) for rocks with excess SiO_2 and K_2O (Turner, 1948, p. 78) showing plots of fresh rocks from the Front Range area studied. K-feldspar and quartz are additional phases. Biotite-muscovite granite (1), granodiorite (4), microcline-quartz-plagioclase-biotite gneiss (9), quartz diorite (16), biotite-quartz-plagioclase gneiss (19 and 24), amphibolite (31), and garnet-quartz gneiss (35).

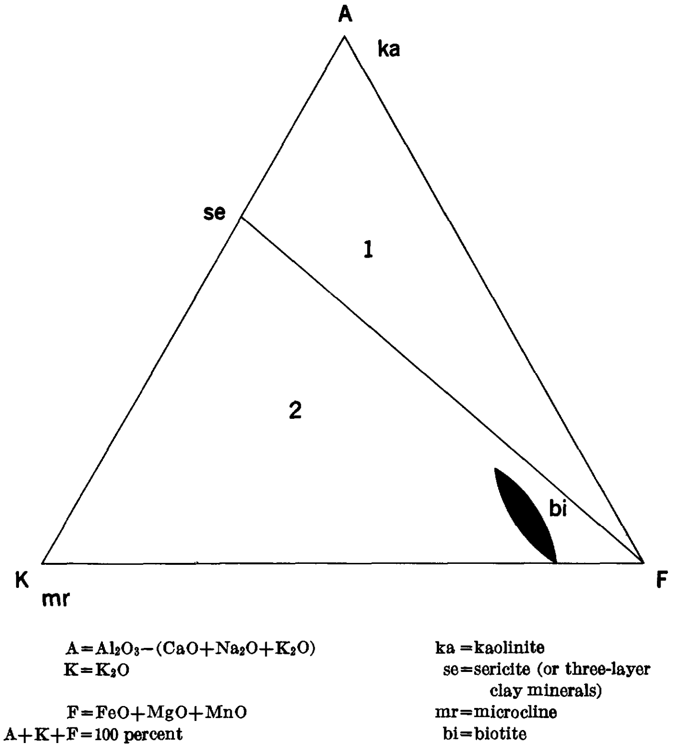


FIGURE 71.—AKF diagram proposed for the hydrothermal assemblage for rocks with excess SiO_2 and H_2O (Creasey, 1959); field 1 contains a lower temperature mineral assemblage than field 2.

The trends of chemical changes (fig. 72) parallel mineralogical changes in progressively more altered wallrocks. Samples 1, 4, 9, 19, and 24, which are fresh rock, are unstable in the hydrothermal environment and are argillized. Weakly argillized rock contains more Al_2O_3 exclusive of that related to CaO, Na_2O, and K_2O in feldspar. Strongly argillized rock is enriched even more in Al. Sericitized rock, however, is enriched in K at the expense of Al. Where pyrite is present in sericitized rock the Fe is also enriched. The alteration of biotite-quartz-plagioclase gneiss (fig. 72, samples 19–23, and 24–26) follows a similar but more variable trend. The samples of biotite-muscovite granite deviate from this pattern in that few clay minerals form in a rock composed predominantly of quartz, microcline, biotite, and muscovite—minerals which are not altered in field 1 and (except for biotite) are stable in field 2.

The loss of alkali and alkaline earth ions, largely from altered plagioclase, shown in figures 28 and 30, results in effective enrichment in Al. Loss of K and Fe is most marked in biotite- and hornblende-bearing rock. In some rocks Fe and Mg are lost gradually; but, in those rocks in which biotite alters to magnetite

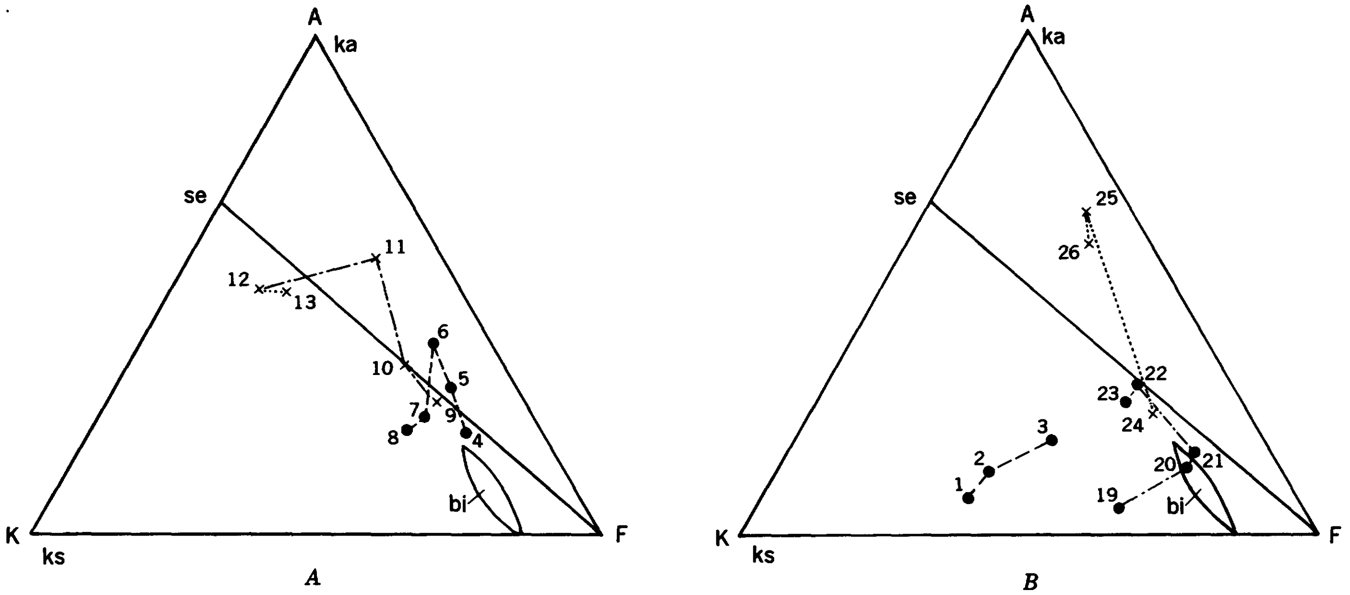


FIGURE 72.—AKF diagrams *A* and *B* show plots of fresh and altered wallrocks in the central part of the Front Range mineral belt: biotite-muscovite granite, 1–3; granodiorite, 4–8; microcline-quartz-plagioclase-biotite gneiss, 9–13; and biotite-quartz-plagioclase gneiss, 19–23, and 24–26. Quartz is an additional phase in all samples and anorthite and various other minerals are additional phases in the fresh rock which is shown only for reference. Rocks of zone 2 and 3 also contain irregular amounts of these partially altered phases. Lines connect related analyses and do not necessarily indicate intermediate paths; microcline (ks), biotite (bi), sericite (se), and kaolinite (ka).

and to chlorite, Fe and Mg do not escape readily from the host rock.

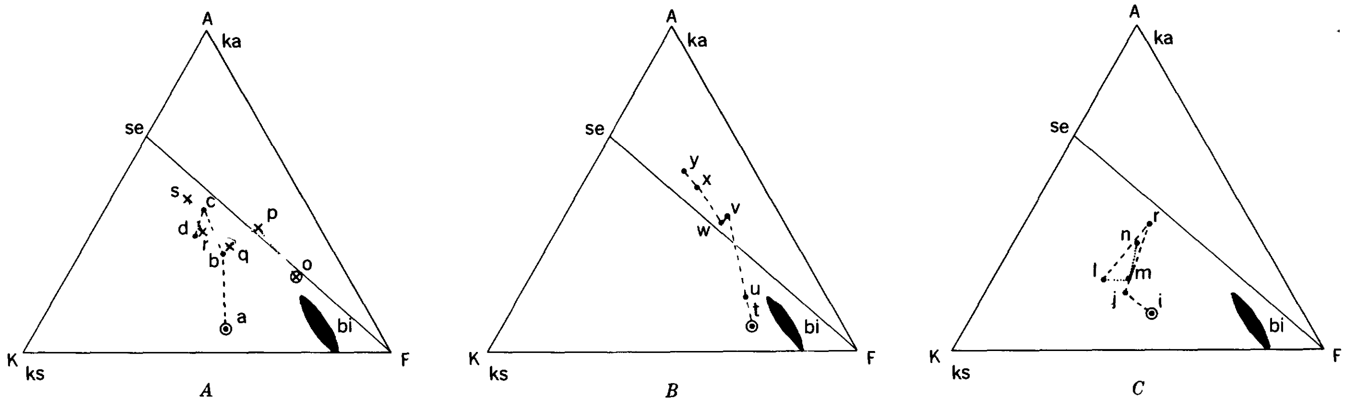
These results conform in a general manner with those from diverse rocks cut by vein deposits of similar type elsewhere. The AKF diagram plots (fig. 73) for fresh and altered Boulder Creek granite, syenite from the Caribou stock in Colorado, and for quartz monzonite at Butte district, Montana, indicate a trend toward Al during argillization and, except for Butte where no marked increase in K is reported, a trend toward K with more intense sericitization.

AKF diagrams suggest that possibly only the final sericitized rock assemblage is in equilibrium with the altering environment, and that the paths of chemical changes in the host rock environment are dependent on the composition and quantity of unstable minerals. Although the paths may vary, the end product of hydrothermal alteration of Al_2O_3-rich rocks and their equilibrium assemblage is a regular feature—a narrow zone parallel to the muscovite-Fe join. Not shown, but indicated by the mineralogy of altered Al_2O_3-deficient rocks is their high-intensity assemblage of three-layer clay minerals, kaolinite, and quartz, which is comparable to the nonequilibrium transitory field 1 assemblage noted above.

HYDROTHERMAL SOLUTIONS

The complex nature of the hydrothermal solutions that altered the wallrocks and formed ore deposits cannot be deduced completely from wallrock data alone; however, these data and those obtained from surface hydrothermal and meteoric waters provide partial insight into the hydrothermal environment in the area studied. First we shall define hydrothermal (after White, 1957a, p. 1638, and 1957b, p. 1661) as pertaining to any water that is relatively warmer (plus 5° C) than its enclosing environment. In this region hydrothermal solutions are inferred to be mixtures of waters of deep origin (for example, juvenile or magmatic water) which are diluted and contaminated nearer the surface by meteoric and metamorphic water. Thus we may infer also, as does Korjinsky (1936), that a moving solution and immobile interstitial pore and bound water solution are mutually involved in the continuous interchange of matter by diffusion of ions and molecules along a concentration gradient.

The classical idea (Kerr, 1955) that acid emanations react with wallrock at depth to produce neutral to alkaline solutions and that these may revert to acid character by oxidation at or near the surface is borne out within the limits of observation in ground-water and hot-spring data at Idaho Springs, Colo. (table 31), and where drilled to a moderate depth at Wairakei, New Zealand (Grange, 1955; Steiner 1953). The radioactive warm springs at Idaho Springs may represent late stages of a long period of hydrothermal activity in the region, and are, as they issue at the surface, worked-out, oxidized, diluted hydrothermal solutions. These solutions are bicarbonate-, sulfate-, and chloride-bearing. In addition they carry abundant Na and Si as



A

a, Fresh Boulder Creek granite, 5th level, Coldspring mine, 35 feet from vein: Quartz, 15–25 percent; orthoclase-microline, 20–30 percent; andesine-oligoclase, 20–30 percent; biotite and hornblende, 10–20 percent; ilmenite and apatite, calcite and sericite minor constituents.
b, Argillized granite (montmorillonite-rich) 10 feet from vein.
c, Argillized granite (kaolinite-rich) 3 feet from vein.
d, Sericitized granite one-half inch from vein.
o, Fresh wallrock.
p, Kaolinite zone.
q, Orthoclase zone.
r, Hydromica zone.
s, Vein.

B

t, Fresh quartz monzonite, 2800 level, a Northwest vein. Values approximate as scaled from diagram and recalculated.
u, "Green" zone (montmorillonite-rich).
v, Boundary "green" zone and kaolinization zone.
w, Kaolinization zone.
x, Sericitization zone.
y, Boundary sericitization zone and vein.

C

i, Fresh syenite, middle section, 1040 level, Caribou mine, 1.7 feet into hanging wall: Orthoclase-plagioclase, 65–75 percent; augite, 10 percent; biotite, 10 percent; quartz <5 percent; apatite, pyrite, sphene, accesories.
j, Weakly altered syenite (montmorillonite-rich) 1 foot into hanging wall.
k, Moderately altered syenite (kaolinite-montmorillonite-rich) 0.6 feet into hanging wall.
l, Intensely altered syenite (sericite-rich) 0.2 feet into hanging wall of vein 0.5 feet thick.
m, Moderate-intense altered syenite 2.2 feet into footwall.
n, Intensely altered syenite 0.6 feet into footwall of vein.

FIGURE 73.—AKF diagrams show plots of fresh and altered wallrocks in (*A*) granite, Nederland district, Colorado (a to d, Lovering, 1941, p. 242; and o to s, Gonzalez-Bonorino, 1959, p. 74), (*B*) quartz monzonite, Butte district, Montana (Sales and Meyer, 1948, p. 30), and (*C*) syenite, Caribou district, Colorado (Wright, 1954, p. 133). Quartz is an additional phase; microcline (ks), biotite (bi), kaolinite (ka), and sericite (se).

TABLE 31.—*Analyses of water samples, in approximate parts per million, from springs near Idaho Springs, Colo.*

Constituent or analysis factor	Sample			
	a	b	c	d
SiO_2	68.0	62.0	55.2	72.9
SO_4	396.3	310.1	223.17	412.35
HCO_3	1,513.8	1,293.64	835.15	1,467.05
CO_3	0	0	0	----------
PO_4	0	0	0	----------
Cl	71.47	73.47	83.49	72.0
Fe	----------	43.75	----------	7.35
Al	----------	----------	----------	----------
Fe_2O_3 Al_2O_3	2.0	84.45	0	10.5
Mn	0	0	0	----------
Ca	145.4	120.64	89.5	145.44
Mg	39.43	8.63	23.21	33.86
K	Tr.	63.24	48.8	----------
Na	573.3	419.04	300.76	561.25
Li	0	0	0	----------
Total	2,478.96	2,466.22	1,659.18	2,782.70
H_2S	0	2.35	0	----------
As	----------	----------	----------	----------
Sr	----------	Tr.	----------	----------
Excess CO_2	545.8	466.48	301.1	14.7
Fe ppt	2.38	1,770.00	3.76	----------
Dissolved solids	2,045.00	15.26	1,244.00	----------
Temp °F	102	58	61	[1] 115 112
Rate of flow gpm	35–50	----------	10–12	----------
pH	----------	----------	----------	[2] 6.5

[1] Measured by W. Niles, U.S. Geol. Survey, Apr. 19, 1956.
[2] Measured by D. Skinner, U.S. Geol. Survey, Apr. 19, 1956.

Sample data:
- a. Hot Soda Spring, No. 85 (Radium Hot Springs?); analysts: H. A. Curtis, R. M. Butters, P. M. Dean and H. R. Mosley (George and others, 1920).
- b. Cold Soda Spring, No. 86; analysts: H. A. Curtis, R. M. Butters, P. M. Dean and H. R. Mosley (George and others, 1920).
- c. Blue Ribbon Spring, No. 87; analysts: H. A. Curtis, R. M. Butters, P. M. Dean and H. R. Mosley (George and others, 1920).
- d. Radium Hot Springs, north end of Radium Hot Springs Hotel; analysts: N. F. Witt and W. B. Peitenpol.

well as a minor amount of Fe^{+3} and Al. One analysis reports H_2S and correspondingly contains more Fe ions. Temperature at the radioactive springs range from 58° to 115° F, and the pH of the Radium Hot Springs was 6.5. The more active nearly boiling springs at Wairakei, have 270° C temperature measured in drill holes at 2,000 feet (Grange, 1955, p. 53), and are of the sodium-chloride type (White, 1957b, p. 1665). Ellis (1955) reports a general increase of pH with depth from 2.5 to 8.0 where drilled to 100 feet and estimates from chemical equilibria that a pH of 9.2 is required at 2,000 feet.

Some mine waters in the Front Range region (Clarke, 1924, p. 644, 646) are slightly acid; however, pH measurements (table 32) of mine water throughout the region shows that moderately to slightly alkaline water occurs in may mines in the intermediate and peripheral ore zones. Ground water at the Running Lode mine (Headden, 1903, p. 179) contains sulfate-carbonate-chloride components as well as moderate to minor amounts of dissolved alkali and alkaline earth ions, silica, iron-alumina, and manganese oxide. In contrast, extremely acid water (tables 33 and 34) from the Argo tunnel, which taps water from mines where pyrite is highly oxidized, contains high sulfate, low halogen ion concentration, moderate amounts of alkali, alkaline earths, silica, aluminum, and iron, and trace amounts of other ions. Ferric iron is more abundant than ferrous iron. The acid waters (table 34) carry more Cu, Zn, Mn, and Fe, and also carry Co, Ti, Y, and Yb which are not found in near-neutral meteoric and ground water (George and others, 1920). These are mostly supergene effects and therefore should not be confused with the hypogene environment.

Data concerning pH must be extrapolated to depth with caution because, as Hemley (1959) indicates, reactions which required low pH at high temperatures may be accomplished at higher pH values at lower temperatures.

A mild hydrothermal solution environment postulated from mineral assemblages may have a range of actual temperatures, but the temperature in any one place was probably only moderately above that of the wallrock. Host-rock mineral structures are not completely destroyed by such solutions; K-feldspar grains are unaltered, quartz is partly recrystallized, and most other silicate minerals are argillized. The sericite zone produced on the vein wall is thin, and leaching here of all elements except silicon and oxygen (thereby creating a porous mass) was not observed.

Synthesis data demonstrate that minerals can be formed artificially in the presence of a solution within certain temperature, pressure, and composition limits. Most of the critical wallrock mineral assemblage may be formed between 100° and 500° C. Some of the extensive laboratory findings are reported by Folk (1947), Ingerson (1955a), Yoder and Eugster (1955), Morey and Chen (1955), and Brindley and Radoslovich (1956). Possibly the most appropriate study for our consideration is that by Hemley (1959) which was concerned specifically with the hydrothermal environment and relates temperature and composition of ore solutions and wallrocks. A lateral temperature gradient for formation of sphalerite in veins from 620° C, in the central ore zone, to about 380° C, in the peripheral ore zone, has emerged from studies by Sims and Barton (Sims, written communication, 1960).

The chemical relations of these wallrocks imply that the altering solutions were slightly acid[14] and slowly differentiating at a source so that their chemistry changed with time. Physical and chemical gradients initially favored migration of H ions a moderate distance into slightly alkaline walls. Ions with low ionic potential were replaced, moved, and added to vein solutions. Solution acidity would thus be reduced upvein initially, but as all attainable fresh wallrock was

TABLE 32.—*Hardness values of mine water in the central part of the mineral belt*

Area and ore zone	Mine	Hardness (pH)
Central City district:		
Central (pyrite) zone	Bobtail	3.4
	Argo (drainage)	2.6
Intermediate zone	Widow Woman	6.3
	Banta Hill	6.7
Peripheral (galena-sphalterite) zone	Carroll	6.8
	Two Sisters (near R.H.D.-McKay shaft).	6.9
Idaho Springs district:		
Intermediate zone	Stanley (near Alma-Lincoln mine).	7.4
	Alma Lincoln	7.4, 7.7
	Sunnyside (near Phoneix mine).	7.7, 7.9
Lawson-Dumont-Fall River district:		
Peripheral zone	Jo Reynolds	8.2
	Watt-Stemple (near Almadin mine).	7.6

TABLE 33.—*Analyses of water samples, in parts per million, from Clear Creek and the Argo tunnel, Idaho Springs, Colo.*

[Samples collected by Lyman Huff, U.S. Geol. Survey; Analyst: F. J. Flanagan, U.S. Geol. Survey.]

Constituent or analysis factor	Sample	
	9	10
Silica (SiO_2)	7.0	38
Aluminum (Al)	.4	30
Iron (Fe^{+2})	.01	25
Iron (Fe^{+3})		175
Manganese (Mn)	1.0	60
Copper (Cu)	.4	16
Zinc (Zn)		5
Calcium (Ca)	18	350
Magnesium (Mg)	7.4	125
Sodium (Na)	5.3	27
Potassium (K)	1.8	5.4
Carbonate (CO_3)	0	0
Bicarbonate (HCO_3)	36	0
Sulfate (SO_4)	54	2,300
Chloride (Cl)	.8	7.0
Fluoride (F)	.6	4.4
Nitrate (NO_3)	1.0	.9
Dissolved solids	117	3,460
Specific conductance ($K \times 10^5$ at 25° C.)	19.3	364
Color	1	70
pH	6.7	2.65
Acidity (as H_2SO_4):		
Total		1,040
Free		130

Localities where samples were collected:
9. Clear Creek, Idaho Springs, Colo., just above Argo tunnel.
10. Drainage from Argo tunnel at mouth, Idaho Springs, Colo.

TABLE 34.—*Qualitative spectrographic examination of water residues of samples from Clear Creek and the Argo tunnel, Idaho Springs, Colo.*

[Samples collected by Lyman Huff, U.S. Geol. Survey. Analyst: K. J. Murata, U.S. Geol. Survey]

	Elements in sample—	
	9	10
Major elements	Na, Ca, Mg	Ca, Mg, Mn, Fe, Al.
Minor elements	Mn, Si	Na, Cu, Zn, Si.
Traces	Pb, Cu, Zn, Fe, Al, Ni, V, Ba, Sr.	Y, Yb, Ni, Co, Ti, Ba, V, Sr.
Not found	Be, Ti, Co, Y, Yb, plus others.[1]	Pb, Ba, plus others.[1]
Total solids (g per gal)	0.484	12.257

[1] Looked for but not found: Tl, As, Sb, Mo, W, Ga, Ag, Bi, In, Cd, Cr, Re, La, Zr, Cb, P.

Localities where samples were collected:
9. Clear Creek, Idaho Springs, just above Argo tunnel.
10. Drainage from Argo tunnel at mouth, Idaho Springs, Colo.

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[14] Lovering (1950, p. 235) states, "* * * the pH range usually considered of geologic interest, say from pH 7 to a pH of 1 (tenth-normal HCl has a pH of 1.1, one-normal HCl has a pH of 0.1), is very small—from 428 parts per million to about 422 parts per million in HCl."

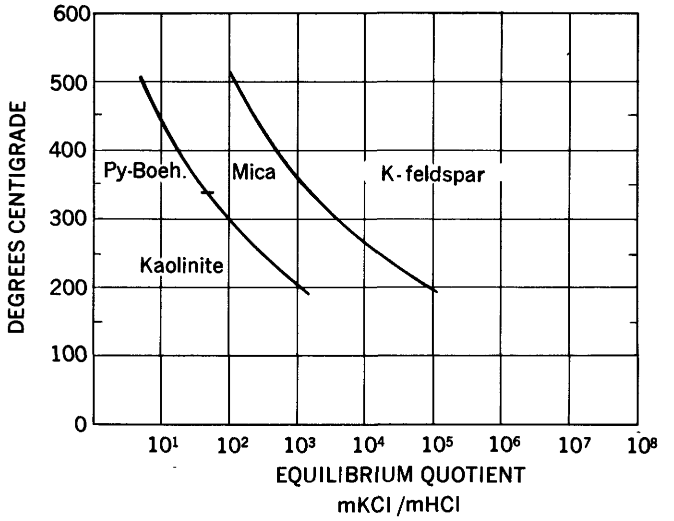

FIGURE 74.—Experimental stability fields of clay mineral-mica-feldspar in the K_2O-Al_2O_3-SiO_2-H_2O system after Hemley (1959 p. 246). Py, pyrophyllite; Boeh, boehmite.

altered down vein and temperature of the solutions increased, the acid solutions rose higher up the vein system. Na, Ca, Mg, Si, Al, and Fe must all have been deficient in the solution initially, and gradually were enriched. K ions were present in the altering solution during its later evolution, but probably were sparse in early differentiates because potassium was leached from unstable minerals in certain plagioclase-rich rocks. Sulfur also appeared late and combined with iron during the K ion-rich phase of rock alteration. The textural position of calcite in fractures implies that CO_2 ions appeared at still a later phase of rock alteration. Under strong hydrothermal alteration conditions all of these delicate yet significant features are obliterated.

The early solution which penetrated and argillized the walls either was incapable of depositing Cu, Pb, Zn, and Ag in argillized rock, or, in the early stages of differentiation of the solutions at the source, these elements (which are subsequently concentrated in the veins) were not present. (Sparse data indicate that the later telluride depositing solution also did not penetrate and deposit in the walls.) Semiquantitative spectrographic data show that the sericite zone does contain minor amounts of these elements above normal rock background. Either sericite-forming solutions contained these ions in addition to K and S and incorporated them, or they result from subsequent leakage from the vein along small fractures. If they were present in a sericite-pyrite forming phase, disseminated copper, lead, or zinc sulfides should have been observed.

The stability fields of clay mineral, mica, and feldspar in the K_2O-Al_2O_3-SiO_2-H_2O system (Hemley, 1959, p. 246), reproduced in figure 74, relate variations in the equilibrium quotient mKCl/mHCl and temperature. If one assumes near isothermal conditions (Lovering, 1950), variations in the K^+/H^+ ratio of pore fluids may account for most mineral changes. Similar alterations are possible, holding the ratio constant and changing temperature. The absence of pyrophyllite and boehmite in the altered rocks has two possible explanatons. If formed under the temperature and equilibrium conditions suggested by altered wallrock mineral assemblages and sphalerite geothermometry, the pyrophllite inverted to kaolinite on subsequent cooling. Hydrothermal pyrophyllite, however, occurs along faults (Ehlmann and Sand, 1959) and some relict pyrophyllite, if present here originally, should have been discovered. If, however, the K^+/H^+ ratio remained within a restricted range kaolinite, mica (sericite and three-layer clay minerals), and K-feldspar could form with increased temperature, and the pyrophyllite field be bypassed. K-feldspar is a metastable mineral in many of the host rocks studied, but almost no secondary K-feldspar was observed in altered wallrock in this area. It was reported by González-Bonorino (1959) in certain Boulder County ore deposits.

It is the writer's conclusion, based on these data, that a vanguard solution argillized the walls that were subsequently sericitized, pyritized, and partly silicified. Lastly, the solutions filled the veins with ore and gangue minerals. Possibly these distinct steps happened at the same time in different levels of the system, but not at one time on one level. Therefore these solutions and vein structures more closely resembled those which produced alteration by "stages" (Lovering, 1941, 1950; Lovering and others, 1949) than those which produced simultaneous fronts (Sales and Meyer, 1948, 1949).

Movements of supergene solutions and normal surface water reached lower with time and erosion, and also acted on the system at a much later time. The fluids were near neutral in the intermediate and peripheral zones and thus did not greatly modify the alteration mineral assemblages. Strongly acid mine waters in the central ore zone have caused minor cation exchange, leaching, physical transportation of rock materials, and thin replacement by limonite and chalcocite.

ALTERED WALLROCKS AS A GUIDE TO ORE

In general, the altered wallrocks of the central part of the Front Range mineral belt cannot be considered to be a useful guide to base-metal sulfide ores. No particular wallrock alteration mineral assemblage accompanies ore. The width of the envelope of altered wallrock about veins outside of the central and intermediate zones at Central City is generally small, measured in a few inches to a few feet. Schwartz (1955,

p. 316) pointed out that "very narrow zones of alteration are of limited help in searching for ore because finding the altered rock is about as difficult as finding the ore." The chances are best that altered rock may be a useful guide if the alteration and metallization processes are nearly contemporaneous. In these areas most rock alteration must have preceded ore formation. Postalteration, but preore strike and perhaps some dip-slip movements on curved vein faults rendered zone faults wholly or partly closed during one phase or the other, thus wide ore-filled veins may adjoin narrow altered rock zones, and often do. However, one may show that there is a local correlation between wide veins and widest altered rock zones in the central and intermediate ore zones in the Central City district.

Whereas little correlation is evident between location and width of base-metal sulfide ore shoots and the location and width of altered granitic wallrocks, altered wallrocks such as amphibolite and garnet-quartz gneiss that contain large amounts of iron-bearing minerals and small amounts of quartz and microcline have a spatial coincidence with some pitchblende deposits in the veins. Regardless of the source, the uranium may have been deposited in open spaces as a consequence of reaction with excess iron in solution that possibly was derived from rock alteration.

The close paragenetic relations between pitchblende and early pyrite (Sims, 1956) strongly suggest that the hydrothermal solution of the late and most intense alteration phase, which altered biotite and provided a source of iron for subsequent early pyrite formation, locally carried uranium. The relation between iron-bearing wallrock and pitchblende metallization also has been noted under slightly differing environment conditions in the Front Range by Adams and Stugard (1956) and Hawley and Moore (1955). Secondary uranium minerals, such as metatorbernite, also appear to have an affinity for the iron-bearing clay mineral (illite-montmorillonite) structures in altered amphibolite and biotite-quartz-plagioclase gneiss (Sims and Tooker, 1955).

Whether this relationship will be useful in finding uranium ore is questionable from the foregoing comments, but this study of alteration mineralogy within a geological framework provides an answer to the seemingly irrational distribution of uranium ore deposits in a well-zoned area of sulfide ore deposits.

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