Geology and Ore Deposits of the Jerome Area Yavapai County Arizona

By C. A. ANDERSON *and* S. C. CREASEY

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With sections on the United Verde Extension mine by G. W. H. Norman and on the Cherry Creek mining district by R. E. Lehner

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GEOLOGY AND ORE DEPOSITS OF THE JEROME AREA, YAVAPAI COUNTY, ARIZONA

By C. A. Anderson and S. C. Creasey

ABSTRACT

The Jerome area, in central Arizona, includes the Mingus Mountain quadrangle and parts of the adjacent Clarkdale, Mayer, and Mount Union quadrangles. The largest copper mines in the area at Jerome are the United Verde Extension and the United Verde. The United Verde Extension, closed in 1938 and the United Verde closed in 1953. The Iron King mine at Humboldt, the most active mine in the area after 1953, produces lead and zinc.

The oldest rocks are the older Precambrian Yavapai series, consisting of metamorphosed volcanic and sedimentary rocks. The Yavapai series is divided into the Ash Creek and Alder groups. The Ash Creek group is divided into seven formations, consisting of basaltic, andesitic, rhyolitic, and dacitic flows and pyroclastic rocks, and a thick sequence of tuffaceous sedimentary rocks containing interbeds of jasper-magnetite and chert. Rocks of the Ash Creek group generally are nonfoliated and relict structures are well preserved; it is easy to distinguish the top and bottom of the lava flows or beds. The Ash Creek group is about 20,000 feet thick. The Alder group is separated from the Ash Creek group by the Shylock fault; the age relationship of the two groups is unknown. The Alder group is divided into six formations, consisting of basaltic, andesitic, and rhyolitic flows and pyroclastic rocks and a sedimentary sequence of slate. The rocks in the Alder group are strongly foliated and the stratigraphy and structure of the group are in doubt. The thickness of the Alder group is not known, but is apparently at least 20,000 and perhaps 30,000 feet.

Intense deformation of the Yavapai series was accompanied by the intrusion of quartz porphyry that is locally foliated. Later, gabbro and then quartz diorite were intruded. The quartz diorite is one of the several intrusive quartz-bearing granitoid rocks that comprise the Bradshaw granite batholith of central Arizona. Swarms of granodiorite porphyry dikes, apparently related to the quartz diorite, intrude the older intrusive rocks and Yavapai series. All these intrusive rocks are of older Precambrian age.

A profound unconformity separates the older Precambrian rocks from the overlying Paleozoic sedimentary rocks. The basal Paleozoic sandstone, deposited on a surface of low relief, has been tentatively correlated with the Tapeats sandstone of Cambrian age exposed in the Grand Canyon. The Tapeats(?) in the Jerome area is overlain with apparent conformity by the Martin limestone of Devonian age, and the Mississippian Redwall limestone overlies the Martin. About 370 feet of the basal part of the Supai formation of Pennsylvanian and Permian age overlies the Redwall in the Jerome area, and to the north and east, the full thickness of the Supai is exposed. About 1,200 feet of Paleozoic rocks are exposed in the Jerome area.

Regional evidence indicates that during Late Triassic, Jurassic, Cretaceous, and early Tertiary time, the Jerome area was a site of active erosion, and a maximum of 3,000 to 3,500 feet of Paleozoic sedimentary rocks and an unknown thickness of Precambrian rocks were removed. By late Tertiary time, an erosional surface of 500 feet or more of relief became the site of deposition of the Hickey formation—1,400 feet of dominantly basaltic lava flows on Mingus Mountain. But along the south margin of the area, a thinner section of intercalated flows and sedimentary rocks is exposed.

After the accumulation of the Hickey formation, faulting blocked out the Black Hills and Verde Valley to the east. The Verde formation, consisting chiefly of calcareous lake deposits as much as 2,000 feet in thickness, accumulated in the Verde Valley.

The dominant structure of the Ash Creek group is the Mingus anticline, essentially domical, plunging northwestward near Jerome and southeastward at Black Canyon. Near Jerome, foliation is locally intense, essentially parallel to the axial planes of the minor folds. Warping of the foliation and axial planes of the minor folds south of Jerome indicates that the Mingus anticline has been warped to a sigmoid pattern and displaced by the eastward-trending Oak fault. In the southern part of the Ash Creek terrane, foliation is discordant to the axial planes of the folds, and may date from the sigmoid warping of the anticline.

The Texas syncline, southwest of the Mingus anticline, plunges southeastward except near the Shylock fault where the plunge flattens appreciably.

Two sets of faults are in the Ash Creek group; the older set strikes approximately east and the younger north. Some of the northward trending faults are occupied by dikes of granodiorite porphyry.

Deformation of the Alder group was intense; the structures are complex and difficult to interpret. The rocks are isoclinally folded and intensely foliated. Possibly as many as three periods of deformation, each with a characteristic trend, were recognized locally in the southwestern part of the Alder terrane.

In the south-central part of the Jerome area, the symmetrical distribution of formations indicates a fold that is interpreted as a southward-plunging isoclinal syncline with an associated anticline to the west. Another isoclinal fold is suggested by the outcrop pattern of the units that compose the Chaparral volcanics. Elsewhere the uniformly steep dip of beds and the variation in the direction that the tops of beds face indicate many isoclinal folds.

Distributive shearing characterizes the Alder group. Locally the shearing apparently resulted in little more than weak foliation. Elsewhere, such as along the eastern exposures of the Alder group and in the Chaparral volcanics, intense distributive

shearing produced broad fault zones in which each lithologic unit is in fault contact with the adjacent unit, and is thinned greatly. The Spud and Chaparral faults bound the Chaparral volcanics in the southwest corner of the Jerome area.

The major Shylock fault which separates the Ash Creek and Alder groups has a minimum stratigraphic throw of 20,000 feet. Locally, it separates quartz diorite from slate of the Alder group, proving some displacement after the intrusion of the quartz diorite. The basal Paleozoic sandstone (Tapeats?) covers the trace of the fault proving that the fault is of Precambrian age.

The Paleozoic and Cenozoic rocks have been tilted gently and faulted. Faulting occurred during three periods: (1) after the close of the Paleozoic but before deposition of the Hickey formation; perhaps this faulting is of Late Cretaceous and early Tertiary age; (2) after the accumulation of the Hickey formation; and (3) after the deposition of the Verde formation. On the northwestern flank of Mingus Mountain, faulting of the first period resulted in the uplift and erosion of the rocks on the west side. By the time of the deposition of the Hickey formation, this faulting and erosion caused lava flows of the Hickey to rest on the Precambrian rocks west of the fault and on Paleozoic rocks east of the fault. Faulting during the second period reversed the relative displacement and the western block was downthrown.

The Verde normal fault on the east side of the Black Hills displaces the Paleozoic rocks and Hickey formation; at Jerome, the vertical separation is about 1,500 feet. The fault also produced about 1,000 feet of vertical separation during the Precambrian, and an unknown but small displacement after the deposition of the Verde formation.

Mining in the Jerome area started about 1870, at which time gold and silver were the chief metals produced. Copper mining of the surface oxide ores started at Jerome in 1883, from which copper, gold, and silver were recovered. W. A. Clark obtained control of the United Verde mine in 1888 and under his guidance successful mining operations were inaugurated. The rich United Verde Extension ore body was discovered in 1914, and these two mines produced more than 99 percent of the total production of the Verde (Jerome) district. Early production at the Iron King mine was in gold and silver, but after a period of inactivity, lead-zinc sulfide ore was mined starting in 1936. By 1952 production had gradually increased to 600 tons daily.

The following approximate amounts of metal have been recovered from the Jerome area: gold, 1,890,000 ounces; silver, 65,600,000 ounces; copper, 1,868,000 tons; lead, 37,725 tons; and zinc, 127,000 tons.

The ore deposits of Precambrian age are of two types: (1) massive sulfide ore, and (2) fissure veins. Most of the metal production has come from the massive sulfide deposits in the United Verde and United Verde Extension copper mines at Jerome, and the Iron King lead-zinc mine at Humboldt. The fissure veins have been mined chiefly for their precious-metal content, such as the gold-quartz veins in the Cherry Creek district and silver-bearing vein of the Shea mine south of Jerome. Copper has been the chief metal from the Verde Central and Yaeger veins. The McCabe-Gladstone vein, west of the Iron King mine, was an important producer of gold and silver, but it cannot be proven that this deposit is of Precambrian age. At Jerome, some copper has been mined from Tertiary lava flows and gravels impregnated with supergene chrysocolla.

The massive sulfide deposits are tabular, lenticular, or pipelike. They are dominantly pyritic and contain variable amounts of chalcopyrite, tennantite, sphalerite, and galena. Alteration of the host rock consists of silicification and sericitization, and at the United Verde mine, chloritization. The writers conclude that the massive sulfide deposits are replacements of the schistose rock.

The massive sulfide body at the United Verde mine was localized in a north-northwestward plunging anticline, intruded by a semiconcordant body of gabbro. Mineralizing solutions were channeled by an "inverted trough" of nonfoliated gabbro. They replaced foliated tuffaceous sedimentary rocks and quartz porphyry with pyrite thus forming a pipelike body of massive sulfide that conforms to the plunging anticline. The country rock on the footwall side (south) of the pyritic pipe was extensively chloritized before the copper was introduced, forming ore shoots in the quartz porphyry, chloritized country rock ("black schist"), and the massive pyritic pipe.

The United Verde Extension ore body is buried beneath Tertiary and Paleozoic rocks to the east of the Verde fault, whereas the United Verde mine west of the Verde fault was exposed at the surface. Much of the ore from the United Verde Extension mine was high-grade chalcocite, but some mixed chalcocite-cuprite-malachite ore came from the higher levels, and some fluxing primary sulfide ore came from the lower levels. The chalcocite ore was formed by secondary enrichment before deposition of the Tapeats sandstone(?).

For 30 years or more the problem has been debated whether or not the United Verde Extension ore body is the downfaulted segment of the United Verde pipe. The solution of this problem depends on accurate determination of the displacements on the Verde fault. Through underground openings in the United Verde and United Verde Extension mines, the trace of geologic contacts can be determined on both the hanging-wall and footwall fault surfaces of the Verde fault. To match the contacts of the Precambrian rocks, about 2,500 feet of vertical separation is necessary, but only 1,500 feet of vertical separation is necessary to match the contacts of the Paleozoic and Tertiary rocks. This smaller displacement of younger rocks indicates that the Precambrian rocks were displaced about 1,000 feet before the deposition of the Paleozoic rocks. With due allowances for possible horizontal separation during both periods of movement, the United Verde Extension ore body could not be the severed top of the United Verde ore body unless the United Verde ore body shifted eastward at some elevation above the present erosion surface, and thus did not follow upward along its known axis (N.20°W., 65°N.).

The Iron King lead-zinc deposit consists of a system of 12 westward-dipping massive sulfide veins that are en echelon in a mylonitic sheared zone. All veins plunge northward, 55° to 60°. The vein material consists of banded fine-grained massive sulfides containing massive quartz at the north end of each. Sulfide vein minerals are pyrite, arsenopyrite, sphalerite, galena, chalcopyrite, and tennantite; nonsulfide minerals are ankerite, quartz, sericite, and a little residual chlorite. Banding in the massive sulfide ore resulted from deposition of ore minerals, chiefly sphalerite, in fractures in early vein filling, and from variation in relative rates of deposition of vein minerals. Gold and silver are most abundant at the north end of the veins, and southward to the point where the maximum content of lead and zinc occurs. The plunge of the north ends of the veins parallels the mineral streaking (lineation) in the wall rocks. This parallelism suggests that the sheared zones that localized the veins were part of the regional deformation which produced the general fabric of the Precambrian terrane.

INTRODUCTION

LOCATION, CULTURE, AND ACCESSIBILITY

The Jerome area is in central Arizona in eastern Yavapai County (fig. 1). The area mapped for this report includes the Mingus Mountain quadrangle bounded by longitude 112° to 112°15′W., and latitude 34°30′ to 34°45′N. (pl. 1). A small part of the Clarkdale quadrangle to the north was also mapped, and parts of the Mayer and Mount Union quadrangles to the south and southwest. The southeast corner of the Prescott quadrangle, west of the Mingus Mountain quadrangle, was mapped contemporaneously by M. H. Krieger and W. E. Bergquist and has been included in this report (pls. 1, 2).

The two largest mines in the area, the United Verde and the United Verde Extension, are at Jerome, near the center of the boundary of the Mingus Mountain and Clarkdale quadrangles. The purpose of this study was to obtain the regional geologic setting for these two important copper mines, and to study in detail the United Verde mine, which was accessible. The United Verde Extension mine has been inactive since 1938 and only parts of it were accessible.

The increasing economic importance of the Iron King lead-zinc mine near Humboldt and its many similarities to the two large copper mines at Jerome made it desirable to include this mine in this report.

The term "Jerome area" is here used to include all the area in which geologic mapping was done for this report. The term probably should include only a small part of the mapped area, centering at Jerome, but no satisfactory geographic term is available for the entire mapped area.

Several mining districts are included: the Verde (Jerome) centering at Jerome, the Black Hills on the west, the Cherry Creek in the southeast, and the northern part of the Big Bug in the southwest, where the Iron King mine is located.

The Jerome area, as defined above, is served by many roads. Highway 89A crosses the northern part and connects Jerome to the west with Prescott, the county seat of Yavapai County. Northeast of Jerome, this highway connects with Highway 66 at Flagstaff. The southwestern part of the area is crossed by Highway 69, which connects Prescott with Dewey and Humboldt. Many dirt and gravel roads join Highway 89A and provide access to much of the northern half of the area. An excellent graded road encompasses Mingus Mountain, south of the highway, and connects with the highway at the summit and at Jerome. This road, known locally as the "pipeline road" because in part it follows the pipeline supplying Jerome with domestic water, gives access to much of the rugged country in the northern part of the area. An excellent graded road extends east from Dewey through Cherry to the Verde Valley, east of the Jerome area, and dirt roads to cattle ranches provide access to the southern part of the Jerome area.

The main line of the Atchison, Topeka and Santa Fe Railway passes north of the Jerome area, and from Ash Fork, a branch line extends southward through Prescott to Phoenix. From Drake, 32 miles north of Prescott, a branch line of the railroad extends to Clarkdale, parallel to the Verde River throughout much of its course. Dewey, Humboldt, and the Iron King mine are served by another branch line that extends from Prescott to Mayer, 7 miles south of Humboldt.

Jerome, Clarkdale, and Cottonwood are the largest towns in the area. Jerome, perched on the steep eastern slope of the Black Hills, is the home of the United Verde and United Verde Extension mines. Population in the town is steadily decreasing owing to the closing of the United Verde Extension mine in 1938 and of the United Verde mine in 1953. Clarkdale, just north of the mapped area, was the smelter town for the United Verde mine; and with the closing of the smelter in June 1951, the population of the town decreased.

Cherry, a small settlement in the southeastern part of the area, has been the site of some small gold production from quartz veins. Most of the mines have long been inactive.

Humboldt is a small town in the southwest corner of the Jerome area that serves some miners working in the Iron King mine. Dewey is essentially a crossroad settlement where the Cherry road joins Highway 69. There are several cattle ranches scattered around the west, south, and east margin of the area.

The 10,000-foot grid based on Arizona (central) rectangular coordinate system has been placed on plate 1, and many references in this report to localities are made to this coordinate system.

PHYSICAL FEATURES

The Jerome area lies within the Mountain region of Arizona (fig. 1), as defined by Ransome (1903, p. 16), a few miles southwest of the Colorado Plateau.

The Colorado Plateau consists essentially of flat-lying Paleozoic and Mesozoic sedimentary rocks, locally capped by extensive basaltic flows and cones. Most of the margin of the plateau is marked by a continuous line of cliffs ranging from 1,000 to 2,000 feet in height, that to the east of the Jerome area are strikingly exposed across the Verde Valley below the Mogollon Rim, the edge of the plateau.

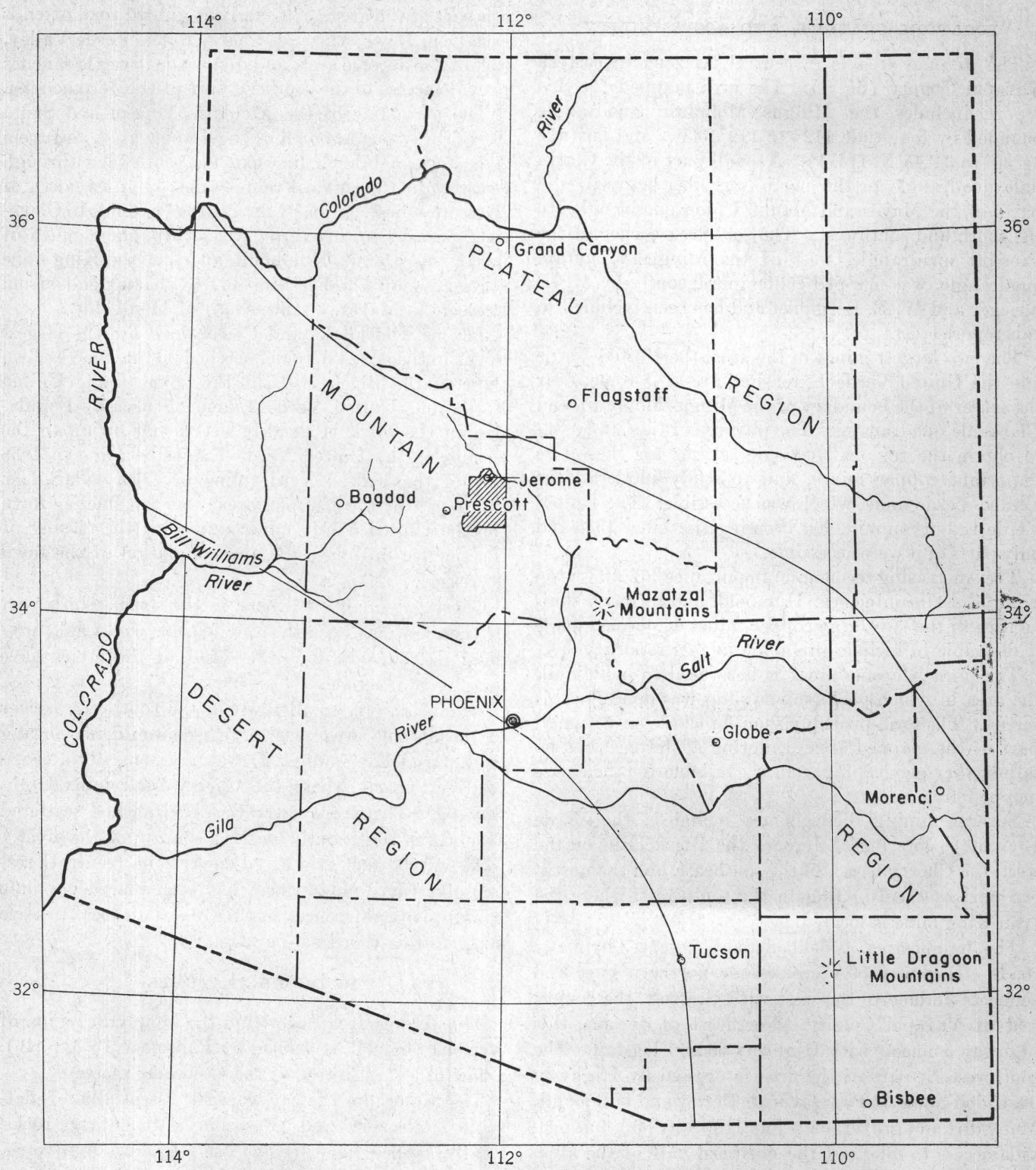

FIGURE 1.—Index map of Arizona showing location of Jerome area. Approximate position of the three principal topographic regions of Arizona outlined. The Plateau region is to the northeast, the Desert region is to the southwest, and the Mountain region lies between them. After Ransome (1903, p. 10).

The mountain region is essentially a broad zone of nearly parallel short ranges, separated in part by valleys, deeply filled with fluviatile and lacustrine deposits. The Black Hills in the Jerome area are an excellent example of one of the north to northwestward-trending ranges in the mountain region, bounded to the east by the Verde Valley and to the west by Lonesome and Chino Valleys. The valleys on both sides of the Black Hills are filled with lacustrine and fluviatile deposits, which are now being eroded to widespread pediments capped by veneers of gravel.

Paleozoic sedimentary rocks in the northern part of the Jerome area continue northward to the plateau and are deeply eroded by the eastward-flowing Verde River. The south margin of the plateau, on physiographic grounds, might be determined by the erosional escarpment of sedimentary rocks of late Paleozoic age (Coconino sandstone and Kaibab limestone) exposed to the north of Verde River. Geologically, the plateau might be extended southward to include the continuous exposures of Paleozoic rocks that crop out in the summit area of the Black Hills, well within the Jerome area.

To the southwest, the Black Hills grade into the Bradshaw Mountains that culminate in a high range southeast of Prescott. To the south, the Black Hills fade into a landscape of low relief.

Three lava-capped mesas comprise the summit region of the Black Hills: Mingus, Hickey, and Woodchute Mountains.

Altitude in the Jerome area ranges from 7,834 feet on Woodchute Mountain to 3,318 feet along the Verde River at Cottonwood. The western slope of Verde Valley rises gradually from the Verde River to the steep eastern front of the Black Hills, and the altitude of the base of this escarpment ranges from 4,200 to 5,000 feet. A long ridge extends southeastward from Mingus Mountain, marking the top of the escarpment along the east margin of the Black Hills; the altitude along this ridge ranges from 6,000 to about 7,000 feet. The south margin of the Jerome area is a surface of low relief, averaging about 4,500 feet in altitude. West of the Black Hills, the lowest altitude is along the Agua Fria River; at Humboldt it is 4,581 feet. The pediment west of the Black Hills slopes gradually from 4,750 feet at the Agua Fria River to about 6,000 feet at the west margin of the Black Hills. The southern part of the west margin of the Black Hills is an area of low to moderate relief; only along the northern half is a pronounced escarpment present.

The Verde River in the northeast corner of the Mingus Mountain quadrangle is the only large perennial stream that flows through the Jerome area. Many intermittent streams drain the east side of the Black Hills to the Verde River. The Agua Fria River drains the southern and southwestern parts of the Jerome area; the Agua Fria is largely intermittent except locally where a small perennial flow is present on bedrock surfaces. During and following heavy summer rains, the Agua Fria may be a raging torrent. North of Coyote Springs Ranch on the west side of the Black Hills, all drainage flows westward to the northward-flowing Granite Creek in the Prescott quadrangle, which is an important tributary to the Verde River.

CLIMATE AND VEGETATION

The temperature of Arizona ranges from the very high summer heat of the southern desert to the extreme winter cold of the high plateaus and mountains to the north. Temperature of the Jerome area is between these two extremes.

Climatic records are available from Clemenceau and Jerome, and from Prescott, to the west. In 1942, a U. S. Weather Bureau office was established at the Love Municipal Airport, 9 miles northeast of Prescott. Records were started in Cottonwood in 1948. The averages from Clemenceau, Jerome, and Prescott give a summary of temperature and precipitation (table 1).

TABLE 1.—*Annual climatic averages, Jerome and Prescott areas*

	Length of record (years)	January	July	Maximum	Minimum
Temperature averages (°F)					
Clemenceau	16	44.3	84.0	110	9
Jerome	39	42.3	78.8	105	7
Prescott	40	35.0	72.5	105	−21

Precipitation averages (inches)

	Length of record (years)	January	February	March	April	May	June
Clemenceau	16	0.95	1.12	0.66	0.72	0.27	0.24
Jerome	30	1.53	1.77	1.27	1.07	.43	.41
Prescott	28	1.80	2.20	1.56	1.20	.44	.34

	Length of record (years)	July	August	September	October	November	December	Annual
Clemenceau	16	1.38	1.80	1.70	0.61	0.77	1.07	11.29
Jerome	30	2.13	2.74	1.59	1.06	.93	1.93	16.86
Prescott	28	2.62	3.39	1.98	.99	1.08	2.38	19.98

The vegetation (Nichol, 1937; Benson and Darrow, 1944) in the Jerome area may be grouped into three types: Pine forest, chaparral, and grassland. Pine forest is limited to altitudes above 6,500 feet, in the Black Hills, except for a small area in the Cherry Creek district. The ponderosa pine (*Pinus ponderosa*)

dominates this forest, but in some breaks, Gambel's oak (*Quercus gambeli*) and Mexican locust (*Robinia neomexicana*) form small but compact thickets.

The heaviest growth of chaparral is south and southeast of Mingus Mountain. Scrub oak (*Quercus turbinella*) and manzanita (*Arctostaphylos pungens*) are the dominant plants in the chaparral, but some mountain mahogany (*Cercocarpus* sp.), cliffrose (*Cowania stanburiana*), and Apache plume (*Fallugia paradoxa*) are locally abundant. In the transition zone between the pine forest and chaparral, pinõn pine (*Pinus edulis*), stringy-bark juniper (*Juniperus utahensis*), and alligator-bark juniper (*Juniperus pachyphloea*) are present. The alligator-bark juniper generally is found at higher altitudes than the stringy-bark juniper.

Grasslands are abundant in Lonesome Valley west of the Black Hills and in Verde Valley to the east. During spring months and after the summer rains, these verdant grasslands are in marked contrast to the duller gray chaparral areas. Small cactus plants, as well as cholla and prickly pear, are widely scattered in the grasslands. Along some of the watercourses, chaparral is abundant. Beargrass (*Nolina microcarpa*) forms scattered clumps in the grassland. Catsclaw or wait-a-minute bush (*Mimosa biuncifera*) is a low-growing shrub that locally forms thickets at lower altitudes.

Century plant or mescal (*Agave parryi*) is common in the lower altitudes of chaparral on the east side of the Black Hills and is sufficiently abundant in the vicinity of Mescal Gulch to suggest this name.

Along the larger streams in the chaparral areas, black walnut (*Juglans rupestris*) and sycamore (*Platanus racemosa*) trees are common. In lower altitudes, the desert willow (*Chilopsis linearis*) and cottonwood (*Populus fremontii*) occur along the watercourses in the chaparral and grassland areas.

FIELD WORK AND ACKNOWLEDGMENTS

Field work was started in October 1945 and essentially completed by October 1951. From January to June 1947, the underground workings of the United Verde mine at Jerome were studied; and from September 1947 to April 1948, Creasey mapped the underground workings of the Iron King mine near Humboldt and made surface geologic maps of parts of the north margin of the Mayer and Mount Union quadrangles. Creasey was absent from the project from September 1948 to June 1949. During much of 1951, Anderson did administrative work not connected with the project. R. E. Lehner joined Creasey in March 1951 to assist in the mapping of the Paleozoic and Tertiary rocks in the Mingus Mountain quadrangle and part of the Clarkdale quadrangle. Lehner later assisted in much of the map compilation and also prepared the section on the Cherry Creek mining district, p. 174.

It is a pleasure to acknowledge the cordial cooperation of Phelps Dodge Corp. in the study of the United Verde mine. J. B. Pullen, general superintendent, United Verde Branch of Phelps Dodge Corp., at the time of our studies of the mine, placed all the facilities of the mine at our disposal and gave us access to all mine data. His successors, C. E. Mills and W. W. Little, provided continued cooperation and aid. Thanks are due to C. R. Kuzell, general manager, Phelps Dodge Corp., for his enthusiastic support of the alteration studies in the United Verde mine and for the chemical analyses of the chloritic alteration products made by the Phelps Dodge laboratories in Douglas, Ariz. L. E. Reber, Jr., resident geologist at the United Verde mine for many years, was generous in advice and counsel. His successor, P. F. Yates, was our constant guide and counselor in our United Verde mine studies, and his remarkable familiarity with the geologic features of the mine was of invaluable aid.

The Mingus Mountain Exploration Co., through its president, Arthur Notman, generously gave us access to the geologic results obtained by its geologist, G. W. H. Norman, in the United Verde Extension mine and much of the country near Jerome. Dr. Norman, who has had much experience in Canadian Precambrian rocks while a member of the Geological Survey of Canada, gave liberally of his time and experience in helping us unravel the complex Precambrian structure. Dr. Norman wrote much of the chapter dealing with the United Verde Extension mine.

H. F. Mills, general manager of the Iron King Branch of the Shattuck Denn Mining Co., kindly permitted the use of all company maps and production data from the Iron King mine. John Kellogg, resident mine geologist, cordially cooperated in many ways.

Eldred D. Wilson of the Arizona Bureau of Mines supplied maps and data on file in his office and gave useful advice in the early stages of the field work.

Our colleagues on the Geological Survey were helpful in many aspects of the project. M. H. Krieger and W. E. Bergquist furnished their geologic map of the southeast corner of the Prescott quadrangle for inclusion in this report. R. S. Cannon, Jr., discussed many problems concerning the ore deposits and structural geology, both in the field and in the office. T. S. Lovering and Earl Ingerson aided in the understanding of the hydrothermal alteration and the metamorphism. George T. Faust and Charles Milton helped in the mineralogical studies. We are grateful to James Gilluly for his pertinent comments in the field about the

structural problems and for his critical review of the manuscript.

PREVIOUS WORK

The following articles, listed chronologically, have been published concerning geologic features in the Jerome area:

1883. Blandy, J. F. The mining region around Prescott, Ariz.: Am. Inst. Min. Metall. Eng. Trans. v. 11, p. 286–291.
Contains a map showing distribution of granite, slate, and limestone. Location of mines of this date are of historical interest.

1905. Jaggar, T. A., and Palache, Charles. Description of Bradshaw Mountains quadrangle, Ariz.: U. S. Geol. Survey Geol. Atlas, folio 126, 11 p.
First adequate description of the Precambrian rocks in the region; only a small part of this quadrangle is included in the study of the Jerome area, but this folio was important pioneer work in the region.

1906. Reid, J. A. Sketch of the geology and ore deposits of the Cherry Creek district, Ariz.: Econ. Geology, v. 1, p. 417–436.
Description of the gold veins in Cherry Creek mining district.

1916. Provot, F. A. Jerome mining district geology: Eng. and Min. Jour., v. 102, p. 1028–1031.
General discussion of the geology of the mines at Jerome.

1916. Ransome, F. L. Some Paleozoic sections in Arizona and their correlation: U. S. Geol. Survey Prof. Paper 98–K, p. 133–166.
Brief description of Paleozoic rocks in the Jerome area and faunal lists proving Devonian age of Martin limestone and Mississippian age of Redwall limestone. Correlates basal Paleozoic sandstone with Tapeats sandstone of Cambrian age in Grand Canyon section.

1918. Rickard, T. A. The story of the U. V. X. bonanza: Min. Sci. Press, v. 116, p. 9–17, 47–52.
Fascinating account of the history of mining at Jerome; Rickard states that the United Verde Extension ore body was not originally connected to the United Verde ore body.

1918. Finlay, J. R. The Jerome district of Arizona: Eng. and Min. Jour., v. 106, p. 557–562, 605–610.
First detailed description of the geology of the United Verde mine, Jerome.

1918. Tovote, W. L. Notes on certain ore deposits of the southwest: Am. Inst. Min. Metall. Eng., Bull. 142, p. 1599–1612.
Main point of this report is the objection to the theory that ore deposits at Jerome are of Precambrian age, contrary to the conclusion in our present report.

1920. Rice, Marion. Petrographic notes on the ore deposits of Jerome, Ariz.: Am. Inst. Min. Metall. Eng. Trans., v. 61, p. 60–65.
Good petrographic description of rocks and ores.

1922. Reber, L. E., Jr. Geology and ore deposits of the Jerome district: Am. Inst. Min. Metall. Eng. Trans., v. 66, p. 3–26.
Excellent description of local geology and ore deposits at Jerome, based on years of observation in the district. Precambrian rocks subdivided and adequately described. Replacement nature of ore deposits demonstrated and first complete description of the ore bodies and their control given.

1923. Jenkins, O. P. Verde River lake beds near Clarkdale, Ariz.: Am. Jour. Sci., 5th ser., v. 5, no. 25, p. 65–81.
First detailed description of the late Tertiary or early Quaternary lake beds found in the Verde Valley.

1925. Fearing, J. L., Jr., and Benedict, P. C. Geology of the Verde Central mine: Eng. and Min. Jour., v. 119, p. 609–611.
Brief account of the history and geology of a copper mine southwest of Jerome.

1926. Fearing, J. L., Jr. Some notes on the geology of the Jerome district, Ariz: Econ. Geology, v. 21, p. 757–773.
Discusses geology and ore deposits in Jerome. Believes that United Verde Extension ore body was not originally connected to United Verde ore body.

1926. Lindgren, Waldemar. Ore deposits of the Jerome and Bradshaw Mountains quadrangles, Ariz.: U. S. Geol. Survey Bull. 782, 192 p.
Main value of this bulletin is the wealth of information on the mines and prospects in the area. Reconnaissance map of Jerome quadrangle included, but regional geology not treated in a comprehensive fashion.

1927. Anderson, C. A. Voltaite from Jerome, Ariz.: Am. Mineralogist, v. 12, p. 287–290.
Brief description of one of the hydrous sulfates formed by the mine fires.

1928. Lausen, Carl. Hydrous sulphates formed under fumarolic conditions at the United Verde mine: Am. Mineralogist, v. 13, p. 203–225.
Interesting description of 9 hydrous sulfate minerals, 5 of them new, formed at or near surface resulting from mine fires.

1930. Hansen, M. G. Geology and ore deposits of the United Verde mine: Min. Cong. Jour., v. 16, p. 306–312.
First paper to suggest Precambrian movement on Verde fault.

1930. Lausen, Carl. The pre-Cambrian greenstone complex of the Jerome quadrangle: Jour. Geology, v. 38, p. 174–183.
Separated a greenstone complex from Yavapai schist (metasedimentary rocks) and suggested an unconformable relationship between the two; Wilson (1939) later showed that a fault relationship exists between these units.

1930. Newhouse, W. H., and Flaherty, G. E. The texture and origin of some banded or schistose sulphide ores: Econ. Geology, v. 25, p. 600–620.
Brief description of ore from the United Verde mine at Jerome.

1930. Ralston, O. C. Possibilities of zinc production in Arizona: Min. Jour., v. 14, p. 11.
Description of zinc occurrence in the United Verde mine, Jerome.

1931. Ingalls, W. R. World survey of the zinc industry: Min. Met. Soc. America, N. Y.
Emphasized the zinc content of pyritic ore at United Verde mine, Jerome.

1932. Ransome, F. L. General geology and summary of ore deposite, *in* Ore deposits of the southwest: 16th Internat. Geol. Cong. Guidebook 14, Excursion C–1, p. 1–23.
Resumé of Arizona geology, but gives considerable attention to ore deposits of Jerome. Emphasized Precambrian movement on Verde fault.

1935. Tenney, J. B. The copper deposits of Arizona, *in* Copper resources of the world: 16th Internat. Geol. Cong., v. 1, p. 167–235.
Historical data about mines at Jerome.

1936. Stoyanow, A. A. Correlation of Arizona Paleozoic formations: Geol. Soc. America Bull., v. 47, p. 459–540.
Described the Devonian Jerome formation, which is termed Martin limestone in this report.

1938. Reber, L. E., Jr. Jerome district, *in* Some Arizona ore deposits: Ariz. Bur. Mines Bull. 145, p. 41–65.
Valuable sequel to Reber's 1922 paper; later developments in Jerome district adequately described and illustrated. Much valuable information on the United Verde Extension mine which closed in 1938.

1938. Schwartz, G. M. Oxidized copper ores of the United Verde Extension mine: Econ. Geology, v. 33, p. 21–33.
Petrographic study of chalcocite and oxide ore; paragenesis of minerals.

1939. Wilson, E. D. Pre-Cambrian Mazatzal revolution in central Arizona: Geol. Soc. America Bull., v. 50, p. 1113–1164.
First attempt to divide rocks of the Yavapai schist of Jerome area into separate stratigraphic units and a forward step in the understanding of the Precambrian geology.

1941. Mills, H. F. Ore occurrence of the Iron King mine: Eng. and Min. Jour., v. 142, no. 10, p. 56–57.
Brief description of the Iron King mine.

1943. Gutschick, R. C. The Redwall limestone (Mississippian) of Yavapai County, Ariz.: Plateau, v. 16, no. 1, p. 1–11.
Presents measured sections of Redwall limestone as exposed in Jerome area.

1945. Huddle, J. W., and Dobrovolny, E. Late Paleozoic stratigraphy and oil and gas possibilities of central and northeastern Arizona: U. S. Geol. Survey Oil and Gas Inv., Prelim. Chart 10.
Description and measured sections of Paleozoic rocks adjacent to the Jerome area.

1946. Yates, P. F. Bottoming of the United Verde sulphide pipe: Mimeographed paper distributed at meeting of Ariz. Section, Am. Inst. Min. Metall. Eng., Tucson, Ariz., Oct. 1946, 20 p.
Contains level plans and section showing character of United Verde ore body and description of exploration activities on lower levels of the mine.

1947. Mills, H. F. Occurrence of the lead-zinc ore at Iron King mine, Prescott, Ariz.: Am. Inst. Min. Metall. Eng. Min. Tech., v. 11, no. 4, Tech. Paper 2190, 4 p.
Excellent brief account of the geology of this mine emphasizing character of ore shoots.

1949. Mahard, R. H. Late Cenozoic chronology of the upper Verde Valley, Ariz.: Denison Univ. Bull. Jour. Sci. Lab., v. 41, p. 97–127.
Suggests a complex history for the Verde Valley before deposition of the Verde formation.

1950. Creasey, S. C. Iron King mine, Yavapai County, Ariz., *in* Arizona zinc and lead deposits: Ariz. Bur. Mines Bull. 156, p. 112–122.
Preliminary statement on the occurrence of ore in the Iron King mine.

1950. Little, W. W. Radial blast holes for drilling an irregular ore body: Min. Engineering, p. 463–465.
Describes copper-zinc ore to north of main ore body, United Verde mine.

1950. Price, W. E., Jr. Cenozoic gravels on the rim of Sycamore Canyon, Ariz.: Geol. Soc. America Bull., v. 61, no. 5, p. 501–508.
Presents evidence to show that before basaltic eruptions on plateau, drainage from Precambrian rocks in Jerome area was to the north.

1951. Anderson, C. A. Older Precambrian structure in Arizona: Geol. Soc. America Bull., v. 62, p. 1331–1346.
Preliminary statement on the Precambrian stratigraphy and structure in the Jerome area.

1951. McNair, A. H. Paleozoic stratigraphy of part of northwestern Arizona: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 503–541.
Presents evidence for believing Cambrian sedimentation did not extend to Jerome area.

1952. Creasey, S. C. Geology of the Iron King mine, Yavapai County, Ariz: Econ. Geology, v. 47, p. 24–56.
Describes the local geology of this mine, the alteration, metallization, and structural control of the ore.

1952. Huddle, J. W., and Dobrovolny, E. Devonian and Mississippian rocks of central Arizona: U. S. Geol. Survey Prof. Paper 233–D, p. 67–112.
Description of these rocks southeast of the Jerome area, with discussion of the regional distribution and history.

1955. Huff, L. C. A Paleozoic geochemical anomaly near Jerome, Arizona: U. S. Geol. Survey Bull. 1000–C, p. 105–118.
Sampling of the Tapeats(?) sandstone revealed anomalous copper concentrations of more than 100 ppm (parts per million) near Jerome and in an area extending about 2 miles southeast from Jerome. Elsewhere this same basal sandstone contains about 20 ppm copper. Anomalous zinc concentrations of more than 100 ppm are distributed in roughly the same area as the copper. Huff concluded that the abnormal copper and zinc content of the Tapeats(?) represents products of erosion from the ore deposits and that geochemical prospecting techniques can be used to identify and trace heavy-metal anomalies in basal sandstones such as the Tapeats(?).

GENERAL GEOLOGY

OLDER PRECAMBRIAN ROCKS

YAVAPAI SERIES

The Yavapai series includes all the older Precambrian volcanic and sedimentary rocks in the Prescott-

Jerome area. Jaggar and Palache (1905) first described this series of metamorphic rocks as the Yavapai schist. According to them, the rocks include slate, phyllite, mica schist, and chlorite schist, as well as local gneiss, granulite, hornfels, and hornblende schist. Jaggar and Palache concluded that the Yavapai schist is of sedimentary origin and equivalent to the Vishnu schist in the Grand Canyon. Later, Lindgren (1926) recognized rhyolitic and basaltic flows in the Yavapai schist, and concluded that the Yavapai schist comprises metamorphosed sedimentary beds and a large amount of interbedded flows and tuffs. Lindgren also extended the Yavapai schist into the Jerome area where metamorphosed volcanic rocks are common.

Lausen (1930) limited the Yavapai schist to the schistose metasediment exposed along the west margin of the Black Hills. For the more massive volcanic rocks to the east, he used the term "greenstone complex," admitting, however, that some interbedded tuff and sedimentary rocks are present. Lausen interpreted the contact of the Yavapai schist and the greenstone complex as an angular unconformity, the greenstone complex being younger. Wilson (1939, p. 1159) correctly concluded that a fault relationship exists between Lausen's Yavapai schist and greenstone complex. Wilson also substituted the term "Yaeger greenstone" for Lausen's greenstone complex, and the term "Alder series" for the schistose metasedimentary rocks to the west, placing both units under the term "Yavapai group."

The present study has shown that the separation of these older Precambrian rocks into two major subdivisions has merit, but the limitation of the term "Yavapai schist" to the western schistose metasediments is untenable because much of the western section contains volcanic flows and tuffs. Furthermore, tuffaceous sedimentary rocks are important constituents of the eastern greenstone complex or Yaeger greenstone, and locally the eastern rocks are schistose. In this report the eastern rocks have been divided into seven formations, and one of these has been subdivided into four lithologic facies; the eastern rocks are assigned a group rank and termed the Ash Creek group. The western rocks have been divided into six formations, and Wilson's Alder series has been modified to Alder group to include them. Because of the fault between the Alder and Ash Creek groups, no statement can be made as to their respective ages. Wilson (1939, p. 1120) suggested that Yavapai schist be modified to Yavapai group, but in keeping with standard stratigraphic nomenclature, Yavapai series is used in this report to replace Yavapai schist.

Although the rocks in the Yavapai series have been metamorphosed, our primary interest in these rocks was not metamorphic processes, but their stratigraphy and structure as a clue to the control of the ore deposits contained in the Yavapai series and associated intrusive igneous rocks. Considerable time was spent searching for relict textures and structures in order to determine the character of the rocks before metamorphism. Billings (1950) has recently emphasized the importance of this approach in the study of metamorphic terranes, which has been standard practice of the Geological Survey of Canada in unraveling the older Precambrian geology of Canada. Following the Canadian procedure, the prefix "meta" has not been used in this report for the rocks of the Yavapai series. Chlorite schist in which relict amygdules and pillow lava can be recognized is classified as andesite or basalt; chlorite schist, in which relict bedding structure is widespread, is classified as andesitic or basaltic sedimentary tuff or tuffaceous sedimentary rock. Many of the rocks of the Yavapai series are so weakly foliated that no doubt exists as to their original character. In the highly foliated rocks, our interpretation may be open to question for particular outcrops, but by tracing formations along the strike, sufficient relict features have been found to confirm our interpretations.

ASH CREEK GROUP

The Ash Creek group is named from Ash Creek, which cuts through a representative section south of Mingus Mountain. The rocks consist of basaltic, andesitic, rhyolitic, and dacitic flows and pyroclastic rocks, and a thick sequence of tuffaceous sedimentary rocks containing interbeds of jasper-magnetite and chert. The Ash Creek group is exposed to the south and to the east of Mingus Mountain, and to the west it is separated from the Alder group by the Shylock fault. Rocks of the Ash Creek group generally are nonfoliated, and relict structures in the lava and tuffaceous sedimentary rocks are well preserved; determinations are easily made of the directions that the beds or lava flows face. Some foliation is present along the east margin of Mingus Mountain; it increases in intensity toward Jerome where locally the tuffaceous sedimentary rocks are highly foliated.

The Ash Creek group has been divided into seven formations. The Gaddes basalt is the oldest, and is overlain by the Buzzard rhyolite. The Shea basalt, dacite of Burnt Canyon, and Brindle Pup andesite are lenticular units about the same age, and separate the Buzzard rhyolite from the younger Deception rhyolite. The Grapevine Gulch formation is younger than the Deception rhyolite.

The section of the Ash Creek group is perhaps 20,000 feet thick, not including the base or top. The sum of

the maximum thickness of each formation is 23,500 feet, but because some of the volcanic formations are lenticular, a lower figure is a more accurate estimate for the group as a whole.

GADDES BASALT

Distribution

Gaddes Canyon, south of Mingus Mountain (1,330,000 N.; 443,000 E.) contains good exposures of the basalt, here defined as the Gaddes basalt. Gaddes Canyon is a tributary to Black Canyon, where excellent outcrops of the basalt are exposed. To the north of Black Canyon, a smaller outcrop is present beneath the Paleozoic rocks. A large outcrop of Gaddes basalt lies east of Mingus Mountain in the drainage basin of Oak Wash (1,342,000 N.; 450,000 E.), where exposures are not generally as good as in Black and Gaddes Canyons.

Thickness and stratigraphic relationship

In Black Canyon where the Gaddes basalt is exposed in the crest of a southeast plunging anticline, the base is not exposed. Precise data on the attitude are so meager that the thickness cannot be accurately determined, but we estimate that 2,000 to 2,500 feet of flows are exposed in Black Canyon. No estimate of thickness can be made in the drainage basin of Oak Wash.

One mile west of Ward Pocket along the contact of Gaddes basalt and the overlying Buzzard rhyolite (1,327,300 N.; 445,000 E.), the basalt is overlain by rhyolitic breccia that dips 40° S. and grades upward into bedded sandy and silty beds that are a foot thick. Excellent graded bedding and channeling prove that the top of the beds face south. In Black Canyon, north of Wards Pocket, rhyolitic flows dip 45° SE. overlying the Gaddes basalt. Three-quarters of a mile southeast of Allen Spring, sandy interbeds in the Buzzard rhyolite dip 20° S. (1,336,900 N.; 449,700 E.), and graded bedding indicates that the beds face south. These exposures prove conclusively that the Gaddes basalt is older than the Buzzard rhyolite, but three small outcrops of rhyolite are intercalated between basaltic flows in Gaddes Canyon (pl. 1) indicating that rhyolite and basalt were erupted, in part, contemporaneously.

Lithology and internal structure

The Gaddes basalt forms black to dark-green outcrops. East of Mingus Mountain where exposures are poor, the hill slopes are covered by dark-brown soil. On unweathered surfaces, the Gaddes basalt is dark green. Pillow structure is common (pl. 3*A*) and is best recognized in walls and the floors of Gaddes and Black Canyons, and in some of the stream floors of the Oak Wash drainage basin. The pillow structures range from 1 to 5 feet in length and average 1½ to 2. A finer grained shell is present on many of the pillows (pl. 3*B*). Chalcedonic filling between pillows is common, particularly in Black Canyon. These silicic deposits are usually nearly triangular in outline but some are irregular to square, coinciding with the openings between pillows. Their size ranges from 2 to 10 inches. Amygdaloidal facies are common in the lava flows (pl. 3*C*); quartz is the chief amygdaloidal filling, but locally calcite and epidote occur.

Weak foliation, striking east and dipping steeply north, is present locally in the southern exposures of the Gaddes basalt. The foliation planes wrap around the pillows. In this southern area, the long dimension of the pillows also plunges north steeply, parallel to the weak foliation. East of Mingus Mountain, the Gaddes basalt is more strongly foliated locally, and the rock is silicified and contains chlorite along fractures parallel to the foliation.

Pyroclastic rocks appear in the Gaddes basalt east of Mingus Mountain, but owing to the poor exposures and small size of the bodies, they were not distinguished in mapping. In part, the pyroclastic facies consists of highly vesicular or amygdaloidal irregular-shaped basaltic fragments resembling lapilli and small bombs. They are closely packed, resembling agglutinates (Tyrrell, 1931, p. 66) formed at the time of eruption by welding of semiplastic ejecta. Other fragmental facies consist of closely packed angular fragments, from 2 to 6 inches across.

Microscopic studies reveal that the major texture of the pillow basalt is pilotaxitic; that is, albite or albite-oligoclase microlites averaging 0.4 millimeter in length are separated by bluish-green or pale-green amphibole needles or chlorite or both. The feldspar crystals are subparallel, forming a trachytoid texture. Minute magnetite crystals are interstitial to the feldspar and mafic minerals. Epidote granules are common, and locally the pillow lavas are strongly epidotized. East of Mingus Mountain, chlorite is common and calcite and sericite occur in streaks and patches. In the southern exposures of Gaddes basalt, chlorite is limited to the northern part of Gaddes Canyon and to quartz veinlets. Quartz is common as amygdaloidal fillings, in places rimmed by chlorite. Quartz also occurs as grains 0.1 millimeter across, interstitial to the plagioclase, but it cannot be determined if the quartz is a primary constituent or was formed during metamorphism. Fragments of lapilli from the pyroclastic facies have the same texture and minerals except that the plagioclase microlites are only 0.2 millimeter long.

The distinction between unmetamorphosed andesite and basalt may be difficult, and where the rocks are

metamorphosed, the changed mineralogical character adds complications. The chemical analysis of the Gaddes basalt (table 2) indicates that the pillow lava is intermediate in composition between average basalt and andesite. In the lava of the Gaddes the SiO_2, Al_2O_3, and MgO contents are intermediate; the CaO and Na_2O contents are near the average andesite but the K_2O content is low even for the average basalt, suggesting a spilitic character. The iron content is nearer to that of the average basalt.

The designation of the pillow lava of the Gaddes as basaltic is arbitrary, and is based mainly on the high content of mafic minerals, and the chemical analyses. Satterly (1941, p. 133) compiled the available analysis of the Canadian Precambrian pillow lavas and most of these are basaltic in composition, but Wilson (1941, p. 18) observed pillows in Precambrian dacitic lava in the Noranda district in Quebec.

The presence of albite in these pillow lavas is not sufficient reason to refer to them as spilites (Dewey and Flett, 1911), for the chemical analysis does not indicate a high soda content. The abundant epidote granules indicate that the original pillow lavas contained calcic plagioclase, and that the albite is a metamorphic product—a conclusion reached by Gunning and Ambrose (1940, p. 6) for Precambrian albitic lava in Quebec.

Table 2.—Chemical analyses of Gaddes basalt and average andesite and basalt, in percent

	1	2	3
SiO_2	54.8	59.59	49.06
Al_2O_3	16.2	17.31	15.70
Fe_2O_3	2.4	3.33	5.38
FeO	9.8	3.13	6.37
MnO	.26	.18	.31
MgO	3.7	2.75	6.17
CaO	4.4	5.80	8.95
K_2O	.36	2.04	1.52
Na_2O	3.2	3.58	3.11
TiO_2	1.0	.77	1.36
P_2O_5	.28	.26	.45
Ignition loss	2.0	------	------

1. Sample of pillow lava (Gaddes basalt) from Black Canyon (1,329,000 N.; 443,500 E.). Rapid analysis by S. M. Berthold and E. A. Nygaard.
2. Average andesite.
3. Average basalt, both from Daly (1933, p. 16–17).

BUZZARD RHYOLITE

Distribution

Buzzard Canyon, southeast of Mingus Mountain, contains good exposures of the Buzzard rhyolite, suggesting this term for the formation defined here. Buzzard Canyon is a tributary to Black Canyon where other excellent outcrops of the rhyolite are exposed. An incomplete section of Buzzard rhyolite is exposed east of Mingus Mountain.

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Thickness and stratigraphic relationship

A complete section of Buzzard rhyolite is exposed in the lower part of Black Canyon where it clearly overlies the Gaddes basalt and is overlain by the Shea basalt. The thickness of the rhyolite here is estimated at 3,500 feet. To the west where the Buzzard rhyolite is separated from the younger Deception rhyolite by the dacite of Burnt Canyon and Brindle Pup andesite, the Buzzard rhyolite is about 2,000 feet thick.

East of Mingus Mountain and north of Oak Wash (1,344,000 N.; 448,000 E.), the Buzzard rhyolite clearly underlies the Shea basalt, because tuffaceous interbeds near the top of the Buzzard rhyolite dip and face northwest, comformable to the basal contact of the Shea basalt. About 2,500–3,000 feet of rhyolitic rocks are exposed in this section, assuming no duplication by folding or faulting. The section is separated by intrusive quartz porphyry. The Buzzard rhyolite is in fault contact (Oak fault) with the Gaddes basalt to the south and is cut off to the east by the Verde fault.

Lausen (1930, p. 181) noted the rhyolitic rocks in Black Canyon and suggested that the rhyolite was erupted after folding of the basaltic rocks; Wilson (1939, p. 1159) stated that the rhyolite is in fault contact with the adjacent rocks. Lausen and Wilson grouped the younger Grapevine Gulch formation with the older basalt, and neither recognized the Shea basalt, dacite of Burnt Canyon, and Brindle Pup andesite that separate the Buzzard rhyolite from the Deception rhyolite. The variations in strike noted by Lausen are an expression of folds rather than an unconformity.

Lithology and internal structure

Much of the Buzzard rhyolite forms bold outcrops except in the forested headwaters of Black Canyon. The outcrops in the lower reaches of Black Canyon are tinted in shades of red whereas those in the headwaters are buff to cream. Contorted flow banding and amygdaloidal and vesicular structures are common in the flows. Fragmental structures are common (pl. 3*D*); some represent flow breccia containing fragments ranging in size from an inch to a foot, but averaging about 3 to 4 inches. Some breccia beds do not have a recognizable lava matrix, and their origin is in doubt. The fragments in many of the breccia beds show a preferred orientation, parallel in their long dimensions. Thin sandy interbeds in breccia are exposed near the base of the Buzzard rhyolite west of Ward Pocket (1,327,300 N.; 445,000 E.) and three quarters of a mile southeast of Allen Spring (1,336,900 N.; 449,700 E.). Similar interbeds of sedimentary rock are exposed near the top of the Buzzard rhyolite north of Oak Wash.

Megascopically, the rhyolitic rocks are commonly porphyritic, containing phenocrysts of quartz ranging from

0.5 to 1.0 millimeter in diameter, and feldspar ranging from 1 to 2 millimeters in length. In some specimens, the feldspar is in glomeroporphyritic aggregates. The ratio of quartz to feldspar phenocrysts is variable, and in some specimens, quartz phenocrysts are absent. The groundmass is aphanitic, ranging in color from pale to dark to purplish gray. Thin sections reveal that all the feldspar phenocrysts are albitic plagioclase and that the groundmass is a microgranular aggregate of quartz and alkalic feldspar. In some specimens the groundmass feldspar is in radiating aggregates separated by interstitial quartz, and in others, feldspar forms small spherulites. In other samples, the groundmass quartz and feldspar are anhedral. The size of the crystals in the groundmass ranges from 0.02 to 0.06 millimeter. Apatite and magnetite are common accessory minerals.

In some of the rhyolitic specimens, sericite is interstitial to the quartz and feldspar, or is in parallel veinlets that displace earlier quartz veinlets, suggesting that the rocks have been slightly sheared. Many rhyolitic rocks contain minute flakes of greenish biotite that commonly are preferentially oriented or are distributed throughout the groundmass in intersecting veinlets. In some of the rhyolite having a dark-gray groundmass, chlorite is interstitial to the quartz and feldspar and is associated with epidote or clinozoisite granules. These varieties may have been originally dacitic in composition, containing enough mafic minerals to form the chlorite and sufficient lime in the plagioclase to form the epidote and clinozoisite.

The sample of Buzzard rhyolite selected for analysis (table 3) is representative of the flow-banded rhyolite that contains no phenocrysts; the rock is essentially sutured microcrystalline quartz and alkalic feldspar with some spherulites of albite. Greenish biotite and some chlorite are in clots and streaks, and some secondary epidote granules are scattered through the rock. The high SiO_2 and low CaO content show that the rock is rhyolitic, and the excess of Na_2O over K_2O indicates that the rock is essentially a soda rhyolite. The ubiquitous albite in all the rhyolite of the Buzzard indicates that the rocks are all sodic.

TABLE 3.—*Chemical analysis of Buzzard rhyolite*

Flow-banded rhyolite, south slope of Black Canyon (1,326,400 N; 445,300 E.). Rapid analysis by S. M. Berthold and E. A. Nygaard]

	Percent
SiO_2	75. 9
Al_2O_3	11. 9
Fe_2O_3	1. 3
FeO	2. 7
MnO	. 10
MgO	. 43
CaO	. 96
K_2O	1. 2
Na_2O	4. 4
TiO_2	. 32
P_2O_5	. 02
Ignition loss	. 86

SHEA BASALT

Distribution

The Shea basalt crops out southeast and east of Mingus Mountain. The southeastern exposures appear north and south of Black Canyon and in the canyon west of the Verde fault. The eastern exposures form a continuous belt northward from Oak Wash (1,343,000 N.) to Mescal Gulch (1,358,500 N.). The thick flows of Shea basalt have a holocrystalline texture and Reber (1938, p. 58) termed this fine-grained dioritic rock the Shea diabase because of the excellent exposures near the Shea mine. As Gunning and Ambrose (1940, p. 26) state, one of the most difficult problems in mapping is to distinguish between intrusive diorite or diabase and holocrystalline facies of basaltic and andesitic flows. Although some intrusive diabase probably is present in the areas mapped as Shea basalt, much of the holocrystalline diabasic rock can be related to the interior of thick flows, and it appears desirable to modify the term of "Shea diabase" to "Shea basalt" to include all the mafic volcanic rocks of this unit.

Thickness and stratigraphic relationship

The thickness of the Shea basalt and interbedded tuffaceous sedimentary rocks is about 2,000 feet on the north side of Black Canyon, assuming that the rhyolitic outcrops along the Verde fault (pl. 1, 1,333,000 N.; 457,000 E.) represent the Deception rhyolite, marking the top of the Shea basalt. The base of the Shea basalt resting on the Buzzard rhyolite is well exposed in Black Canyon. No other place provides satisfactory data about thickness, because south of Black Canyon, the Shea basalt is intruded by quartz diorite and is partly covered by sedimentary rocks of Paleozoic age.

East of Mingus Mountain, the base of the Shea basalt is exposed north of Oak Wash (1,344,000 N.; 447,000 E.). Interbedded tuffaceous sediments near the top of the Buzzard rhyolite dip northwest, conformable with the rhyolite-basalt contact. Graded bedding in the tuffaceous sedimentary rocks and vesicular flow tops indicate that the tops of the tuffaceous and flow rocks face northwest, and the Buzzard rhyolite is older than the Shea basalt. On the south side of Mingus Mountain, the site of probable contact between the Shea basalt and Deception rhyolite is covered by Paleozoic rocks (1,322,000 N.; 448,000 E.), but the eastward extension of the Deception rhyolite from where it overlies the Buzzard rhyolite indicates that the Deception rhyolite is younger than the Shea basalt which lenses out to the west. In the Mescal Gulch area, the structure involving the Shea basalt-Deception rhyolite contact is complex, and poor exposures in part mask the contact. The major structure in the Mescal Gulch area appears to be

a northwestward-plunging anticline, so that younger formations appear progressively to the northwest. By this structural interpretation, the Shea basalt is the oldest and is overlain by the Deception rhyolite and the Grapevine Gulch formation in that order. Supporting evidence occurs in Hull Canyon (1,360,000 N.; 436,000 E.) where the Grapevine Gulch formation clearly overlies the Deception rhyolite. Rhyolite crops out in small masses in the Shea basalt, and some basalt or andesite appears in the Deception rhyolite indicating that rhyolitic and basaltic rocks were erupted simultaneously.

Lithology and internal structure

The Shea basalt is a dark-green to black rock, forming good exposures only in Black Canyon and south of the Copper Chief mine. Throughout much of the exposed area east of Mingus Mountain, the surface is covered with dark-brown soil, and bedrock is exposed only in stream gulches or in limited outcrops.

The Shea basalt is composed mostly of lava flows and intercalated tuffaceous beds, and in Black Canyon, many coarse fragmental deposits contain vesicular fragments as much as 6 inches in diameter. These fragments are associated with pyroclastic deposits of lapilli and small bombs. Flows commonly contain quartz and chlorite amygdules, and in a few places, pillow lavas. The weathered surfaces of some of the centers of thick flows reveal an excellent diabasic texture. Some small masses of intrusive diabase probably occur in the Shea basalt but they are not distinguished on the geologic map (pl. 1).

The interbedded tuffaceous sedimentary rocks are commonly thin, from 2 to 4 feet thick, and many contain chert beds that range from 2 to 4 inches in thickness. In the northern exposures of Shea basalt, the tuffaceous sedimentary rocks range from 20 to 40 feet in thickness and contain fragments of lava as much as an inch in diameter. These sedimentary rocks contain an appreciable amount of chlorite and are dark green to greenish gray.

Thin sections reveal some variation in texture and mineral composition. Many of the amygdaloidal lava flows are similar to the Gaddes basalt in having a pilotaxitic texture; they are composed of albite or sodic oligoclase separated by hornblende or chlorite or both. Chlorite is more common in the northern exposures than in the southern. Quartz, sericite, epidote, and calcite are the common accessory minerals.

The massive lava from the thick flows has a diabasic texture, and commonly tabular albite crystals 1 millimeter long are separated by bluish-green hornblende and interstitial quartz. Near the Copper Chief mine where the diabasic facies is common, chlorite is the only mafic mineral, and secondary calcite is present in veinlets or disseminated granules. Slender apatite needles as well as magnetite grains are abundant.

Adjacent to the quartz diorite south of Black Canyon, the Shea basalt is massive and of uniform texture; the grain size averages about 1 millimeter across. Thin sections reveal that the rocks consist largely of sodic oligoclase, greenish-brown hornblende, interstitial quartz, and scattered magnetite grains. The hornblende in places is poikilitic, surrounding small crystals of the other constituents. In other places, the basalt has a granular texture that resembles hornfels.

The chemical analyses (table 4) indicate the basaltic composition of the Shea. The pillow lava near the Shea mine, analysis 1, has a pilotaxitic texture of albitic plagioclase separated by bright-green hornblende with some interstitial chlorite. Epidote granules are common, and probably contain the lime expelled from the plagioclase. The holocrystalline facies, analysis 2, is from the interior of one of the thick flows of basalt; the texture is essentially diabasic (ophitic) with albitic plagioclase crystals separated by chlorite and interstitial quartz. Secondary epidote and calcite are common. Analyses 3 and 4 are from outcrops where the holocrystalline diabasic facies is common, and presumably both samples are from the interior of thick flows. Appreciable alteration is indicated by the high CO_2 and H_2O content; it is possible that the low-silica content of analysis 4 is due to a loss of SiO_2 through hydrothermal alteration.

TABLE 4.—*Chemical analyses of Shea basalt, in percent*

	1	2	3	4
SiO_2	51.0	54.4	49.88	44.02
Al_2O_3	13.6	13.6	14.48	18.05
Fe_2O_3	3.2	5.9	----	----
FeO	12.9	8.6	[1] 6.75	[1] 10.06
MnO	.33	.14	----	----
MgO	3.5	4.0	7.24	7.96
CaO	6.8	3.6	6.94	3.99
K_2O	.14	.28	1.14	2.45
Na_2O	2.6	3.7	4.59	4.44
TiO_2	2.5	2.2	.20	.10
CO_2	----	----	5.37	4.29
P_2O_5	.70	.40	----	----
Cu	----	----	.01	.02
H_2O	----	----	3.79	4.84
Ignition loss	2.0	3.1	----	----
			100.39	100.22

[1] All iron determined as FeO.

1. Pillow lava, near Shea mine (1,346,700 N.; 447,350 E.).
2. Holocrystalline lava, southeast of Shea mine (1,345,700 N.; 448,100 E.). Analyses 1 and 2 are rapid analyses by S. M. Berthold and E. A. Nygaard.
3. Shea basalt along "pipeline road" (1,346,500 N.; 445,900 E.).
4. Shea basalt, west of Copper Chief mine (1,348,000 N.; 446,600 E.). Analyses 3 and 4 furnished by courtesy of Phelps Dodge Corp.

BRINDLE PUP ANDESITE

Distribution

The Brindle Pup andesite occurs as a thick lens to the south of Mingus Mountain, separating the older Buzzard rhyolite from the younger Deception rhyolite. It is well exposed in Brindle Pup Gulch (1,323,000 N.; 442,000 E.) and this name has been used for the formation defined here.

Thickness and stratigraphic relationship

The Brindle Pup andesite ranges in thickness from about 2,500 feet in the widest outcrop to a wedge line. The andesite is cut off by the younger intrusive quartz porphyry along the southeastern contact and covered by Paleozoic sedimentary rocks along part of the northern contact. Intercalated Buzzard and Deception rhyolitic flows are distinguished on the geologic map (pl. 1), but intercalated basaltic flows similar to the Shea basalt are not. These intercalated flows indicate that rhyolitic, basaltic, and andesitic lava flows were erupted simultaneously, and indicate that the Brindle Pup andesite is of the same general age as the Shea basalt; both separate the older Buzzard rhyolite from the younger Deception rhyolite.

Lithology and internal structure

The Brindle Pup andesite is a dark-gray rock which weathers to light-brown surfaces on which cream-colored plagioclase phenocrysts, 4 to 6 millimeters long, are conspicuous. Small clots of indeterminate mafic minerals are present in some specimens. Vesicles and quartz amygdules are present, increasing in abundance near the flow tops where the plagioclase phenocrysts are smaller, averaging about 1 millimeter. The groundmass is aphanitic to very finely crystalline, and in the highly vesicular and amygdaloidal facies, the groundmass is darker than in the main porphyritic facies.

Thin sections of the andesite reveal that the phenocrysts range in composition from albite to sodic oligoclase; in some specimens, sericite and epidote are conspicuous inclusions in the feldspar; and in others, only scattered granules of epidote appear. Original mafic phenocrysts are represented in some slides by aggregates of green hornblende. The groundmass is pilotaxitic, and the albitic plagioclase crystals average about 0.1 millimeter in length. In some specimens, green hornblende is the chief mafic mineral in the groundmass; whereas in others, greenish biotite in flakes ranging from 0.02 to 0.03 millimeter in length are interstitial to the feldspar. In one slide, only chlorite and epidote comprise the groundmass. Tiny quartz crystals interstitial to the groundmass feldspar may be original or metamorphic. Minute magnetite crystals and apatite needles are common minor accessory minerals.

The intercalated basaltic flows are highly vesicular greenish black lava, containing sparse plagioclase phenocrysts in a finely crystalline groundmass. Microscopic studies reveal that the groundmass is pilotaxitic and albitic microlites 0.2 millimeter in length are separated by green hornblende needles. The scattered albite phenocrysts, 1 millimeter in length, contain sericite and epidote. Magnetite crystals are abundant in the groundmass.

The chemical analysis (table 5) shows that the Brindle Pup is andesitic in composition, slightly more silicic than the average andesite, but not so silicic as to warrant classification as dacite. The analyzed sample is porphyritic; albitic to sodic oligoclase phenocrysts, in places glomeroporphyritic, are embedded in a fine-grained pilotaxitic groundmass of sodic plagioclase and flakes of greenish biotite. Secondary epidote and sericite are present.

TABLE 5.—*Chemical analysis of Brindle Pup andesite*

Porphyritic andesite from near top of section of Brindle Pup andesite (1,320,900 N.; 443,800 E.). Rapid analysis by S. M. Berthold and E. A. Nygaard]

	Percent
SiO_2	62.6
Al_2O_3	15.4
Fe_2O_3	1.7
FeO	6.6
MnO	.24
MgO	2.1
CaO	3.2
K_2O	2.7
Na_2O	3.1
TiO_2	.64
P_2O_5	.22
Ignition loss	1.8

DACITE OF BURNT CANYON

Distribution, thickness, and stratigraphic relationship

The dacite of Burnt Canyon as defined here is exposed only south of Mingus Mountain where Burnt Canyon cuts into this unit (1,333,000 N.; 434,000 E.).

This dacite unit is about 2,000 feet thick in the widest outcrop and pinches out to the southeast. To the north, the dacite is covered by Paleozoic sedimentary rocks.

The dacite of Burnt Canyon appears to be at about the same stratigraphic position as the Brindle Pup andesite and Shea basalt; it separates the older Buzzard rhyolite from the younger Deception rhyolite. Because basaltic lava flows, similar to the Shea basalt, are intercalated between the Brindle Pup andesite flows, the dacite of Burnt Canyon, Brindle Pup andesite, and Shea basalt are of the same general age.

Lithology and internal structure

The dacite of Burnt Canyon contains conspicuous feldspar phenocrysts, ranging from 1 to 2 millimeters

in length, in a finely crystalline groundmass in which small quartz crystals are visible. Except for the quartz, the rock is generally similar to the porphyritic Brindle Pup andesite. Local vesicular facies are common, particularly in the southeastern exposures.

In thin section, the albite phenocrysts are embedded in a groundmass of quartz and alkalic feldspar whose crystals range from 0.05 to 0.1 millimeter in diameter. Microphenocrysts of embayed quartz are as large as 0.6 millimeter but average nearer 0.2 millimeter in diameter. Some spherulites of alkalic feldspar range from 0.1 to 0.2 millimeter in diameter. Streaks of sericite and subordinate chlorite occur in the groundmass associated with some calcite. Magnetite and apatite are accessory minerals. Sufficient lime is probably present in the calcite indicating that the original plagioclase was oligoclase or andesine, and that the rock is dacite.

The chemical analysis (table 6) indicates the dacitic composition of the lava of Burnt Canyon. The lime content is a little low for the average dacite, but too high for rhyolite. The other constituents fall within the usual range of dacite.

TABLE 6.—*Chemical analysis of dacite of Burnt Canyon*

[Porphyritic dacite west of Burnt Canyon (1,333,600 N.; 431,250 E). Rapid analysis by S. M. Berthold and E. A. Nygaard]

	Percent
SiO_2	69. 6
Al_2O_3	14. 0
Fe_2O_3	1. 8
FeO	2. 7
MnO	. 15
MgO	. 84
CaO	2. 1
K_2O	1. 5
Na_2O	4. 0
TiO_2	. 40
P_2O_5	. 15
Ignition loss	2. 7

DECEPTION RHYOLITE

Distribution

The Deception rhyolite is well exposed in the drainage basins of Mescal and Deception Gulches (1,362,200 N.; 440,000 E.), which is the name for the deep rugged part of Hull Canyon south of Jerome. A second area of exposed Deception rhyolite is south of Mingus Mountain between Ash Creek and Black Canyon.

The rhyolitic rocks in the Mescal Gulch area were given the noncommital term of greenstone by Reber (1922, p. 12), but he recognized lava flows, fragmental volcanic material, waterworn pebbles and sand grains, and sedimentary stratification. Reber also noted that, along Deception Gulch near the quartz porphyry contact, the nature of the rock was less certain. The appearance was similar to known volcanic rocks, but locally obscure features indicated the fragmental character of the rock. Reber reported that Finlay had named the rock Deception porphyry; probably this reference is to a private report, as we found no published account where Finlay introduced this term.

Later, Fearing (1926, p. 759) suggested that the Deception porphyry should be distinguished from the greenstone and on the basis of two sills of quartz porphyry, the older sill was related to the Deception porphyry. Lindgren (1926, p. 56) stated that in Deception Gulch, the greenstone is a quartz porphyry with typical phenocrysts of quartz and that it is older than the normal intrusive quartz porphyry of the district. Hansen (1930, p. 309) limited the term Deception porphyry to fragmental rhyolitic greenstone.

Our study indicates that the Deception porphyry, as exposed in Deception Gulch and in part of Mescal Gulch, is a hydrothermally altered facies of rhyolitic flows and fragmental rocks; the term is therefore modified to Deception rhyolite to include all rhyolitic rocks older than the Grapevine Gulch formation and younger than the Shea basalt, dacite of Burnt Canyon, Brindle Pup andesite, and Buzzard rhyolite.

Thickness and stratigraphic relationship

The most complete section of Deception rhyolite is exposed between Ash Creek and Black Canyon, where the outcrop is about 4,000 feet wide. The upper (southern) contact dips 60° SW., the lower contact 40° SW., and the average dip in this area is probably between 45° and 55°, indicating that the rhyolite in this section is about 3,000 feet thick.

In Mescal Gulch and north to Jerome, the Deception rhyolite is folded and faulted. East of the quartz porphyry mass in Mescal Gulch (1,355,800 N.; 437,700 E.), no large-scale folds or faults were recognized in an outcrop more than 4,000 feet wide; dips range from 40° to 60° NW., suggesting a thickness of about 3,000 feet.

In the Ash Creek drainage basin, the upper contact with the Grapevine Gulch formation presents some problems. Evidence is generally convincing that the Grapevine Gulch formation lies conformably above the Deception rhyolite, particularly along the contact east of the northward-trending granodiorite porphyry dike (441,400 E.). West of this dike, embayment of the Grapevine Gulch formation within the rhyolite may be interpreted as intertonguing, or may be the trough of a small syncline. However, proof of a fold by reversal of dips or by the outcrop pattern of beds, was not obtained, suggesting that the rhyolitic rocks do intertongue with the younger Grapevine Gulch sedimentary rocks. West of the embayment, there is a slight structural discordance between beds in the Grapevine Gulch formation and the contact with the Deception rhyolite. In part this discordance may represent a depositional contact of different facies on the older rocks, but less than a mile east of Kendall Camp, the discordance is

greater, and the occurrence of silicified zones may imply a fault relationship. North of Kendall Camp (1,333,000 N.; 429,500 E.), the Grapevine Gulch formation rests directly on the dacite of Burnt Canyon, whereas east of Kendall Camp (pl. 1, 1,327,500 N.; 436,500 E.), the dacite of Burnt Canyon is 2,000–3,000 feet stratigraphically below the contact between the Grapevine Gulch formation and Deception rhyolite. Because bodies of rhyolitic rock are commonly lenticular, the disappearance of the rhyolite northward beneath the Grapevine Gulch formation cannot be used as positive evidence of a fault relationship or unconformity.

In Hull Canyon in the northern exposures of Deception rhyolite (1,360,000 N.; 436,000 E.) the Grapevine Gulch formation dips and faces west, and overlies a small body of rhyolite, which is separated from the main rhyolitic outcrops by intrusive quartz porphyry. Tuffaceous interbeds in the rhyolite are essentially conformable to the overlying Grapevine Gulch formation.

Lithology and internal structure

Near Jerome, the outcrops of Deception rhyolite are a reddish hue, changing to buff and cream in Mescal Gulch, and are pale colored near Ash Creek.

Rhyolitic rocks in the Ash Creek drainage area generally are so similar to the Buzzard rhyolite, that a description of the megascopic and microscopic characters would be repetitious. A jasper-bearing facies of rhyolite is present at the top of the Deception rhyolite in the drainage basin of Ash Creek and in Hull Canyon, west of the intrusive quartz porphyry. Small outcrops of similar rhyolite are in the Buzzard rhyolite north of Oak Wash and about 1,500 feet west of the Verde fault. Much of the jasper-bearing rhyolite is brecciated; flow-banded fragments ranging from 1 to 3 inches in diameter are cemented by jasper. Some jasper-bearing rhyolite also has concentric banding brought out by jasper veinlets and minute streaks of hematite. These form ellipsoidal structures ranging from 2 inches to a foot in length on favorable outcrops. In Hull Canyon there is a weak columnar jointing in the jasper-bearing rhyolite, and some of the ellipsoidal banding is related to the jointing because it is only on surfaces essentially perpendicular to the columns. Probably alteration, controlled in part by jointing, is the cause of the concentric structure. Jasper veins 1–3 inches wide cut through the elliptically banded rhyolite. This jasper-bearing facies of the rhyolite is so distinctive that it is tempting to assume a single age, but its presence in the Buzzard rhyolite demonstrates that it has no stratigraphic significance. A small dike of similar jasper-bearing rhyolite, too small to plot on the geologic map (pl. 1), cuts the Texas Gulch formation 2¼ miles north of the Cherry Road, proving that this facies is not limited to lava flows.

In Mescal Gulch, the breccia contains interbeds of tuffaceous sandstone and cherty shale. Individual breccia beds range from 2 to 40 feet in thickness, and a particular bed generally contains angular to subrounded fragments of somewhat uniform size; the diameter ranges from an inch in some beds to a foot in others. A few breccia beds have little sandy matrix; others contain appreciable amounts.

The tuffaceous sandstone and cherty shale interbeds range from several inches to 2 feet in thickness. Stratification (pl. 3*E*) and graded bedding commonly indicated the direction that beds face.

Weathered outcrops reveal relict textures and structures not visible on unweathered surfaces, and some relict flow banding and fragmental textures in the Mescal Gulch area are only recognizable on weathered surfaces.

A minor facies of the Deception rhyolite in the Mescal Gulch area contains sericitized plagioclase phenocrysts in a microcrystalline groundmass. This facies may be intrusive in part, for locally it appears to be discordant to adjacent bedded breccias. Some bodies of this facies are from 50 to 200 feet wide and appear dikelike, but owing to the uncertainty of their origin and narrow width, they are not differentiated on the geologic map (pl. 1).

Massive and amygdaloidal andesitic or basaltic flows are interbedded with the Deception rhyolite north of Mescal Gulch. They resemble the underlying Shea basalt and undoubtedly represent late eruptions of mafic lava flows.

Five outcrops of andesitic or basaltic agglomerate are intercalated between the rhyolitic rocks in the Mescal Gulch area (pl. 1), but these may represent only one or two bodies of mafic ejecta, duplicated by folding or separated by structural movement. Similar agglomerate is exposed in Burnt Canyon (1,330,000 N.; 434,500 E.). The agglomerate forms dark-green to black outcrops in contrast to the adjacent rhyolite. The fragments are highly vesicular, irregular, and suggest andesitic or basaltic bombs and lapilli. The matrix is rich in chlorite and thin sections reveal some microgranular quartz and leucoxene associated with it. Two intersecting movement planes produced "pencils" in some of the chlorite-rich facies; in thin section, the planes are marked by intersecting flakes of brownish biotite.

Hydrothermally altered facies.—Most of the rhyolitic rocks exposed in Mescal Gulch and north to Jerome have been hydrothermally altered, the intensity of alteration increases to the north, so that in Hull Canyon

(Deception Gulch) the rhyolitic rocks are massive except for a weak northwestward-trending foliation. Relict flow and bedding structures are obliterated except locally. The rocks are greenish gray on unweathered surfaces, and are marked by irregular dark-greenish streaks. Some outcrops show angular to wispy black "fragments" or "spots" ranging from 1 to 2 inches in diameter. These may represent fragments in a clastic rock, flow breccia, or chloritized areas. Thin sections of these rocks are similar whether the outcrops suggest fragmental or flow structure.

The principal minerals are quartz, sericite, chlorite, and carbonate (calcite?). Larger relict quartz phenocrysts, or clastic grains, occur in rounded to angular grains about 1 millimeter in diameter scattered throughout the thin sections. These grains are also easily recognized megascopically. The quartz grains in the "groundmass" range from 0.01 to 0.1 millimeter in diameter; they are arranged in microgranular aggregates separated by pale-green chlorite and sericite in irregular patches and streaks. Epidote granules and apatite prisms are present in small amounts.

The chemical analysis of the Buzzard rhyolite (analysis 1), is repeated in table 7 for comparison with the Deception rhyolite from near Ash Creek (analysis 2). The higher SiO_2 content of the Deception rhyolite may be due to introduced quartz as veinlets, as a few were noted in thin section. Otherwise the two analyzed rhyolites are similar, and the unaltered Deception rhyolite is also sodic and properly should be termed soda rhyolite.

TABLE 7.—*Chemical analyses of Buzzard and Deception rhyolites, in percent*

	1	2	3	4	5
SiO_2	75.9	78.9	72.34	76.42	73.40
Al_2O_3	11.9	12.4	13.04	12.35	14.02
Fe_2O_3	1.3	1.4	.77	3.26	5.14
FeO	2.7	.30	4.64	.85	------
MnO	.10	.02	------	------	------
MgO	.43	.36	3.22	.26	3.45
CaO	.96	.45	.77	1.30	------
K_2O	1.2	1.8	.93	.81	------
Na_2O	4.4	3.5	.87	.97	------
TiO_2	.32	.22	------	------	------
P_2O_5	.02	.03	------	------	------
H_2O+	------	------	.65	.13	2.60
H_2O-	------	------	.11	.11	
Ignition loss	.86	.92	------	------	------
S	------	------	1.64	1.80	------
CO_2	------	------	1.15	1.90	------
			100.13	100.16	

1. Flow-banded Buzzard rhyolite, southern slope of Black Canyon (1,326,400 N.; 445,300 E.).
2. Porphyritic Deception rhyolite, containing abundant albite phenocrysts, near Ash Creek (1,319,700 N.; 444,800 E.). Analysis 1 and 2 are rapid analyses by S. M. Berthold and E. A. Nygaard.
3. Altered Deception rhyolite, south side of Deception Gulch (1,361,000 N.; 439,500 E.).
4. Altered Deception rhyolite, Deception Gulch. Exact location unknown.
5. Altered Deception rhyolite, lower end of Deception Gulch. Analyses 3, 4, and 5 furnished by Phelps Dodge Corp.

The analyses from Deception Gulch (table 7, analyses 3, 4, and 5) are the hydrothermally altered facies of Deception rhyolite. These analyses show an alkali content much lower than in normal rhyolite, indicating an appreciable leaching of the alkalies. Gains in iron content are indicated, and in two samples (table 7, analyses 3 and 5) appreciable grains in MgO, which is reflected mineralogically by the introduced chlorite. The low MgO content indicates that analysis 4 did not contain much chlorite. Introduced pyrite and carbonate are shown by the high S and CO_2 content in analyses 3 and 4.

GRAPEVINE GULCH FORMATION

Distribution

The Grapevine Gulch formation here defined is well exposed southwest of Mingus Mountain where it forms a broad northwestward-trending belt, cut off by the Shylock fault to the west and by younger quartz diorite to the south and east. Because of the excellent exposures along Grapevine Gulch (1,315,000–1,320,000 N.; 425,500 E.), the term Grapevine Gulch formation is used. A small section of the formation is exposed in Hull Canyon east of the Warrior fault (1,361,000 N.; 435,000 E.).

Thickness and stratigraphic relationship

The Grapevine Gulch formation rests on the dacite of Burnt Canyon southwest of Mingus Mountain (1,332,000 N.; 429,500 E.) but elsewhere it rests upon, and intertongues with, the Deception rhyolite. The top of the Grapevine Gulch formation is not exposed.

Southwest and south of Mingus Mountain, where the most complete section of the Grapevine Gulch formation is exposed, it appears in the trough of the Tex syncline, which plunges southeast. On the northeast limb of the syncline where the Grapevine Gulch formation rests on the dacite of Burnt Canyon, the thickness of the formation is about 8,000 feet. A greater thickness is indicated in the plunging trough of the syncline, but many minor structural irregularities may here make the apparent thickness much greater than the actual one. Probably 8,000–10,000 feet is a fair estimate.

Lithology and internal structure

Six map units were recognized in the Grapevine Gulch formation: (1) coarse-grained lithic tuffaceous sedimentary rock, (2) fine-grained tuffaceous sedimentary rock interbedded with chert and siltstone, (3) volcanic breccia, (4) jasper-magnetite beds, (5) intercalated dacitic flows that in part grade into dacitic intrusive masses, and (6) intercalated andesitic flows.

Lithic tuffaceous sedimentary rock.—The coarse-grained lithic tuffaceous sedimentary rocks generally are poorly bedded; the layers range from 4 to 50 feet in

thickness. The units shown on the geologic map (pl. 1) contain fine-grained tuffaceous interbeds ranging from 2 to 50 feet in thickness; some of the units contain about 50 percent of the fine-textured interbeds. The lithic tuffs are pale-greenish gray or greenish black where fresh and are buff colored where weathered. The fragments generally range from 0.1 to 1.0 inch in diameter, although locally, rare fragments are as long as 12 inches.

The lithic fragments consist of light-colored rhyolite and dacite and dark-brown to black andesite. No basaltic fragments resembling the Gaddes or Shea basalts were observed. In some localities, the fragments are dominantly andesitic, and the matrix is dark colored. Elsewhere they are dominantly rhyolitic, and the matrix is either light or dark colored. In many outcrops, andesitic, rhyolitic, and dacitic fragments occur together.

Microscopic studies reveal that many of the andesitic fragments contain minute plagioclase laths embedded in a black opaque groundmass, heavily charged with magnetite dust. The rhyolitic fragments are similar to the underlying Buzzard and Deception rhyolites, in that albite and quartz phenocrysts are embedded in a microcrystalline groundmass of quartz and alkalic feldspar. The dacitic fragments contain albite phenocrysts in a microcrystalline base of quartz, alkalic feldspar, chlorite, and epidote. The matrix of the lithic tuffs consists of variable amounts of quartz, angular albite locally charged with sericite, chlorite, epidote, greenish-brown biotite, and calcite. Magnetite-ilmenite grains are present in some specimens, but in others, leucoxene occurs and black opaque minerals are absent.

One variety of light-colored lithic tuff contains quartz grains ranging from 3 to 4 millimeters in diameter, but no similar quartz phenocrysts were observed in any of the associated lithic fragments. Presumably these large quartz grains were liberated from a rhyolitic magma during volcanic eruptions and distributed as crystal fragments.

Lithic tuffs generally are more andesitic in the upper part of the section, stratigraphically near the volcanic breccia unit, but andesitic fragments are also found in the lower beds near the dacite of Burnt Canyon and Deception rhyolite.

Fine-grained tuffaceous sedimentary rock.—The fine-grained tuffaceous and cherty sedimentary rocks are well-bedded, the layers range in thickness from ⅛ to 2 inches. Graded bedding is well defined in the coarser grained facies, and locally, crossbedding and channeling are present. In Hull Canyon and in the pit of the United Verde mine at Jerome, fine-grained tuffaceous beds are foliated (pl. 4*A*), but the lithic tuff beds are not. Minor folds from 2 to 20 feet across occur in the fine-grained tuffaceous rocks in Yaeger and Hull Canyons and in the pit (pl. 4*B*) and underground workings of the United Verde mine.

The crystal tuff beds consist of angular feldspar and subordinate quartz grains, from ½ to 1 millimeter in diameter, embedded in a fine-grained chloritic matrix. The feldspar weathers into conspicuous pale-yellow grains; variations in the size of the crystals accentuate the bedding. Thin-section study reveals that the feldspar grains are albitic and commonly sericitized; the matrix consists of microgranular quartz, albite, hornblende, epidote, chlorite, apatite, and ilmenite or leucoxene.

In some of the very fine grained bedded rocks, only scattered quartz grains represent apparently original clastic grains, and these are embedded in a microcrystalline aggregate of quartz, chlorite, and magnetite-ilmenite. In other specimens, chlorite is replaced by greenish-brown biotite.

The cherty rocks are dark gray and weather buff. In some places, the chert beds are 1 to 2 inches thick, separated by shaly interbeds, a quarter to half an inch thick. Thin sections of the chert reveal microcrystalline quartz containing variable amounts of greenish-brown biotite crystals, ranging from 0.005 to 0.01 millimeter in diameter. Fine lamination is distinct, and is caused by variation in the amount of biotite in the rock.

Adjacent to the quartz diorite, the tuffaceous beds are recrystallized to massive rocks which are intruded locally by many narrow dikes of quartz diorite. The texture of this massive facies is granular; green hornblende in granules or poikiloblastic grains separate granular quartz and plagioclase.

Volcanic breccia.—The breccia consists of angular to subangular fragments of lava, averaging from 2 to 4 inches in diameter, separated by a matrix of small lava chips and crystals. A few fragments are as large as 12 inches across. The fragments are dominantly porphyritic andesite containing numerous plagioclase phenocrysts from 1 to 2 millimeters long embedded in a microcrystalline groundmass ranging from pale gray to dark gray to reddish brown. In some fragments, aggregates of hornblende and chlorite represent pseudomorphs after original mafic phenocrysts.

Light-colored rhyolitic and dacitic fragments are scarce; they are similar to the flow rocks of similar character in the Grapevine Gulch formation and Buzzard and Deception rhyolites.

The structure of the breccia is massive, and no foliation has been observed. Microscopic study indicates some metamorphism; the plagioclase crystals are albitic and heavily charged with sericite and clinozoisite;

and bluish-green hornblende, chlorite, and epidote and secondary quartz are common in the matrix.

Amygdaloidal andesitic flows are intercalated near the base of the breccia. Phenocrysts of plagioclase and augite, 1 millimeter long, are embedded in a light-green microcrystalline groundmass. Thin sections reveal that the plagioclase is altered to sericite and clinozoisite, and the augite is pale-greenish gray, partly altered to chlorite. The groundmass consists of minute tablets of plagioclase separated by an indeterminate greenish-gray base.

The breccia unit intertongues at its base with the tuffaceous sedimentary rocks. The breccia unit must be lenticular as it is exposed only in the west end of the trough of the Tex syncline, and although the east margin of the breccia is in fault contact with the tuffaceous rocks, the displacement must be small as the fault cannot be traced for a great distance. The breccia unit is not the youngest unit in the Grapevine Gulch formation; in the trough of the southeastward-plunging syncline younger rocks are exposed.

Jasper-magnetite beds.—The jasper-magnetite beds occur throughout the Grapevine Gulch formation, and have no stratigraphic significance. In the southeastern exposures, east of Ash Creek, these beds are intercalated between dacitic flows. The beds generally range from 1 to 10 feet in thickness, but locally reach 30 feet. In places the beds contain as much jasper as magnetite, and grade along the strike into pure jasper. Other beds contain interbeds, a fraction of an inch thick, of fine-grained tuffaceous sedimentary rock. Some persistent jasper-magnetite beds are as much as a mile long, but others are much shorter.

The beds are black except where jasper replaces magnetite, and in favorable outcrops the beds form ribs that stand out above the adjacent rock. Mapping of individual beds has helped in deciphering local structure.

Polished surfaces reveal that magnetite is the principal opaque mineral, appearing as ramifying streaks and bands, separated by microcrystalline quartz, 0.01 millimeter in diameter, or by very fine grained, aggregates of quartz and albite that range in diameter from 0.05 to 0.1 millimeter. In the feldspathic facies, chlorite and pale-brown biotite are conspicuous accessory minerals. Individual magnetite grains are in the matrix. Hematite is scarce, appearing as separate streaks adjacent to magnetite, or as microscopic intergrowths.

Jasper-magnetite beds generally resemble the Precambrian iron formation of the Canadian shield (Leith, and others, 1935, p. 21), except that quantitatively, the magnetite beds are unimportant in the Mingus Mountain quadrangle. The jasper-magnetite beds probably represent sedimentary deposits in which the iron and silica formed as chemical precipitates. The magnetite may have formed by later metamorphism of some other iron mineral (Gunning, 1937, p. 10), or perhaps it was deposited as primary magnetite (James, 1954, p. 263).

Dacitic flows and intrusive masses.—The intercalated dacitic lava flows crop out extensively in the most southeasterly exposures of the Grapevine Gulch formation; these flows are probably 2,000 to 3,000 feet thick. A much smaller thickness of dacitic flows is exposed near Kendall Camp, and several separate flows have been mapped west and south of the camp (1,320,000–1,330,000 N.; 420,000–430,000 E.).

The dacitic lava is gray to dark gray on unweathered surfaces; cream-colored plagioclase phenocrysts, from 1 to 3 millimeters in length, are conspicuous on the buff to gray weathered outcrops. Quartz crystals from 0.1 to 0.2 millimeter in diameter appear in the microcrystalline groundmass. Vesicular and amygdaloidal facies are common, but most of the lava is massive.

Thin sections reveal that the plagioclase phenocrysts are albitic, containing a variable amount of clinozoisite, which indicates an original plagioclase more calcic than albite. Original mafic phenocrysts are outlined by clots of chlorite or brownish-green biotite. The groundmass contains interlocking alkalic feldspar and quartz, locally having a micrographic texture. Locally the quartz is in streaks separating alkalic feldspar, suggesting some silicification. Other minerals in the groundmass are epidote, chlorite, brownish-green biotite, and locally, green hornblende. Apatite is a common accessory mineral. Sphene occurs in specimens containing no black opaque mineral, indicating that ilmenite is the black opaque accessory mineral in other specimens that contain no sphene.

Two intrusive masses of dacite were recognized; one to the east of Kendall Camp (1,327,000 N.; 431,500 E.) and the other to the east of Ash Creek (1,315,000 N.; 447,000 E.). The intrusive character of these two masses is clearly demonstrated by the crosscutting of the adjacent sedimentary beds in the Grapevine Gulch formation. The mass near Kendall Camp grades into a flow intercalated between the tuffaceous sedimentary rocks, indicating that these two intrusive masses are volcanic plugs filling old vents.

The intrusive dacite resembles the dacitic flow rock except that the plagioclase phenocrysts are larger, ranging from 3 to 6 millimeters in length, and the structure is massive. Thin sections show that the groundmass is slightly coarser grained than in the lava, and quartz grains from 0.1 to 0.2 millimeter in diameter are more

abundant. Clots of green hornblende or chlorite probably represent original mafic phenocrysts.

The chemical analysis (table 8) shows that the dacite of the Grapevine Gulch formation varies slightly from the average dacite by a higher content of FeO and Na_2O, and by a lower content of CaO and K_2O. The analysis indicates, however, that the rock is sodic dacite, in keeping with the sodic nature of the rhyolite deposits in the Ash Creek group.

TABLE 8.—*Chemical analysis of dacite in Grapevine Gulch formation and of average dacite, in percent*

	1	2
SiO_2	65. 5	65. 68
Al_2O_3	15. 2	16. 25
Fe_2O_3	3. 0	2. 38
FeO	4. 4	1. 90
MnO	. 13	----
MgO	1. 3	1. 41
CaO	2. 0	3. 46
K_2O	1. 3	2. 67
Na_2O	5. 6	3. 97
TiO_2	. 57	. 57
P_2O_5	. 20	. 15
Ignition loss	. 95	----

1. Dacitic flow, Ash Creek (1,303,000 N.; 440,700 E.). Rapid analysis by S. M. Berthold and E. A. Nygaard.
2. Average dacite, Daly (1933, p. 15).

ALDER GROUP

The Alder group is exposed along the west margin of the Black Hills, in the outlying hills in Lonesome Valley, and in the hills marking the south margin of the valley. It is separated from the Ash Creek group by the Shylock fault. The rocks have been divided into six formations, which fall into two classes: those of probable known stratigraphic succession, and those of unknown stratigraphic succession. The former comprises three volcanic sequences (from oldest to youngest)—the Indian Hills, the Spud Mountain, and the Iron King; the latter is divided into the Green Gulch, the Chaparral volcanics, and the Texas Gulch formation. Each of the formations contains two or more lithologic subdivisions; all the formations have a volcanic source except the slate unit of the Texas Gulch formation, which is probably terrigenous; and the composition of the group ranges from rhyolitic to basaltic.

The thickness of the Alder group is not known; it appears to be 20,000 feet or more, and may be as much as 30,000 feet.

Wilson (1939, p. 1121 and 1158) first applied the term "Alder series" to these rocks in the Jerome area because of the lithologic similarity of our Texas Gulch formation to his Alder series as exposed in the type sections in the Mazatzal Mountains, about 60 miles southeast of the Jerome area. A map by Wilson (1939, p. 1158) of part of the Jerome area shows that his Alder series includes other formations lithologically distinct from the Texas Gulch formation, indicating that Wilson included all the volcanic and sedimentary rocks west of the Shylock fault in his Alder series. Our work has demonstrated that one unit of the Spud Mountain volcanics is lithologically similar to the Texas Gulch formation, indicating that in a broad sense, correlation of the diverse formations in the Jerome area with the Alder series of Wilson (1939) in the Mazatzal Mountains has merit. It seems desirable therefore to retain Wilson's term of Alder for these rocks in the Jerome area, changing it to group rank because of our subdivision of it into six formations.

The rocks of the Alder group are foliated, and the texture and mineralogic assemblages are typical of regionally metamorphosed rocks belonging to the greenschist facies. Chlorite, actinolitic hornblende, clinozoisite (zoisite), albite, sericite, and quartz are the dominant metamorphic minerals; microcline is a common stable relic in rhyolitic rocks.

In metamorphic terranes characterized by pronounced foliation and isoclinal folds, the normal approach to determine stratigraphy cannot be used. Stratigraphy and structure are determined concurrently through mapping of distinctive lithologic units whether they are beds, zones, or formations. Structural elements and data on the direction that tops of beds face are diligently searched for in all exposures. In this manner, the stratigraphy and structure unfold together, and the final interpretation of the geology results from the integration of the stratigraphic and structural information.

During the mapping of the Alder group, structures indicating the direction that tops of beds face were so few and scattered that, in most places, one could not draw a distinction between normal and inverted sequences. It was not until all mapping in the Alder group was completed that we decided upon the most probable order of superposition. It should be emphasized that the proposed stratigraphy of the Alder group is based on too few facts (pl. 6), and additional mapping in the Alder group south of the Jerome area may indicate the need for revision. But we believe that with the information at our disposal, a stratigraphic and structural interpretation, however tentative, has some merit.

INDIAN HILLS VOLCANICS

Distribution

The Indian Hills volcanics as here defined is named from the exposures in the low hills of this name that protrude through the Cenozoic rocks along the western front of the Black Hills. The only other major outcrops of this formation are 2 miles to the north in the low hills west of the Coyote Spring ranch.

Thickness and stratigraphic relationship

The thickness of the Indian Hills volcanics is not known, for the base of the formation does not crop out, and the rocks dip at unknown angles. Judging by the attitude of the adjacent formations, the dip of the formation is presumably steep and the outcrop widths approximate thicknesses. The outcrops in the hills west of the Coyote Spring ranch reveal a maximum width of 6,500 feet. The absence of stratigraphic subdivisions of the Indian Hills does not permit recognition of small folds, so that duplication of flows probably would not be noticed.

The stratigraphic relationship of the Indian Hills volcanics is not firmly established, but the stratigraphic relationship postulated harmoniously fits the known data, and permits integration of the Indian Hills into the structural fabric of this highly deformed terrane. The contact between the Indian Hills and Spud Mountain volcanics in the Indian Hills area appears conformable, and the presence of an intercalated rhyolitic flow northeast of the main contact further supports this interpretation. Here, however, as we do not know the direction that beds face, we rely upon stratigraphic relationship in other areas.

Along the south margin of the map area (near 1,267,000 N.; 391,000 E.), two determinations of the directions that beds face suggest that the Spud Mountain volcanics underlies the Iron King volcanics. Evidence for this relationship occurs near the Indian Hills (1,317,000 N.; 416,000 E.); here the Iron King faces east and lies east of the Spud Mountain. In the Indian Hills, therefore, it is most likely that the eastern part of the Spud Mountain faces east. Once this probability is accepted, it is equally probable that the Indian Hills volcanics is older than the Spud Mountain. The reason is this: reversal of the direction that the Spud Mountain volcanics faces in the Indian Hills through a major fold is not likely because the Indian Hills volcanics and the Iron King volcanics are different and therefore cannot be correlated. Because of the probability of an east facing section in the Indian Hills, the conformity of the Spud Mountain and the Indian Hills suggests that the latter is older.

The base of the Indian Hills volcanics does not crop out, thus its relationship to older rocks is in doubt. Presumably, the Texas Gulch formation underlies it, but whether directly or whether other rocks intervene is not known.

Lithology and internal structure

The Indian Hills volcanics is divided into two interbedded units: one andesitic and basaltic, the other rhyolitic. These units are, however, lithologic and not stratigraphic divisions.

The Indian Hills is less foliated than any other formation of the Alder group. Local zones common enough to give the regional trend are foliated, but they are by no means ubiquitous, as in the Texas Gulch formation of the Alder group. Local zones common for the absence of a pervasive foliation is difficult to determine. Perhaps such a thick series of flows possessed sufficient strength to resist deformation, or perhaps stresses may have been weaker in this local area.

The andesitic and basaltic flows are greenish black. They generally are blocky, but are cut by local zones in which the rock is somewhat fissile. Amygdules, composed of members of the epidote group, and vesicles are abundant. The rocks are porphyritic with a holocrystalline, but commonly aphanitic, groundmass. The phenocrysts, as much as three quarters of an inch long, are altered plagioclase and subordinate hornblende. The matrix, as seen in thin section, comprises granular albite, granular members of the epidote group, and small flakes of chlorite, which show a poor preferred orientation. The groundmass has been completely reconstituted to a metamorphic assemblage, and its texture is metamorphic. The phenocrysts, however, are probably altered relics of the original rock.

The rhyolitic flows are very pale orange or greenish gray on the fresh surface, and grayish orange-pink or pale yellowish brown on the weathered surface. The rocks are largely massive, but locally foliated. Flow banding is widespread, but not conspicuously abundant. Amygdules and vesicles mark local areas, and secondary veinlets of quartz appear characteristic.

The rhyolitic flows are porphyritic, but inconspicuously so. The groundmass is aphanitic; in places, it appears siliceous to the point of being flinty. The scattered phenocrysts generally range from 0.5 to 0.75 millimeter in maximum dimension, but in the few flows where phenocrysts are abundant, they are as much as 2½ millimeters long. They consist chiefly of quartz and albite, but a few phenocrysts of orthoclase also occur. The matrix consists of a microcrystalline granular aggregate of quartz, alkalic feldspar, and small amounts of sericite. Minor amounts of chlorite and members of the epidote group are accessory minerals. In the zones of sheared rhyolite, sericite is abundant.

SPUD MOUNTAIN VOLCANICS

Distribution and thickness

The Spud Mountain volcanics defined here comprises four lithologic units: andesitic breccia, andesitic tuff, rhyolitic tuff, and andesitic and basaltic flows (pl. 1). A representative section is exposed in the area shown on the southwest corner of plate 1, from Spud Mountain to the Iron King mine (1,273,000 N.; 390,000–397,000

E.). The name Spud Mountain is adopted for this volcanic formation because of the excellent exposures on and near this mountain. A complete section of the Spud Mountain volcanics is nowhere exposed in the Jerome area, preventing determination of the total outcrop width or stratigraphic thickness from any one continuous section. The consistently high dips of the Spud Mountain volcanics suggest that their outcrop widths may about equal the stratigraphic thickness for local or partial sections, but small folds and thinning through regional deformation generally obscures the stratigraphic thickness.

Andesitic breccia.—The andesitic breccia west of the Iron King mine occurs in a northeastward-trending zone 3 miles long and about 5,500 feet wide; some beds may be duplicated by folding. This zone passes beyond the map area to the southwest, and beneath the Cenozoic rocks in Lonesome Valley to the northeast. The zone reappears from beneath the Cenozoic rocks on the east side of the valley in the Indian Hills (1,324,000 N.; 410,000 E.), where the maximum outcrop width is also about 5,500 feet. From the Indian Hills northward to within 1½ miles of the northern boundary of the map area, scattered outcrops of andesitic breccia protrude through the Quaternary gravel beds.

A large area underlain by andesitic breccia is between the Texas Gulch formation and the quartz diorite about 2½ miles east of Humboldt (pl. 1). This andesitic breccia crops out in a wedge-shaped zone that trends northward for about 8 miles from the southern boundary of the map area. The outcrop width narrows from 5,000 feet at the southern boundary of the map area to a wedge line at the northernmost outcrop. The true thickness of this breccia is not even approximated by outcrop widths, for both the east and west contacts are faults, and the zone contains small folds.

Andesitic tuffaceous rock.—The distribution of the andesitic tuffaceous rock is like that of the andesitic breccia; zones of andesitic tuffaceous rock are near the Iron King mine, east of the Indian Hills, and southeast of Humboldt. The tuffaceous rock near the Iron King mine trends northeastward for a strike length of 1½ miles, and ranges in outcrop width from 1,200 to 1,700 feet. To the south, it passes beyond the map area. To the north, the tuffaceous rock is concealed by the Cenozoic rocks in Lonesome Valley, but reappears east of the Indian Hills. Here the maximum outcrop width is about 5,000 feet, including a zone of rhyolitic tuff, 1,000 feet wide; we believe that this section is duplicated by folds. Near the Indian Hills, the andesitic tuffaceous rock crops out in a northward trending belt from Grapevine Gulch (1,312,500 N.; 416,000 E.) northward for about 7 miles. To the north, the zone is covered by Paleozoic and Cenozoic rocks and to the south by Cenozoic rocks.

The zone of andesitic tuffaceous rock southeast of Humboldt trends slightly east of north, and ranges in outcrop width from 7,000 to 9,000 feet, including a belt of rhyolitic tuff 600 feet wide. Cenozoic rocks in Lonesome Valley conceal the northward extent of the zone except for a band, 1,000 feet or less wide, that extends northward to a point near the Cherry road (1,286,000 N.; 413,000 E.). Beneath the Cenozoic rocks, this belt probably joins the andesitic tuffaceous rock east of the Indian Hills. Southward, this zone passes beyond the limits of the map (pl. 1).

Three large fault-lenses of andesitic tuffaceous rock occur within the Texas Gulch formation. They range in outcrop width from a point to 1,500 feet, and in length from 1 to 5 miles. The most northerly one is east of the Indian Hills, and the two southerly ones are east and northeast of Dewey. The latter two are probably continuous beneath the Cenozoic fill of Lonesome Valley.

Rhyolitic tuff.—Belts of rhyolitic tuff are widespread in the Spud Mountain volcanics: east of the Iron King mine, east of the Indian Hills, and east of Humboldt. Rhyolitic tuff and conglomerate beds crop out about 500 feet southeast of the Iron King mine in a zone about 200 feet wide that trends northeastward from the southern boundary of the map area for 6,500 feet before passing beneath the Cenozoic rocks in Lonesome Valley. A zone of rhyolitic tuff, which ranges from a few feet to 1,000 feet in width, crops out from a point about 2 miles southeast of Indian Hills northwestward to the Paleozoic overlap north of Highway 89A. At least one other rhyolitic tuff bed occurs in the andesitic tuffaceous rocks southeast of Indian Hills, but it is too narrow to show on the scale of plate 1. Another belt of rhyolitic tuff, which is displaced by an oblique fault, occurs intercalated in the andesitic tuffaceous rocks about 1 mile east of Humboldt. This belt trends slightly east of north for about 2 miles and is from 350 to 650 feet wide. To the north, it is concealed by the Cenozoic rocks in Lonesome Valley.

Andesitic and basaltic flows.—The largest mass of andesitic and basaltic flows of the Spud Mountain volcanics forms a northward-trending belt intercalated in the andesitic tuffaceous rocks north of Indian Hills (1,343,500 N.–1,339,500 N.). It has a strike length of about 4,000 feet and a maximum width of 1,500 feet. In addition, two thin flows occur in the breccia unit about 3,000 feet west-southwest and 6,000 feet southwest of the Iron King mine. Both are less than 400 feet wide and are about 3,500–4,000 feet long.

Minor andesitic and basaltic flows occur elsewhere in the Spud Mountain volcanics, but they are not distinguished on the map (pl. 1). The general distribution of these flows is indicated on plate 1 by the symbols for pillow lavas. East of Humboldt, a flow occurs in the andesitic tuffaceous unit about 600 feet west of the contact with the rhyolitic tuff. It trends northeastward, is about 100 feet wide, and occurs in two gulches 3,500 feet apart. As no attempt was made to map it, the northern and southern limits were not determined. Possibly other thin flows in the andesitic breccia and andesitic tuffaceous rocks were not recognized, but it is believed that any section as much as a few hundred feet thick would not have been missed.

Stratigraphic relationship

The Iron King volcanics apparently overlies the Spud Mountain volcanics, and the Indian Hills volcanics underlies them. The basal contact of the Spud Mountain with the Indian Hills volcanics crops out in the Indian Hills. It apparently is gradational, for a rhyolitic flow is intercalated in the Spud Mountain about 500 feet above the main contact. The upper contact of the Spud Mountain with the Iron King volcanics, shown on the south-central margin of plate 1, is conformable and appears gradational, but this may be because of shearing along the contact.

The lower part of the Spud Mountain volcanics appears to be andesitic breccia. This grades upward into andesitic tuffaceous rocks, which comprise the great bulk of the upper part of the formation and which contain intercalated rhyolitic tuff.

We regard the rhyolitic tuff as a lithologic facies of dubious correlative value, but wish to point out the correlations permissible on the basis of lithologic similarities and to point out the chief obstacles to such correlations. Rhyolitic tuff intercalated in the andesitic tuff occurs in three areas: about 500 feet southeast of the Iron King mine, about 5,000 feet east of Humboldt, and east and northeast of the Indian Hills. Whether any two of these are the same stratigraphic zone was not determined with certainty, but as the rhyolitic tuff is similar, correlations merit consideration. The rhyolitic tuff southeast of the Iron King mine is lithologically similar to the southern part of the rhyolitic tuff in the east edge of the Indian Hills. However, the rhyolitic tuff near the Iron King is at the top of the Spud Mountain volcanics, whereas the rhyolitic tuff near the Indian Hills is overlain by about 1,000 feet of andesitic tuffaceous rock, in fault contact with the Texas Gulch formation. The rhyolitic tuff east of Humboldt is lithologically similar to the northern part of the rhyolitic tuff in the east edge of the Indian Hills. The tuff near Humboldt lies about 1,500 feet below the top of the Spud Mountain volcanics, a stratigraphic position similar to the tuff east of the Indian Hills. If the discrepancies of the stratigraphic position of the rhyolitic tuff east of the Iron King mine can be resolved through original lenticularity of the andesitic tuffaceous rocks or through removal of hundreds of feet by shearing, tentative lithologic correlation of the rhyolitic tuff in these areas is permissible. We believe that neither this degree of lenticularity nor the necessary intensity of shearing would be exceptional in the rocks and structures of the Jerome area.

The andesitic and basaltic flows are lithologic units only. They occur intercalated in the andesitic tuff and in the andesitic breccia.

Lithology and internal structure

The color of Spud Mountain volcanics depends largely on abundance of ferromagnesian minerals, chiefly chlorite. Where chlorite is abundant, such as in some fine-grained tuffaceous sedimentary rocks and in some crystal tuff, the rock is dark-grayish green on fresh surfaces, and weathers to a dark reddish brown. Where chlorite is less abundant, rock is light grayish green and weathers light tan or buff. Feldspathic rhyolitic tuff is characteristically gray with a greenish cast owing to sparse disseminated chlorite. The rhyolitic tuff northeast of the Indian Hills is white or light cream, mottled with pinkish- to brick-red stain, owing to disseminated iron oxide derived from oxidation of carbonate and probably pyrite. The fine-grained light colored tuff southeast of Humboldt is white, cream or very light green, spotted and streaked with brick red iron oxide on fresh fracture, but is commonly deep red to nearly black on weathered surfaces.

Foliation varies in the Spud Mountain volcanics, particularly in breccia facies. Breccia west of the Iron King mine and in the Indian Hills is foliated but generally does not cleave readily parallel to that structure, owing to predominance of fibrous actinolitic hornblende over chlorite. Breccia east of the Texas Gulch formation is well foliated toward the north, where it is essentially a fine-grained chloritic schist containing white streaks that probably represent original clastic feldspar grains, phenocrysts, and leucocratic rock fragments. South of latitude 34°30′ foliation is obscure in breccia beds, and the original character of the breccia is evident; interbeds of tuffaceous rocks, however, are foliated. The transition from foliated to chiefly nonfoliated breccia takes place in about 2,000 feet parallel to the strike. Fine-grained tuffaceous facies produce conspicuous foliation—the finer the grain size, the more pronounced the foliation.

Attenuation of fragments in breccia facies is extremely variable, even within local areas and from bed to bed.

Fragments in some beds are not perceptibly deformed, but in others the length is commonly 10 times greater than the width and thickness. In the well-foliated breccia east of the Texas Gulch formation, fragments are visible only as "ghosts" in widely scattered outcrops. Fragments are commonly difficult to discern in poorly foliated outcrops where the matrix has essentially the same composition and grain size as the fragments.

Relict bedding occurs in all belts of Spud Mountain volcanics, but it is much more common in areas where foliation is weak. It is most common in andesitic and rhyolitic tuffs, and is particularly discernible in water-worn outcrops in the larger gulches. The strike of beds generally parallels the northerly to northeasterly trend of the belts of Spud Mountain volcanics, and dips are high eastward and westward, but chiefly westward. Graded bedding, manifest by distribution and size of saussuritized plagioclase grains, occurs sporadically in tuffaceous interbeds.

Breccia facies of Spud Mountain volcanics consist of interbedded breccia beds—some probably more than 500 feet thick—fine-grained tuffaceous sedimentary rocks, and granular crystal tuff. Breccia predominates. Fragments range from less than 1 to about 18 inches in length and are commonly subrounded, although many are so deformed that the original shape is conjectural. Fragments vary in size in different beds, but within individual beds, the size of most fragments is uniform.

Fragments are chiefly andesite but rhyolite is common, especially in local areas. Other fragmentlike masses, whose original composition is obscure, are composed largely of quartz and epidote; possibly they may be metamorphic. Almost all andesitic fragments are porphyritic and some are conspicuously porphyritic with somewhat equant saussuritized plagioclase in clusters and individual crystals. In certain beds fragments are highly vesicular and amygdaloidal. The matrix between fragments is largely andesitic, abounding in equant plagioclase. Crystal tuff matrix was observed in some beds where the fragments are chiefly nonporphyritic. Beds composed of rhyolitic fragments in a chloritic matrix occur in the western part of the breccia west of the Iron King mine. In the breccia facies east of the Texas Gulch formation, similar rocks are associated with rhyolitic fragments in rhyolitic matrix and with massive rhyolite of uncertain origin. The rhyolitic fragments and the massive rhyolite are light colored and generally felsitic, and in places contain small quartz phenocrysts. Tuffaceous interbeds are indistinguishable from the tuffaceous facies of Spud Mountain volcanics.

In the andesitic breccia east of the Texas Gulch formation, several purple slate interbeds and a zone that contains several 10-foot thick marble beds interrupt the breccia sequence. The largest purple slate interbed is 50 feet wide by 300 feet long. West of the Iron King mine in the first zone of andesitic breccia, an iron-jasper bed, 10 feet wide and 1,000 feet long, was noted.

East of the eastern strand of the Shylock fault, just north of line 1,280,000 N., breccia facies of the Spud Mountain volcanics on the margin of the quartz diorite has been metamorphosed to hornfels. In part, the hornfels retains the identity and foliation of the breccia, although it is generally no longer fissile. The breccia and crystal tuff beds still stand out in sharp contrast to the intercalated finer grained tuffs, owing to preservation of the large equant saussuritized plagioclase grains. Apparently the finer grained rocks and the finer grained fraction of the breccia and crystal tuff were more susceptible to recrystallization than the large saussuritized plagioclase. The recrystallized parts of these rocks consist of a compact fine-textured mass of interlocking crystals. Metamorphic hornblende and brown biotite replace chlorite as the dominant mafic constituent, although minor amounts of chlorite remain. The reconstituted plagioclase is granoblastic albite or oligoclase, but much saussuritized plagioclase still remains, and is commonly flecked with sericite. Granoblastic quartz is abundant.

Andesitic tuffaceous rock.—The transition from the andesitic breccia to the overlying andesitic tuffaceous rock is gradational. For nearly 1,000 feet above the contact, which is arbitrary, conglomeratic beds as much as 10 feet thick occur sporadically, especially in thicker beds of crystal tuff. In the upper part of the andesitic tuffaceous rock, coarse-grained crystal tuff and associated conglomerate are limited to a few scattered beds, in which the fragments are of pebble or cobble size.

The andesitic tuffaceous rock is essentially a sedimentary rock formed predominantly of andesitic detritus. The tuff ranges in color from grayish green, through dusky-yellow green, to grayish-yellow green. The color appears to be a function of mineral composition and grain size; the smaller the grain size the lighter the hue. The fragments range from the size of cobbles to that of clay, but most range between coarse sand and silt. The grain size varies abruptly across the strike, and commonly beds of uniform grain size are only a few inches thick. This feature helps greatly to distinguish the andesitic tuffaceous rocks, for it strongly implies stratification. In individual exposures, however, the possibility of differential shearing cannot be eliminated. Variation in the average grain size in the sequence from one place to another is difficult to

ascertain, but we have the impression that the top of the unit is finer grained.

Megascopically the andesitic tuffaceous rock consists chiefly of chlorite and saussuritized plagioclase; quartz and to a lesser extent sericite are common, but by no means always present. The saussuritized plagioclase dominates the megascopic constituents, for it is the most conspicuous mineral; its distribution has produced the obvious heterogeneities in the rock, such as bedding. The quartz, chlorite, sericite, other plagioclase, and other constituents together form a granular aggregate that comprises the matrix or background for the pseudoporphyritic saussuritized plagioclase.

The finer grained facies, which are not uncommon, are phyllite or slate. In most outcrops no individual minerals are conspicuous, although commonly small grains of quartz or saussuritized plagioclase can be resolved with a hand lens. The most conspicuous feature of these rocks is the sheen on the cleavage surfaces, the product of the chlorite and to a lesser extent, the sericite. Presumably because of their fine grain these rocks have formed a very pronounced and regular fissility.

Minor beds of rhyolitic tuff and associated marble are intercalated in the andesitic tuffaceous rock in zones too small to show on the maps. In general, these are of two types: (1) the coarse-grained rhyolitic tuff and conglomerate, purple slate, and associated marble, like those of the Texas Gulch formation; and (2) the very fine grained rhyolitic tuff, which appears much like porcelaneous slate. Type 1 occurs in two beds not separately mapped in the region southeast of the Indian Hills; type 2 is more abundant in the andesitic tuffaceous rock along the southern border of the map area.

Rhyolitic tuff.—The rhyolitic tuff intercalated in the Spud Mountain volcanics is of two lithologic varieties; one coarse-grained and feldspathic, with intercalated conglomeratic and marble beds; the other fine-grained, finely laminated, and dense.

The feldspathic rhyolitic tuff is like the rhyolitic tuff in the Texas Gulch formation. It is very light gray, but commonly has a greenish cast, owing to small amounts of disseminated chlorite. Medium- to coarse-grained rhyolitic tuff predominates, but minor amounts of fine-grained slate or phyllite occur in thin beds. Interbeds of conglomerate and marble, like those in Texas Gulch formation, make up a small part of the rhyolitic tuff southeast of the Iron King mine and southeast of the Indian Hills. Quartz, potash feldspar, albite, and sericite are the chief constituents, and generally all are megascopic; in some specimens a little disseminated chlorite also can be detected. The quartz and feldspar are commonly in lenticular grains or granular aggregates around which the interwoven oriented flakes of sericite produce a minutely wavy foliation. The pebbles and cobbles in the conglomerate, which generally are attenuated, consist of jasper, quartz, porphyry, and other rocks in a matrix of granular rhyolitic tuff. The marble is pinkish, gray, or very light gray; it is granular and shows flow structures.

The fine-grained rhyolitic tuff is characterized by its grain size, thin bedding, and an abundance of fine lamination, which apparently is stratification, further emphasized in most places by a parallel foliation. East and northeast of Indian Hills, this rock is sericite slate, so white and dense that it appears porcelaneous. In the slate west of the Yaeger mine, zones as much as 5 feet wide were altered by carbonization, silicification, and pyritization. The quartz occurs as a general impregnation, and in ramifying veinlets with pyrite. Whether any of the widespread sericite and disseminated quartz that give the rock the porcelaneous aspect are hydrothermal was not determined, but the possibility cannot be dismissed.

The fine-grained rhyolitic tuff southeast of Humboldt has several distinctive features. Beds are characteristically thin, and internal stratification abundant. The color ranges from very light gray to dusky-yellow green, depending on the ratio of sericite to chlorite. The most characteristic feature is abundant, chiefly disseminated, magnetite which is probably a metamorphic mineral derived from ferruginous sediments. The possibility that it was introduced however cannot be dismissed, for veinlets containing quartz, carbonate, and pyrite(?) cut these rocks, and attest to some hydrothermal alteration. This type of alteration commonly produces sericite also, but in these rocks hydrothermal sericite would be difficult to distinguish from metamorphic sericite.

The porcelaneous slate east and northeast of the Indian Hills is characterized by sericite and quartz. Carbonate and, in some rocks, sparse chlorite are accessory constituents. The rhyolitic tuff east of Humboldt contains sericite, chlorite, biotite, garnet, quartz, albite-oligoclase, carbonate, and magnetite. Typical assemblages are: sericite-quartz-carbonate, which may be due to hydrothermal alteration, chlorite-biotite-albite or oligoclase, and biotite-sericite-quartz-garnet. The index of refraction of the garnet is 1.745, indicating a dominance of grossularite. The presence of biotite and garnet resulted from a higher grade of metamorphism locally along the south margin of the map area.

Sericite-albite-quartz characterizes the feldspathic rhyolitic tuff southeast of the Iron King mine. Traces of clinozoisite or zoisite and of chlorite are also present.

Basaltic and andesitic flows.—Preservation of the original textures and structures of the basaltic or andesitic flows ranges from excellent to poor. In some

outcrops, the presence of pillow lavas, vesicles, amygdules, and flow breccia tops clearly reveal the rocks as flows; in others, however, such relicts are absent, and other features must be used to distinguish tuffaceous rocks from flows. Flows in the Alder group generally are massive, uniform in appearance, and lack the abrupt changes in grain size and texture every few inches or feet so characteristic of tuffaceous beds.

In a few samples, thin sections of flows reveal a relict porphyritic and felty (pilotaxitic) texture. Amygdules comprising minerals stable under conditions of moderate stress and temperature, chiefly quartz, epidote, and carbonate, remain as "eyes," even where stress reduced the massive rock to a schist. Relict breccia structure, presumed to be flow breccia, was spectacularly preserved along the east margin of the flow lying west of the Iron King mine. Abundant amygdules and vesicles(?) occur in both the fragments and the matrix of the breccia. In local areas, pillow lavas survived the regional deformation, and they are noted on plate 1 by a symbol. The deformation necessary to destroy a pillow structure was incredible; probable pillow structures were noted whose lengths exceeded widths by 15 to 20 times.

The megascopic minerals of the lava comprise chlorite, fibrous hornblende, saussuritized plagioclase, and members of the epidote group. The saussuritized plagioclase grains that originally were phenocrysts are the only mineral grains that retain a resemblance to the original shape.

The andesitic and basaltic lava flows north of Highway 89A have a relict microcrystalline felty or pilotaxitic texture, and are composed of chlorite, epidote, and albite. The lava west of the Iron King mine has a relict porphyritic texture, and consists of actinolitic hornblende, clinozoisite, and oligoclase; quartz, chlorite, and sericite are accessory minerals.

In some places, flows may have been mistaken for concordant intrusive masses, for the distinction between flows and intrusive masses that have a fine-grained porphyritic or diabasic texture is most difficult. In the absence of other diagnostic features, textures were used to distinguish one from the other. Porphyritic mafic rocks with a phanerocrystalline groundmass, fine-grained equigranular phanerocrystalline mafic rocks, and phanerocrystalline diabases were regarded as intrusive rocks, even though the interior parts of thick flows may have similar textures. Such a mass, which was regarded as intrusive, but which may be a flow, lies immediately to the east of the andesitic and basaltic flows north of Highway 89A (pl. 1). This mass is dominantly fine to medium grained diabase, but local areas of very fine grained diabase suggest a flow.

Microscopic features.—In thin section, the andesitic tuffaceous rock and andesitic breccia consist of chlorite, actinolitic hornblende, albite, and albite-oligoclase, epidote, clinozoisite (probably some zoisite), quartz, and carbonate. Apatite, leucoxene, and magnetite are common accessory minerals. Sericite, other than small flakes disseminated in saussuritized plagioclase, occurs only in some beds in the andesitic tuff, particularly in the upper part of the unit. Actinolitic hornblende is limited to the andesitic breccia and to intercalated andesitic or basaltic lava. The epidote mineral in some specimens is clinozoisite (or zoisite) and in others epidote; clinozoisite (or zoisite) is most abundant. Quartz is distributed erratically in andesitic breccia and the lower port of the andesitic tuffaceous rock. It is abundant only in the upper part of the andesitic tuffaceous rock, where it is commonly associated with sericite.

The mineral assemblage chlorite-albite-epidote (clinozoisite), with or without quartz, is characteristic of the andesitic tuffaceous rock and of the intensely sheared andesitic breccia east of Texas Gulch formation. This assemblage is also common in breccia facies in Spud Mountain and Indian Hills, but there chlorite may be partly to completely replaced by actinolitic hornblende. Chlorite-sericite-albite-quartz-clinozoisite (or epidote) is a common assemblage in some beds in the upper part of the andesitic unit. This assemblage probably results from a mixture of andesitic and rhyolitic detritus in certain beds.

Albite and hornblende in some intercalated mafic lava flows are the only relicts recognized. Some albite crystals are clear except for scattered sericite flakes; in others the content of the epidote group of minerals varies. Only part of the plagioclase crystals originally present survived metamorphism, generally only those whose long dimension lay parallel to the foliation. All the relict plagioclase crystals show some crushing; some are granulated only around the periphery, others can only be surmised from lenticular streaks of granoblastic albite. Relict bedding is obscure in thin section. However, aggregates of the epidote group, with or without minor amounts of albite, chlorite, quartz, and sericite appear to preserve the approximate size and shape of original plagioclase and also the bedding structures.

IRON KING VOLCANICS

Distribution, thickness, and stratigraphic relationship

Iron King volcanics, named here from the exposures in Iron King Gulch, occurs in a northward-trending belt south of Humboldt in the southwest corner of the Jerome area (pl. 1). The formation crops out from the southern boundary of the map area northward for

about 8,000 feet where it is covered by Cenozoic rocks in Lonesome Valley.

The belt of Iron King volcanics ranges in width from 9,500 near Humboldt to 11,500 feet at the south edge of the map area. Because of duplication by the major syncline (pl. 6), the formation is only about 5,500 feet thick, of which all but the upper 500–1,000 feet is basalt; the upper unit is andesitic tuffaceous sedimentary rock. East of the Agua Fria River, however, determination of the direction that beds face from pillow structure revealed at least one intraformational fold. Similar folds may or may not be in the lava on the west flank of the regional syncline, for local reversals in the stratigraphic succession could not be recognized, owing chiefly to pronounced foliation. Hence, a thickness of 5,500 feet for the exposed section is a maximum and the true thickness is probably somewhat less.

Except for the Chaparral volcanics and the Green Gulch volcanics whose stratigraphic positions are unknown, the Iron King volcanics is the youngest formation in the Alder group, according to our interpretations. The Iron King volcanics grades downward into the Spud Mountain volcanics; the contact between the two formations is placed where lava predominates over tuff.

Lithology and internal structure

The Iron King volcanics comprises andesitic tuffaceous sedimentary rock and basaltic and perhaps andesitic flows. Mafic flow is perhaps a more accurate descriptive term, as a distinction in the field between metamorphosed andesite and basalt is not believed possible. The chemical analysis indicates the lava flows are basaltic, and probably basalt predominates, but the possibility of some intercalated andesite cannot be dismissed. Although the areas mapped as flows are predominantly as indicated, tuffaceous beds occur in the transition zone with the underlying Spud Mountain volcanics and as interbeds in the flows. To distinguish these flows from tuffaceous rock, relatively uniform textures and structures over widths of hundreds of feet were found to be the greatest aid. In a few favorable exposures, tops of flows, revealed chiefly by pillow structure and by vesicles and finer grained textures, were recognized contiguous to tuffaceous beds. The bedded tuffaceous rock forming the upper unit of the formation is free of lava flows, at least within the map area.

The basaltic flows range in color from grayish green to greenish black. The color depends, in part, on whether the mafic mineral is chlorite or actinolitic hornblende, and in part, on the degree of foliation. The chloritic and more foliated flows are lighter in color. Megascopic structures consist chiefly of foliation, pillows, and amygdules. Foliation is more apparent in the lava west of the Agua Fria than in those east, owing to chlorite; but whether chlorite connotes more intensive shearing during regional deformation is problematical. Relict pillow structures are obvious, in other places probably present, and in many places, pillow-like structures invite fruitless speculation. Where pillow structures are obvious, it is generally possible to determine the direction that the tops of the flows face. Amygdules are locally abundant, and in well foliated facies, they are attenuated.

Megascopically, most lava consists of irregular grains set in a very fine grained granular green matrix. The megascopic grains comprise dull, light-gray or tan minerals of the epidote group and vitreous albite and quartz. Megascopic relict textures in the flows are rare. In a few places where foliation is weak, the relict saussuritized plagioclase phenocrysts retain the lathlike shape, and in still fewer places, the felty (pilotaxitic) groundmass texture is preserved. Pillow lava facies are very dense and almost completely aphanitic, although very small needles of plagioclase show in some outcrops. Commonly ill-defined patches or areas of mafic rocks with a diabasic or fine-grained equigranular texture crop out sporadically. Whether these rocks are the interior parts of thick flows or small intrusive masses was not generally determined, but in one or two exceptionally good exposures, these rocks appeared to grade into altered lava.

Most of the flows have metamorphic textures and mineral assemblages. A few, however, contain relict plagioclase phenocrysts, fractured and granulated around the periphery; generally only the phenocrysts whose long dimension lay parallel to foliation survived. In one section, relict phenocrysts of plagioclase in a relict pilotaxitic groundmass of plagioclase needles in a chloritic base attest the original character of the rock. Two mineral assemblages characterize the flows, (1) chlorite-albite-epidote, and (2) actinolitic hornblende-soda-oligoclase-epidote (or clinozoisite). Quartz, sericite, pyrite, magnetite, and leucoxene are accessory minerals. Both flows and tuff contain carbonate, but how much of it is metamorphic is questionable, for carbonate in crosscutting veins and irregular pods indicates that some is hydrothermal.

The chemical analysis of typical flow rock is given in table 9. It clearly shows that the rock is a basalt and Daly's average of all basalts is included for comparison. The most significant variations between the two are the higher CaO content and lower Na_2O and K_2O content in the flow from the Iron King volcanics.

The andesitic tuffaceous rock, which appears identical to that in the Spud Mountain volcanics, is pale-yellowish green, dusky-yellow green, and grayish green. The

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TABLE 9.—*Chemical analyses of basaltic flows, in percent*

	1	2
SiO_2	47.6	49.06
Al_2O_3	15.4	15.70
Fe_2O_3	3.5	5.38
FeO	8.6	6.37
MnO	.19	.31
CaO	11.9	8.95
MgO	6.6	6.17
K_2O	.14	1.52
Na_2O	2.0	3.11
TiO_2	1.2	1.36
P_2O_5	.19	.45
Ignition loss	2.2	
H_2O		1.62

1. Rapid analysis of basalt, near Agua Fria River (1,266,600 N.; 403,400 E.). Rapid analysis by S. M. Berthold and E. A. Nygaard.
2. Average basalt, Daly (1933, p. 17).

texture ranges from very fine- to coarse-grained, and alternating zones of fine-, medium- and coarse-grained tuffaceous rock characterize the unit. This alteration is believed to reflect original bedding. Foliation is ubiquitous, but is more pronounced in the finer grained beds.

The megascopic minerals are chlorite, sericite, members of the epidote group, quartz, and carbonate. Abundant sericite and quartz are limited to certain zones. We believe that they indicate admixing of rhyolitic detritus to the original andesitic tuffaceous sediments. The textures and mineral assemblages of the andesitic tuffs are almost completely metamorphic. Two mineral assemblages were recognized: chlorite-albite-epidote (or clinozoisite), and chlorite-sericite-albite-quartz-carbonate.

TEXAS GULCH FORMATION

Distribution

The Texas Gulch formation defined here, composed of alternating belts of rhyolitic tuff and purple slate, crops out in a band about 2,500 feet wide along the western slope of the Black Hills. The formation extends from the southern border of the Jerome area northward for about 14 miles to a point north of the Jerome highway (1,341,000 N.; 419,000 E.) where the formation is covered by Paleozoic rocks (pl. 1). Texas Gulch exposes a representative section (1,290,000 N.; 415,000 E.) and this name has been adopted for the formation. Lausen (1930, p. 176) recognized the distinct lithology of the Texas Gulch formation by referring to quartz-sericite schist and conglomerate (rhyolite tuff) and to the slate. Wilson (1939, p. 1157–1158) described the rhyolitic tuff as arkosic sandstone and squeezed conglomerate and included these rocks and the associated slate in the Alder series.

Thickness and stratigraphic relationship

The thickness and stratigraphic relationship of the Texas Gulch formation are not known. Within the formation, rhyolitic tuff bands alternate with those of purple slate and the explanation of the distribution of the rhyolitic tuff and purple slate within formational boundaries is dependent on the interpretation of structure in local areas. The formation has been deformed so intensely that commonly the abrupt termination of a particular slate or tuff band cannot be determined as a fold, a lens, or a gross boudin (link in boudinage structure). It appears most reasonable from the outcrop pattern that two or more beds of purple slate and of rhyolitic tuff are present. West of the Shylock mine (1,317,000 N.; 417,000–419,000 E.) a maximum of four bands of rhyolitic tuff alternate with purple slate, but whether any of these represent duplication by folds or faults is uncertain. The thickness of individual units, especially the purple slate, changes abruptly along the strike.

Both the east and west boundaries of the outcrop area of the formation are faults. Furthermore, most of the contacts within the formation, such as between purple slate and rhyolitic tuff, are probably mechanical. In our interpretation of the regional structure, however, the Texas Gulch formation is probably older than the Indian Hills volcanics, but whether or not it directly underlies Indian Hills volcanics is not known.

Lithology and internal structure

The rhyolitic tuff is a distinctive rock. It is light, generally gray or white but also commonly cream or light green. Abundant sericite produces a silvery sheen on cleavage surfaces. Quartz grains enclosed in a thin coating of tightly molded sericite produce, a characteristic spherulitic appearance on foliation planes.

Rocks are erratically foliated partly as the result of differences in grain size and probably partly in response to local stress conditions. In some places foliation is pronounced, in others almost imperceptible. Fine-grained tuff generally is smoothly foliated, whereas some pebble beds have a hackly foliation that is apparent from distances where the pebbles cannot be resolved.

Lineation is common. It results from elongation of sericite flakes, or less commonly chlorite, on foliation planes, and from elongation of pebbles.

Rhyolitic tuff ranges from fine-grained tuff through coarse-grained tuff to pebble conglomerates or to cobble conglomerates in local zones south of the road to Cherry. In a few restricted areas fine-grained tuff and intimately associated interbedded gray slate indicate gradation from tuff to slate, but the clean separation into rhyolitic tuff and purple slate generally is most remarkable. The bulk of the rhyolitic tuff is medium grained, and coarse grained tuff is more abundant than fine. Changes in grain size across the strike are com-

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mon, but systematic variations through several beds are rare, resulting in general uncertainty on local stratigraphic sequences.

Abundant quartz and sericite and locally minor amounts of feldspar are the chief megascopic constituents of the tuff. However, appreciable amounts of feldspar and sufficient chlorite to color the tuff a light green occurs in the westernmost exposures between 1,300,000 N. and 1,320,000 N. In some beds, pebbles are massed almost to the exclusion of matrix, elsewhere beds consist of a few scattered pebbles surrounded by rhyolitic tuff, generally coarse grained. Conglomerate beds vary in abundance, but four or five in a section of rhyolitic tuff 500 feet wide are not uncommon. Pebbles in conglomeratic beds, associated with well-foliated rhyolitic tuff, are commonly stretched so that lengths are several times widths; in other beds stretching was not evident.

Scattered throughout the rhyolitic tuff are beds that show few or no fragments on surfaces perpendicular to foliation but have lenticular and linear streaks and patches rich in sericite on the foliation planes. Origin of these streaks and patches is uncertain; possibly they represent completely sheared fragments of rhyolitic pumice.

In thin section, quartz, albite, microcline, and sericite are the common minerals, and opaque minerals, epidote, and apatite occur in small amounts. Jasper and rhyolite are the dominant lithic fragments, but fragments composed of quartz, microcline, and albite and of microcline-quartz graphic intergrowth also are present.

Lithic fragments and the larger grains of relict quartz, microcline, and albite are set in a very fine grained (about 0.03 mm. maximum diameter) matrix composed of disseminated sericite flakes and an aggregate of quartz and alkalic feldspar. Quartz is the most abundant mineral, forming as much as 50 percent of the rock. It occurs both in granoblastic mosaics and in relict, angular to subangular clastic grains, some of which are fractured, granulated around the margins, and strained. Sericite is abundant as small flakes and patches; "ribbons" of sericite anastomose around quartz and feldspar grains forming the chief microscopic indication of the foliation. Albite and microcline occur in relict clastic grains, and alkalic feldspar, subordinate to quartz, occurs in small granoblastic grains in the matrix.

Rhyolitic fragments and sharp embayments on quartz grains testify to the volcanic origin of most of the rhyolitic tuff. However, the fragments of jasper, quartz-microcline in graphic intergrowth, and of somewhat equigranular quartz-microcline-albite rock were probably derived from nonvolcanic sources. The remarkably clean separation of purple slate and rhyolitic tuff indicates relatively rapid deposition of the tuff, otherwise contamination by purple slate would have resulted.

Purple slate is chiefly a purple to deep maroon rock, but some local areas are gray, and others grade along the strike into green slate for a short distance. Several bands of green slate, one locally 300 feet wide, are intercalated with purple slate in the more westerly exposures in the southwestern part of the Jerome area. Purple slate has a very pronounced cleavage that in places is cut by cross slips or shears that have no recognizable systematic pattern. Post cleavage deformation produced local masses and individual occurrences of crumpled cleavage; some of the axes of the small folds thus formed plunge northward, others southward.

Variations in lithology and texture are manifest chiefly by interbedded marble beds and by local variations of granularity. The latter, recognized only in isolated outcrops, is the result of sandy layers intercalated with the slates. The marble beds, or perhaps layers would be more precise terminology, as no assurance exists that the present layers have similar widths or similar locations relative to adjacent rocks as when originally formed, range from a few inches to about 20 feet in outcrop width, and from a few feet to about 3,000 feet in length.

Relict bedding occurs scattered sparsely throughout the purple slate. Generally it is recognized with assurance only in outcrops where bedding and cleavage are not parallel, for here bedding shows on cleavage planes as parallel bands slightly different in color or, less commonly, in grain size.

Microscopically, typical well-foliated purple slate consists of abundant oriented sericite, detrital quartz, and hematitic dust. Apatite is a common accessory mineral. The quartz grains are about 0.05 millimeter in maximum dimension, individual sericite flakes are about 0.03 millimeter long, and aggregates or clusters of sericite are about 0.06 millimeter long. Locally the purple slate near 1,295,000 N.; 418,000 E. and in the area west of the Yaeger mine contains megascopic quartz-carbonate alteration; the foliation in the altered slates is obscure, and the rock has a dull aspect in comparison with typical slates. Thin sections of these altered slates reveal that sericite is partly to completely absent and in its place very fine grained (about 0.08 mm. long) fibrous flakes of an unknown mineral occur. The birefringence of this mineral is low and indices of refraction probably lie between 1.560 and 1.580. In one section, the hematitic dust is in patches composed of abundant closely spaced discrete grains.

Chemical analysis of the purple slate (table 10) reveals that the slate was probably a normal terrigenous

TABLE 10.—*Chemical analyses of slate and shale, in percent*

	1	2	3
SiO_2	62. 2	60. 96	60. 15
Al_2O_3	19. 4	16. 15	16. 45
Fe_2O_3	7. 5	5. 16	4. 04
FeO	. 30	2. 54	2. 90
MnO	. 02	. 07	Tr.
MgO	1. 6	3. 06	2. 32
CaO	. 08	. 71	1. 41
K_2O	3. 6	5. 01	3. 60
Na_2O	1. 3	1. 50	1. 01
TiO_2	. 75	. 86	. 76
P_2O_5	. 20	. 23	. 15
Ignition	3. 3	-------	-------
H_2O-	-------	. 17	. 89
H_2O+	-------	3. 08	3. 82
BaO	-------	. 04	. 04
SO_3	-------	-------	. 58
NH_3	-------	. 01	-------
CO_2	-------	. 68	1. 46
C	-------	-------	. 88
Total	-------	100. 23	100. 46

1. Rapid analysis of purple slate, south of Shylock mine (1,312,900 N.; 418,500 E.). Analysts: S. M. Berthold and E. A. Nygaard.
2. Purple slate, Castleton, Vt. From Clarke, F. W. (1924, p. 554).
3. Composite analysis of 51 Paleozoic shales, by H. N. Stokes. From Clarke, F. W. (1924, p. 552).

sedimentary rock rather than a fine-grained rhyolitic tuff, as might be expected here because of the dominance of volcanic rocks. The significant data from the analysis are the high Al_2O_3 content and the high ratio of K_2O to Na_2O. All analyses of rhyolite from the Jerome area show a much lower Al_2O_3 content and a high ratio of Na_2O to K_2O. The admixing of some rhyolitic detritus, however, cannot be dismissed. The analysis of purple slate from Vermont and the composite analyses of 51 samples of shale are included for comparison (table 10). Both show a K_2O to Na_2O ratio somewhat similar to the purple slate from the Jerome area, and both have a higher Al_2O_3 content than the rhyolite from the Jerome area, although not as high as the purple slate.

CHAPARRAL VOLCANICS

Distribution

The Chaparral volcanics, named here from the good exposures in Chaparral Gulch, crops out in a northeastward-trending belt in the southwest corner of the area shown on plate 1. It extends beyond the southern limit of the map area, and its northern limit is hidden beneath the Cenozoic rocks in Lonesome Valley, somewhere between Dewey and the Indian Hills. The formation crops out over a strike length of 4½ miles and over a width ranging from 1,500 to 3,500 feet. The formation does not occur elsewhere in the Jerome area.

The Chaparral volcanics is divided into two lithologic units: rhyolitic tuffaceous sedimentary rock and andesitic tuffaceous sedimentary rock. At the southwest corner of the map area, rhyolitic tuffaceous rock crops out in three zones, but one of these is less than 50 feet wide and only about 1,500 feet long. The other two, which are believed to be opposite limbs of a northward-plunging anticline, trend northeastward, and merge near the boundary of the Prescott and Mount Union quadrangles. Here quartz diorite splits the rhyolitic tuffaceous rock into two zones. The western, which ranges in width from 125 to 850 feet, crops out continuously from this point northeastward for about 3 miles to the overlap of Cenozoic rocks. The eastern zone of rhyolitic tuffaceous rock, about 250–600 feet wide, trends northeasterly from the quadrangle boundaries for about a mile, where it divides into two zones, which may be due to a reversal of plunge of the anticline. The outcrops of both zones trend northeastward, and extend for nearly 2 miles to the Cenozoic rocks. In addition, there is a separate narrow zone of rhyolitic tuffaceous rock east of the other rhyolitic zones near the quadrangle boundary. It crops out northeast of the boundary for about 4,000 feet and southwest about 5,500 feet.

Distribution of the andesitic and rhyolitic tuffaceous rocks is virtually the same, except that most of the andesitic rocks lie east of the rhyolitic. The andesitic tuffaceous rock is more abundant in the Mount Union quadrangle, but rhyolitic rock preponderates in the Prescott quadrangle.

Thickness and stratigraphic relationship

The thickness of the Chaparral volcanics is not known, nor can it be approximated. Strong phyllonitization (Knopf and Ingerson, 1938, p. 190), isoclinal folds, and faults that removed both the top and bottom of the formation, vitiates any approximation of the original thickness.

Because the Chaparral volcanics lies between two major faults, the Chaparral fault on the northwest and the Spud fault on the southeast, its stratigraphic relationship is indeterminate. This stratigraphic isolation is one of two reasons that compel the designation of this unit as a separate formation. The second is the absence of lithology diagnostic of any other formation. The proximity of the Spud Mountain volcanics and the similarity between the rhyolitic and andesitic tuffaceous rocks in the two formations intice correlation, but alone, none of these lithologies are diagnostic of the Spud Mountain volcanics. To be Spud Mountain volcanics without doubt, a rock unit must contain either the andesitic breccia or coarse-grained crystal tuff, containing the large equant plagioclase crystals such as those in the andesite fragments that characterize the breccia. Lacking either of these two diagnostic rock types, the Chaparral volcanics cannot be correlated with the Spud Mountain volcanics, and it is given tentative formational rank.

If the interpretation of the structure of the Chaparral volcanics is correct, andesitic tuffaceous rock overlies and underlies the rhyolitic rock.

Lithology and internal structure

The Chaparral volcanics comprises interbedded rhyolitic and andesitic tuffaceous sedimentary rocks. During deposition, there was limited mixing of rhyolitic and andesitic detritus; sporadic rhyolitic tuffaceous beds too small to map are intercalated in the andesitic rock, and conversely. Minor amounts of chlorite in the rhyolitic rock and of sericite in the andesitic rock indicate mixing.

The andesitic tuffaceous rock is somewhat variable in texture, structure, and mineral composition. The most abundant type is a fine-grained, well-foliated, fissile rock that ranges in color from green or greenish gray to green mottled with irregular patches and streaks of grayish-yellow green or pale-yellowish green. The color changes reflect variation in relative amounts of the component minerals. Intercalated in these finer grained andesitic tuffaceous rocks in subordinate amounts (but more abundantly to the east than to the west), are beds from 1 to 50 feet thick of medium- to coarse-grained andesitic tuff, characterized by abundant saussuritized plagioclase crystals. These rocks are grayish green, owing to much chlorite. Although these coarse-grained rocks do possess a foliation, it is much less pronounced and more irregular than in the finer grained rocks.

In a few isolated outcrops and in one bed traced for 3,000 feet, greatly elongated lithic fragments attest to the accumulation of at least minor amounts of coarse material in the Chaparral volcanics. In one exposure, the fragments formed a lineation that plunges northeastward at 60°. Intense shearing, accompanied by recrystallization, has generally obliterated the original lithologic composition of the fragments, but in one locality the fragments appear to be porphyritic andesite.

The fine-grained, fissile andesitic tuffaceous rock is characterized by the assemblage, chlorite-sericite-albite-epidote-quartz, and the coarser grained, more massive andesitic tuff by the assemblage, chlorite-albite-epidote. The textures are dominantly metamorphic, and are typical of foliated, low-grade metamorphic rocks.

The rhyolitic tuffaceous rock is a fine-grained finely laminated, fissile phyllite that ranges in lithologic composition from a relatively pure rhyolitic rock to a mixed rhyolitic and andesitic rock. It is yellowish gray to very light gray, and locally, where sufficiently chloritic, has a greenish cast. Quartz, orthoclase, microcline, and albite are the only truly megascopic minerals, but the luster or sheen on cleavage flakes obviously indicates sericite. In thin section, the micro-granular matrix is sericite, quartz, some potash feldspar, and a little albite.

Some of the more massive beds range in color from pale red to yellowish gray, and contain megascopic quartz and feldspar. The matrix is mainly quartz and orthoclase.

The mixed rhyolitic-andesitic rocks, many of which are finely laminated, are greenish gray to dark gray and contain variable amounts of epidote, actinolite, and biotite, or chlorite, together with quartz, potash feldspar, albite, and sericite. The three types are gradational, even on a microscopic scale.

GREEN GULCH VOLCANICS

Distribution

The Green Gulch volcanics, named here for good exposures along Green Gulch in the southeastern part of the Prescott quadrangle, forms a triangular-shaped mass in the extreme southwest corner of the Jerome area (pl. 1). This volcanic formation is bounded on the east by the Chaparral fault; on the west, it extends beyond the map area; and to the north is covered by Cenozoic rocks of Lonesome Valley. On the regional geologic map of the Jerome area (pl. 1), the Green Gulch volcanics is not subdivided, although work currently in progress by M. H. Krieger for the Survey will result in at least a twofold lithologic subdivision that consists of a basaltic or andesitic flow unit and of a tuffaceous unit of variable composition. The flow unit lies east of the tuff.

Thickness and stratigraphic relationship

The thickness of the Green Gulch volcanics is not known because neither the top nor the bottom of the formation is exposed. The outcrop width is about 10,500 feet, but the attitude revealed by the direction that beds face indicates at least several folds of uncertain magnitude so that the outcrop width is greater than the stratigraphic thickness. Small intrusive dikes and lenses of gabbro, diabase, and alaskite further increase the outcrop widths throughout the stratigraphic thickness; in the basaltic unit, however, the proportion of intrusive material is believed small.

The stratigraphic relationship of the formation could not be determined as the formation is bounded on the southeast by the Chaparral fault and on the west by intrusive rocks, chiefly gabbro and alaskite. Locally, breccia, crystal tuff, and rhyolitic tuffaceous beds in the Green Gulch volcanics somewhat resemble the Spud Mountain volcanics. The basaltic flows, however, more closely resemble those in the Iron King volcanics, which overlie the Spud Mountain volcanics. Despite the local

similarities, the lithology of the Green Gulch volcanics is not sufficiently like that of any other formation to permit a correlation. Until such time, therefore, that the Green Gulch volcanics is found in depositional contact with other units, its stratigraphic position will remain unknown and the unit should be given formational rank.

Lithology and internal structure

The Green Gulch volcanics comprises basaltic and andesitic flows and sedimentary tuffaceous rocks of variable composition. Most of the flows are medium to dark gray with a greenish cast. The exposures in the northeastern outcrops appear somewhat darker than those in the southwest, possibly owing to more abundant metamorphic hornblende. Chloritic rocks, which generally have a pronounced foliation, are grayish green. Locally in the flows, relict pillow and breccia structures, amygdules, vesicles, and bedding are preserved, and are well exposed in the larger gulches; pillow structures were only recognized in Green Gulch and the next gulch to the south (pl. 1, indicated by symbol). Vesicles and amygdules of quartz and quartz-epidote are more abundant and widespread than the pillow lavas. Locally, calcite-filled vesiclelike structures occur, but whether these are true vesicles or the result of metamorphism was not determined. Relict bedding and drag folds are preserved in tuffaceous interbeds separating the flows and breccia.

Fragments of rhyolite and andesite or basalt form the breccia structure. They are stretched like the pillow lavas; lengths of some fragments exceed widths by as much as 10 times.

Small saussuritized plagioclase phenocrysts and porphyroblasts or aggregates of hornblende are the most abundant megascopic minerals in flows. Locally chlorite also can be recognized. Relict felty (pilotaxitic) texture was recognized in thin section from one flow, but metamorphism has obliterated most original textures, and transformed the flows into schist.

Two mineral assemblages were recognized from the flows: actinolitic hornblende-albite-epidote, and hornblende-oligoclase-epidote. Brown biotite, sphene, and apatite are associated with this latter assemblage, and chlorite and carbonate with the former, where the rocks have been more highly sheared.

No analyses of the flows have been made, but the predominance of mafic constituents and the lack of pronounced porphyritic textures in the outcrops where relict textures are preserved, lead us to believe that the flows are basaltic in composition rather than andesitic.

The tuffaceous rocks range in composition from rhyolitic to basaltic. The most characteristic rock ranges in color, from greenish black to greenish gray or dark to light olive gray and has a characteristic sheen on cleavage surfaces due to sericite and variable amounts of admixed chlorite. Fine-grained, dark-gray siliceous rocks that weather to a light pinkish-gray, and lesser amounts of light-gray porcelaneous slate are also present.

The tuffaceous rocks range from finely laminated phyllite to more massive and locally coarser grained rocks. A few lithic pebbles, mostly of rhyolite, are locally present. Relict bedding and drag folds are apparent in water worn outcrops along Green Gulch, but not in weathered outcrops on the adjacent ridges. Small phenocrysts of feldspar and (or) quartz are sparsely to abundantly distributed throughout the rock. Thin sections of the tuffaceous rock in Green Gulch reveal small grains of saussuritized plagioclase and locally quartz and albite in a matrix composed of quartz and variable amounts of sericite, chlorite, brownish biotite, magnetite, minerals of the epidote group, and sphene. Porcelaneous slate and rhyolitic tuffaceous rocks contain a few scattered fragments of albite and quartz crystals in a matrix composed of quartz and alkalic feldspar including minor amounts of sericite, epidote, and locally biotite. Magnetite is generally present and locally it is abundant.

ORIGIN OF YAVAPAI SERIES

The rocks of the Yavapai series are in large part normal products of volcanic eruptions, such as lava flows, flow breccia, and agglomerates. Pillow lavas are present in nearly all the basaltic-andesitic flow units. A submarine origin for pillow lavas is commonly assumed because of the association of many Paleozoic and younger pillow lavas with marine sedimentary rocks, particularly radiolarian chert. However, Fuller (1931) has described excellent pillow structures from the Tertiary basalts of the Columbia Plateau that formed by outpouring of lava into lakes. Tyrrell (1929, p. 38) has suggested that extrusion into soft water-soaked sediments, beneath ice sheets, or into rain-soaked air, may induce sufficiently rapid cooling to produce the pillow structure. No evidence of glacial conditions was found in any of the formations in the Yavapai series, and this possibility cannot be considered seriously in the absence of any supporting evidence. However, it cannot be denied that subaerial eruptions of the pillow lavas are a possibility.

The agglomeratic deposits apparently represent volcanic ejecta erupted from nearby sources. The lack of bedding in most of these deposits might be used as evidence of subaerial eruptions, for if these ejectamenta were deposited subaqueously, current action might redistribute them into stratified beds, unless they were quickly buried by succeeding lava flows.

Volcanic breccia beds can form as a product of volcanic eruptions or by transportation of volcanic debris by water or mudflows (Anderson, 1933, p. 246). The association of many of the breccia beds with interbeds of sandy tuffaceous rock indicates that the breccia largely results from water transportation of coarse volcanic debris. The rocks do not contain evidence which indicates the environment of deposition. It cannot be denied that some of the breccia may have formed by volcanic eruptions that broke up older lava flows and redistributed the fragments near volcanic vents.

The bedded tuffaceous rocks so common in the Alder and Ash Creek groups may represent subaerial reworking of debris derived from the flow rocks and deposited along flood plains or in lakes. Conversely it must be admitted that marine fossils are found in similar bedded tuffaceous rocks of Paleozoic or younger age (Heyl, 1948, p. 17).

The origin of bedded chert is an unsolved problem; some believe that the process is diagenetic, and that silica derived from siliceous organisms (Bramlette, 1946) or from volcanic ash (Rubey, 1929), is added to shale. The others suggest that the bedded cherts are formed by direct precipitation of silica (Davis, 1918) of magmatic source related to contemporaneous volcanism (Taliaferro, 1933). Because many bedded cherts are radiolarian-bearing or associated with marine diatoms, the assumption is made that the chert beds are of marine origin, but as marine fossils are absent from the Precambrian bedded chert of the Grapevine Gulch formation, it may be equally possible that these chert beds are of lacustrine origin. The origin of iron formation, now represented in the Grapevine Gulch formation by the jasper-magnetite beds, presents the same unsolved problem.

Only the Texas Gulch formation in the Alder group contains sedimentary rock that in part is clearly terrigenous, as shown by the nonvolcanic character of some pebbles and cobbles in the interbedded conglomerate. The purple slate appears to represent a normal terrigenous shale, and the intercalated limestone probably represents chemical precipitation in quiet water. Even though terrigenous material was added with rhyolitic debris to the basin where the Texas Gulch formation accumulated, this evidence does not indicate anything relating to the environment of deposition. Minor amounts of similar slate, limestone, and interbedded conglomerate in the Spud Mountain volcanics indicate intermittent supply of terrigenous material during the time volcanic material was the main type of sediment.

In summary, the rocks of the Yavapai series do not contain diagnostic evidence on their environment of deposition. They may represent nonmarine accumulation of lava flows, pyroclastic ejecta, reworked volcanic debris, and terrigenous sediment deposited on broad flood plains or in lakes or both. Some of the rocks may be nonmarine, and some may be marine. And it is equally possible that the entire sequence is marine, as the pillow lavas, bedded chert, and jasper-magnetite beds possibly suggest.

INTRUSIVE ROCKS

INTRUSIVE RHYOLITE

Distribution and relation to other rocks

The intrusive rhyolite intrudes only the eastern mass of the andesitic breccia unit of the Spud Mountain volcanics south of 1,285,000 N. It forms five separate masses, but other masses may occur in the same general area. Hydrothermal alteration of the andesitic breccia adjacent to the quartz diorite locally resulted in irregular-shaped masses that megascopically appear so similar to the intrusive rhyolite that we could not everywhere distinguish between the two.

The irregular outlines of the intrusive rhyolite—particularly of the largest mass—implies an intrusive origin. The structure of the andesitic breccia near the largest mass of intrusive rhyolite is essentially homoclinal, so that the many divergent prongs of rhyolite forming the mass must have originated through intrusion. The age of the intrusive rhyolite relative in the other intrusive rocks is unknown.

General characters

The intrusive rhyolite is light, commonly cream or white. It consists of a few phenocrysts of quartz and plagioclase set in a felsitic groundmass. Secondary veinlets and disseminated grains of pyrite—both oxidized—mark most of the outcrops with streaks and patches of brick-red iron stain. Commonly the intrusive rhyolite has a poor cleavage that cannot be related to the orientation of any minerals.

Thin sections reveal relict phenocrysts of quartz and euhedral albite in a microcrystalline groundmass. A little chlorite in irregular masses and sericite, liberally sprinkled throughout, are the only other constituents. The albite has remarkably perfect crystal outlines, in complete crystals and in parts of crystals. Several crystals of albite have cores of groundmass material. The quartz phenocrysts have irregular shapes, and one has smooth indentations, characteristic of resorption. The groundmass consists of a microcrystalline aggregate whose undulatory extinction suggests divitrified glass. The sericite appears to replace chlorite, penetrate quartz, and rim the albite phenocrysts. Sericite is probably secondary and related to the pyritization.

QUARTZ PORPHYRY

Distribution

Most of the quartz porphyry intrusive bodies crop out in the belt of Precambrian rocks on the east side of Mingus Mountain, extending northward from Oak Wash to Jerome. Three large and several small bodies are in the Ash Creek drainage area, south of Mingus Mountain. East of Humboldt, a large body of quartz porphyry appears east of the Shylock fault, and several small bodies lie to the west, and one small mass is exposed south of Humboldt. Eastward- and northeastward-trending dikes of quartz porphyry are in the drainage area of Black Canyon and east of Mingus Mountain, south of Oak Wash.

Because of excellent exposures on Cleopatra Hill near Jerome, the term "Cleopatra quartz porphyry" (Fearing, 1926, p. 758; Reber, 1938, p. 61), has been used for this rock, although Lindgren (1926, p. 57) referred to the rock as rhyolite porphyry. In this report the term "quartz porphyry" will be used.

At the south margin of the exposures of Texas Gulch formation (pl. 1, 1,265,000 N.; 415,000 E.), a mass of feldspar porphyry of unknown size has been included with the quartz porphyry. Not enough information is available to justify this assignment, but it was deemed inadvisable to add another cartographic unit for a rock of such limited extent.

Relation to other rocks

The quartz porphyry intrudes both the Alder and Ash Creek groups. South of Humboldt, the quartz porphyry intrudes andesitic tuffaceous rock of the Iron King volcanics and, to the southeast of Humboldt, andesitic breccia of the Spud Mountain volcanics. South of Mingus Mountain, the quartz porphyry intrudes the Grapevine Gulch formation, Deception rhyolite, Brindle Pup andesite, and the dacite of Burnt Canyon. East of Mingus Mountain, the porphyry intrudes the Buzzard and Deception rhyolite, and Shea basalt, and at the United Verde mine, the Grapevine Gulch formation. The eastward-trending dikes of porphyry intrude the Gaddes basalt, Buzzard and Deception rhyolite, and Shea basalt.

South of Mingus Mountain, the gabbro intrudes the quartz porphyry. This relationship is brought out clearly at 1,319,500 N., 425,500 E.; 1,307,000 N., 437,000 E.; and 1,308,000 N., 433,500 E., where dikes of gabbro cut through the quartz porphyry (pl. 1). This relationship is important to the local history at the United Verde mine, for no dikes of gabbro have been found to cut the adjacent quartz porphyry in the mine and a screen of tuffaceous sedimentary rocks of the Grapevine Gulch formation partly separates the two intrusive rocks. However, the quartz porphyry is highly foliated (N. 20° W.) in and near the United Verde mine and the gabbro is massive, except at its contact where locally a narrow foliated zone is parallel to the margin (N. 30° E.). On the west side of the gabbro mass (pl. 5, 350 N.; 1,150 W.) highly foliated quartz porphyry is present, indicating that the regional foliation (N. 20° W.) was developed in the quartz porphyry before the intrusion of the gabbro and proving that here also the quartz porphyry is older than the gabbro.

South of Black Canyon, one of the quartz porphyry dikes is cut off by the quartz diorite, proving that the widespread quartz diorite is younger than the quartz porphyry.

Possibly two periods of quartz porphyry intrusions are represented in the Mingus Mountain quadrangle; the eastward-trending dikes may be younger than the larger intrusive masses of porphyry. Evidence for this possibility is present in the United Verde mine pit and underground where the fine-grained tuffaceous and cherty beds of the Grapevine Gulch formation are folded into small northward plunging anticlines and synclines from 2 to 20 feet across. Sills of quartz porphyry, from 1 to 2 inches thick, are concordant to folded beds, and offshoots locally cut across the folds. In one exposure underground in the mine, a fold 1 foot wide is clearly cut off by quartz porphyry, indicating that the quartz porphyry was intruded after the formation of that particular fold.

In the United Verde pit, excellent exposures reveal that the foliation in the folded Grapevine Gulch formation parallels the axial planes of the folds (axial-plane foliation) and the equally well-defined foliation in the adjacent quartz porphyry is parallel to the same plane. This geometric relationship indicates that the foliation in both the sedimentary rocks and quartz porphyry formed during the folding. The cross-cutting relationship of dikes to minor folds indicates therefore that the intrusion took place at the time of folding, and while the axial plane foliation was forming.

In Mescal Gulch (1,358,000 N.; 440,000 E.), however, an eastward-trending dike of quartz porphyry clearly cuts across folds several hundred feet wide in the Deception rhyolite. If the larger mass of porphyry was intruded during the period of folding, the dike is clearly younger than folding. Because none of the eastward-trending dikes intersect any of the gabbro masses, their age relationship is unknown.

The large masses of quartz porphyry are essentially concordant with the structure of the host rock, and where the host rocks are generally homoclinal, elongate sill-like bodies of quartz porphyry occur, such as at Jerome where the quartz porphyry body is 2,000 to 3,000 feet wide and more than 2 miles long, conformable

with the adjacent Grapevine Gulch formation and Deception rhyolite. North of Oak Wash, two bodies of quartz porphyry trend northeastward, essentially conformable to the Shea basalt and Buzzard rhyolite. Where the host rocks are folded, the quartz porphyry masses generally conform to the folds, such as south of Mescal Gulch where the west-northwestward-plunging fold in the Shea basalt is reflected in the western contact of the quartz porphyry. In detail, however, the quartz porphyry masses crosscut all the flows and sedimentary beds along their contacts.

South of Black Canyon, two masses of quartz porphyry are discordant to the host rock. One mass cuts across the Deception rhyolite (1,324,000 N.; 439,000 E.) and the other cuts off the Brindle Pup andesite (1,320,000 N.; 447,000 E.).

General characters

The quartz porphyry is characterized by abundant quartz phenocrysts averaging about 2 millimeters in diameter. Near the United Verde mine, the quartz porphyry is strongly foliated, and the quartz phenocrysts form "eyes" in a sericitic matrix. Elsewhere, most of the quartz porphyry contains plagioclase phenocrysts, from 2 to 4 millimeters long; in local areas, the feldspar crystals are 6 millimeters long. Some of the masses of quartz porphyry contain local facies in which quartz phenocrysts are absent, but quartz is abundant in the groundmass. The groundmass is finely crystalline and minute quartz and feldspar grains are recognizable with a hand lens; locally the groundmass is microcrystalline and no minerals can be recognized except under a microscope. Most of the porphyry has a light to dark-gray groundmass, but local facies are green or purple.

Microscopic studies add little to the descriptions given by Rice (1920, p. 62); the quartz phenocrysts are embayed and strained and some are shattered and recrystallized. The plagioclase phenocrysts are albitic in large part, locally containing sericite, epidote, and calcite. In a few specimens, oligoclase remnants were recognized, veined and partly replaced by albite. No mafic phenocrysts were observed, but scarce aggregates of chlorite or hornblende may represent original mafic minerals. The groundmass consists of a microgranular aggregate of quartz and alkalic feldspar, and some polysynthetic twinning indicates albite in part. The individual crystals range from 0.04 to 0.1 millimeter in size. Some of the alkalic feldspar forms minute spherulites. Green biotite, hornblende, epidote, calcite, magnetite, and chlorite are common in crosscutting veinlets or filling spaces between the groundmass quartz and feldspar. Chlorite is abundant in many specimens from the northern exposures.

Near the United Verde mine, particularly where the quartz porphyry has been foliated, feldspar phenocrysts are not visible. Some thin sections show mats of sericite that may represent original plagioclase phenocrysts, but in many only the original quartz phenocrysts remain and the groundmass is an aggregate of microgranular quartz, sericite and chlorite streaks, and scattered carbonate rhombs and streaks. Presumably some hydrothermal alteration has occurred, particularly in the formation of the chlorite-rich facies of foliated quartz porphyry. Mineralogically, the quartz porphyry near the United Verde mine is similar to the hydrothermally altered Deception rhyolite, except that the quartz phenocrysts are larger in the quartz porphyry.

The quartz porphyry mass cutting the Buzzard rhyolite north of Oak Wash appears more granular in its southern exposures, and the rock is pale pink. Under the microscope, however, the texture is revealed as seriate, and the crystals range in diameter from 0.08 to 2 millimeters. Quartz and alkalic feldspar are in micrographic intergrowths; some of the host feldspar is albite and some is orthoclase or twinned microcline. Sericite is present along fractures. Northward, the seriate facies grades into normal quartz porphyry.

TABLE 11.—*Chemical analyses of quartz porphyry, in percent*

	1	2	3	4	5	6	7
SiO_2	74.6	78.29	77.00	74.60	72.80	72.80	72.36
Al_2O_3	12.5	11.30	10.11	10.54	14.90	14.57	14.17
Fe_2O_3	1.6	------	------	1.71	------	------	1.55
FeO	2.0	[1] 2.98	[1] 1.78	1.87	[1] 2.84	[1] 2.86	1.01
MnO	.10	------	------	------	------	------	.09
MgO	.66	.87	.20	.58	1.80	1.48	.52
CaO	1.2	.13	1.42	2.09	.90	1.98	1.38
K_2O	2.6	.73	1.08	1.90	2.38	------	4.56
Na_2O	3.6	5.47	3.61	2.08	1.26	------	2.85
TiO_2	.26	.20	------	------	------	------	.33
P_2O_5	.06	------	------	------	------	------	.09
H_2O+	------	------	.04	.07	------	------	} 1.09
H_2O-	------	.43	.58	.12	------	------	
Ignition loss	1.2	------	------	------	2.22	2.67	------
S	------	------	1.60	1.03	.50	.06	------
CO_2	------	.34	.85	3.35	------	1.35	------
Cu	------	.01	------	------	.04	.02	------
Zn	------	------	------	------	------	.04	------
	------	100.75	------	------	100.79	------	------

[1] Total iron reported as FeO.

1. Quartz porphyry north of Ash Creek (1,325,000 N.; 438,000 E.). Rapid analysis by S. M. Berthold and E. A. Nygaard.
2. Quartz porphyry west of Copper Chief mine, approximate location 1,348,000 N.; 447,500 E.
3. Brown quartz porphyry, Hull Canyon, approximate location 1,360,000 N.; 437,000 E.
4. Gray quartz porphyry, south side of Hull Canyon near Deception rhyolite contact. Approximate location 1,360,500 N.; 439,000 N.
5. Typical quartz porphyry, handpicked at United Verde smelter.
6. Quartz porphyry, 2,100 level, United Verde mine. Analyses 2, 3, 4, 5, and 6 furnished through the courtesy of Phelps Dodge Corp.
7. Average quartz porphyry (Daly, 1933, p. 9).

The only available chemical analysis of quartz porphyry not collected from areas of hydrothermal alteration is of analysis 1, table 11; it compares closely to the average composition of quartz porphyry (analysis 7), except for a higher Na_2O and lower K_2O content. The remaining analyses are of quartz porphyry that

has undoubtedly been hydrothermally altered. The most southerly of these, analysis 2, is richer in Na_2O and poorer in K_2O than the quartz porphyry in and near the United Verde mine where sericite is an important constituent. The CaO content is very low in analysis 2, indicating that albitization must have taken place; this is confirmed in part by microscopic studies of quartz porphyry in the region around the Copper Chief mine. The higher MgO content in analyses 5 and 6 is related to the introduction of chlorite in the rock, a common feature near the United Verde mine. The high CO_2 content of analyses 4 and 6 indicates the presence of calcite or dolomite. The classification of the quartz porphyry is difficult because of the alteration of much of the porphyry in and near Jerome, and Reber's statement that the "usual composition is that of a normal rhyolite, often as albite rhyolite and rarely a dacite" (Reber, 1922, p. 13) is an excellent summary.

A breccia facies of the quartz porphyry is exposed in Mescal Gulch, and it has been distinguished on plate 1. The angular fragments in the breccia range in size from 1 inch to 1 foot and consist of tuffaceous sedimentary rock from the Grapevine Gulch formation and quartz porphyry and jasper. The matrix consists of quartz porphyry locally flow banded. Dike offshoots of quartz porphyry cut adjacent less-brecciated rock. Screens of nonbrecciated tuffaceous sedimentary rock between masses of quartz porphyry breccia can be traced for several hundred feet.

West of the main breccia mass, the quartz porphyry is generally massive but swirls of flow-banded porphyry crop out locally. The west margin of the quartz porphyry body in Mescal Gulch is less intensely brecciated, and the fragments are dominantly of quartz porphyry in a quartz porphyry matrix. The western breccia facies grades into the massive facies.

Similar but smaller breccia masses of quartz porphyry are exposed in Hull Canyon along the west margin of the quartz porphyry. Here, the fragments are composed largely of quartz porphyry in a quartz porphyry matrix and the fragments average from 3 to 4 inches in length. In some outcrops the fragments are oriented, the long axes strike north and plunge 80° E., whereas in other outcrops, the fragments are not oriented. An inclusion of tuffaceous sedimentary rock, 2 feet wide and 10 feet long, is in the breccia and is injected locally by breccia dikes. The matrix of some of the breccia is reddish purple as are some of the tuff beds. Thin sections reveal that the pigment is finely divided hematite.

Breccia with foreign as well as quartz porphyry fragments occurs in the quartz porphyry mass near the Copper Chief mine, east of Mingus Mountain. Along the margins of some of the masses of quartz porphyry in the Ash Creek drainage area, chilled border facies are locally broken and cemented by quartz porphyry.

Presumably the breccia facies was formed by the shattering of the wall rocks and early consolidated border zone of the quartz porphyry by later intrusions of quartz porphyry. If this is the proper explanation, probably there were several periods of intrusions and some quartz porphyry masses may be products of multiple intrusion.

GABBRO

Distribution

Gabbro, diorite, diorite porphyry, pyroxenite, and diabase are present as intrusive bodies in most exposures of the Yavapai series. Most of these mafic rocks are probably gabbro, and all have been indicated as gabbro on the geologic map (pl. 1). In terms of ore control, the most important mass of gabbro is exposed at Jerome, cropping out west of the United Verde mine. Pyroxenite is exposed only north of Blowout Wash on the east side of Mingus Mountain (1,355,000 N.; 448,500 E.). The masses of gabbro on the east side of Mingus Mountain are small, but south of Mingus Mountain, large bodies of gabbro occur. Similar gabbro crops out in Lonesome Valley in the northwest quadrant of the Mingus Mountain quadrangle, and smaller bodies are exposed east and west of Dewey in the southwestern corner of the mapped area (pl. 1).

The term "United Verde diorite" was given by Reber (1922, p. 14) to the gabbro exposed at Jerome, but in this report, the term "gabbro" will be used to include all the gabbroic and related rocks exposed in the Jerome area.

Relation to other rocks

The gabbroic rocks intrude most of the formations of the Alder and Ash Creek groups, proving that the gabbroic rocks are younger than the Yavapai series. The gabbro is younger than the quartz porphyry, as described earlier. Quartz diorite dikes cut the gabbro west of Ash Creek, proving that the quartz diorite is younger.

The gabbroic bodies generally are essentially concordant to the structure of the host rocks, and the arcuate form of several masses of gabbro south of Mingus Mountain reflect the trough of the plunging major syncline in that region. In detail, however, they crosscut the bedded rocks, and the form of the mass of gabbro at Jerome is irregular as discussed in the section on the geology of the United Verde mine, page 104.

West of Spud Mountain in the Prescott quadrangle, some gabbro is essentially parallel to the northward-trending Green Gulch volcanics but the dikelike masses of gabbro have a northeast trend, parallel to the Chap-

arral and Spud faults. Renewed movement on these faults produced a northeastward-trending foliation in some of the gabbro.

Diabase dikes, from 50 to 100 feet wide, are spatially associated with some of the larger masses of gabbro, particularly those in Lonesome Valley. The dikes cut the gabbro as well as the Yavapai series, but because of the close spatial relationship and mineralogic similarity to the gabbro, the diabase is presumably related genetically to the gabbro.

General characters

The gabbro is a dark granular rock containing hornblende and saussuritized plagioclase. The ratio of hornblende to feldspar is variable and the rocks of low hornblende-content are termed diorite. The diorite porphyry contains phenocrysts of saussuritized plagioclase and hornblende embedded in a fine-grained groundmass of feldspar and hornblende. The contacts between diorite porphyry, diorite, and gabbro are gradational and some of the larger masses contain all three facies. Diorite porphyry generally is present at the margins of the larger masses or in small masses and dikes, but there are also many dikes of granular diorite and gabbro.

The grain size is variable; some coarse-grained facies contain crystals ranging from 5 to 10 millimeters in length, but generally most of the granular rocks are medium grained, the crystals range from 3 to 5 millimeters in length.

All the gabbroic rocks are metamorphosed, and perhaps should be termed metadiorite, metagabbro, and metadiorite porphyry, but the original igneous textures are preserved and the extent of recrystallization is revealed only by microscopic study. A few relict augite grains were recognized in scattered samples, but augite generally is replaced by amphibole or chlorite. The amphibole is variable; it ranges in composition from fibrous green actinolitic amphibole, to needles and coarse crystals of bluish-green amphibole, to coarse crystals of greenish-brown amphibole (hornblende). The brownish varieties of amphibole are limited to outcrops of gabbro near the quartz diorite, and probably represent some thermal metamorphic effect. Actinolitic hornblende is more common in the masses of gabbroic rocks west of the Shylock fault, but it is present in the gabbroic mass west of the United Verde mine, subordinate to chlorite.

The plagioclase is altered to saussurite, a granular aggregate of clinozoisite and zoisite, containing rims or inclusions of albitic plagioclase, and scarce sericite. Locally, remnants of andesine or labradorite were observed. Calcite is a common associate with zoisite in the altered plagioclase of the gabbro at the United Verde mine.

Ilmenite(?) is common in triangular networks in many specimens of the gabbro, but in chlorite-rich facies, leucoxene takes its place. Coarse apatite crystals are common in all specimens. Veinlets and aggregates of epidote are in many of the thin sections.

Quartz grains interstitial to the altered feldspar and hornblende are common in most of the rocks containing a moderate amount of mafic minerals, but in the mafic-rich facies, quartz is absent. Some of the quartz may be primary, but much of it may be a product of metamorphism.

The diorite porphyry shows the same variation in mineral composition as the granular gabbro, and some relict augite phenocrysts partly altered to chlorite were noted in thin sections under the microscope. The groundmass is finely crystalline, composed of albitic plagioclase, quartz, and greenish-blue hornblende in some specimens, and of andesine, chlorite, and magnetite in other specimens. The plagioclase phenocrysts are altered to clinozoisite and zoisite, associated with albitic feldspar.

The diabase has a relict diabasic texture recognizable on weathered outcrops and in thin section. The euhedral plagioclase is changed to granular clinoboisite and zoisite rimmed by albite and embedded in clots of amphibole crystals.

The pyroxenite mass east of Mingus Mountain consists of granular augite crystals, from 5 to 10 millimeters long, separating a few interstitial saussuritized plagioclase crystals. In thin section, the pale augite (diallage?) is rimmed by brownish to bluish-green hornblende and chlorite. Nests of chlorite and epidote are present. The zoisite aggregates formed from original plagioclase comprise less than 10 percent of the rock.

Six complete and one partial chemical analyses are available of the gabbroic rocks (table 12); five of the six complete analyses are of the gabbro exposed at Jerome, the United Verde diorite of Reber (1922, p. 14). The average of the six complete analyses compares closely with the average analysis of 41 gabbro analyses compiled by Daly. The average of 70 diorite analyses, compiled by Daly, is given for comparison. The exact chemical and mineral composition of the original rock is in doubt because of the alteration, but the ubiquitous development of zoisite and clinozoisite and limited amount of albitic feldspar in the pseudomorphs of original plagioclase feldspar indicate that the original plagioclase must have been of the calcic variety, bytownite or anorthite, additional proof of the gabbroic character of most of these rocks. However,

TABLE 12.—*Chemical analyses of gabbro, in percent*

	1	2	3	4	5	6	7	8	9	10
SiO_2	45.73	45.22	51.16	48.90	49.50	50.84	45.60	48.57	48.24	56.77
Al_2O_3	19.45	21.94	19.55	22.76	19.08	19.17	20.40	20.32	17.88	16.67
Fe_2O_3	5.28	------	------	------	------	------	11.00	------	------	------
FeO	3.18	[1]8.14	[1]9.88	[1]7.95	[1]9.80	[1]9.50	------	[1]8.87	[1]8.79	[1]7.24
MnO	------	.14	.16	.13	.14	.15	------	.12	.13	.13
MgO	6.24	7.46	4.00	3.11	4.27	4.15	4.60	4.87	7.51	4.17
CaO	13.86	9.51	7.55	10.33	6.41	7.23	10.70	9.15	10.99	6.74
K_2O	.32	.57	.23	.35	.28	.30	3.94	.34	.89	2.12
Na_2O	.64	2.30	3.03	3.03	4.12	2.87		2.66	2.55	3.39
TiO_2	.23	.44	.90	.66	.80	.79	------	.64	.97	.84
H_2O-	1.57	.18	.18	.12	.19	.22	3.76	4.02	1.45	1.36
H_2O+	3.56	4.37	3.49	3.07	3.61	3.58				
S	------	Tr.	Tr.	.18	Tr.	.02	------	------	------	------
Cu	------	.01	.01	.03	.01	.01	------	------	------	------
Zn	------	.02	.07	None	None	.03	------	------	------	------
Fe	------	.02	.13	.21	.10	.08	------	------	------	------
SO_3	------	None	.02	.05	.01	.01	------	------	------	------
CO_2	.28	.27	.28	.26	1.87	1.32	------	------	------	------
Total	100.34	100.77	100.64	101.14	100.24	100.27	------	------	------	------
Specific gravity	------	2.93	2.92	2.93	2.84	2.87	------	------	------	------

[1] Fe_2O_3 calculated as FeO.

1. West of Yava Wash, Bradshaw Mountains quadrangle, G. Steiger, analyst. From Jaggar, T. A., and Palache, Charles (1905, p. 4). On Mayer quadrangle (1949 edition), Yava Wash is labeled "Yarber Wash."
2. United Verde mine. Analyst, W. O. Hamilton.
3. United Verde mine 500 level. Analyst, W. O. Hamilton.
4. United Verde mine 1,500 level. Analyst, W. O. Hamilton.
5. United Verde mine 2,100 level. Analyst, W. O. Hamilton.
6. United Verde mine 3,000 level. Analyst, W. O. Hamilton.
7. Surface of United Verde mine; average sample of coarse-, medium-, and fine-grained facies collected by L. E. Reber, Jr. Analyst, J. A. Handy. Analyses 2–7, furnished through courtesy of Phelps Dodge Corp.
8. Average of analyses 1–6.
9. Average of 41 gabbro analyses from Daly (1933, p. 17).
10. Average of 70 diorite analyses from Daly (1933, p. 16).

some of the facies containing limited amounts of mafic minerals may have been dioritic, though it is impossible to postulate the composition of the original feldspar.

QUARTZ DIORITE

Name and distribution

Quartz diorite crops out throughout much of the southern half of the Mingus Mountain quadrangle, where it forms large erosional mounds that interrupt the usual drab gray cover of chaparral. Smaller bodies are exposed west of the Iron King mine in the Prescott and Mount Union quadrangles (pl. 1). Lindgren (1926, p. 16) called the plutonic rock in the Mingus Mountain quadrangle granite, and correlated it with the Bradshaw granite, the term given by Jaggar and Palache (1905) for most of the plutonic rock in the Bradshaw Mountains quadrangle. Work currently done by M. H. Krieger in the Prescott quadrangle shows that the Bradshaw granite is a complex of diverse petrographic types, and that true granite is rare. In a broad sense, the quartz diorite in the Jerome area belongs to the Bradshaw granite but in this report, the term will not be used, and the more precise nomenclature of quartz diorite is adopted.

The west margin of the quartz diorite in the Mingus Mountain quadrangle is essentially the Shylock fault, although some small bodies of quartz diorite appear to the west of the fault. The most northerly of these bodies is a narrow strip, about 2,500 feet long, exposed north of the Jerome highway (1,340,000 N.; 415,500 E.). The largest of these separate blocks is south of the Jerome highway (1,327,000 N.; 418,000 E.). The smallest body is about a mile west of the Shylock fault (1,320,500 N.; 414,000 E.) where the northern and southern extensions are covered by Quaternary gravel. The most southerly exposure of quartz diorite is east of the Shylock fault, and separated from the main mass by older gabbro (pl. 1).

The quartz diorite in the Mingus Mountain quadrangle is partly separated by a southeastward-trending prong of Ash Creek group and intrusive gabbro, and the eastern part of the quartz diorite includes the basin surrounding the village of Cherry. The Verde fault marks the east margin and along the southern border of the quadrangle, the quartz diorite is largely covered by the Tertiary Hickey formation.

In the western part, the quartz diorite is deeply weathered to a sandy aggregate beneath the Tertiary sedimentary rocks and in the areas between the Hickey outcrops. Presumably this weathering occurred before the deposition of the Hickey, for in the areas of present erosion away from the Hickey exposures bold outcrops weather from the quartz diorite.

Two outlying masses of quartz diorite are exposed in the Mingus Mountain quadrangle, north and northwest of the eastern part of the quartz diorite mass. The larger mass is in Ward Pocket, on the south wall of Black Canyon (1,325,000 N.; 450,000 E.) and the smaller mass, only 600 feet in its longest dimension, is more than 2½ miles from the main body of quartz

diorite. This small mass is exposed in one of the tributaries to Ash Creek (1,317,800 N.; 435,500 E.).

West of the Iron King mine, the most southerly exposure of quartz diorite is shown in the southwest corner of plate 1, and only the northern tip of this body has been mapped. A band of quartz diorite, more than 3 miles long and 1,500 feet maximum width, is exposed in the Prescott quadrangle between the Chaparral and Spud faults. A narrow strip of quartz diorite is present along the Spud fault and this strip almost connects with the larger southern body.

Relation to other rocks

The quartz diorite is the youngest of the Precambrian major intrusive bodies exposed in the Jerome area. Apophyses of quartz diorite intrude the Ash Creek group, and the Ash Creek rocks adjacent to the quartz diorite are metamorphosed to hornfels. Southwest of the Iron King mine, apophyses of quartz diorite intrude the breccia facies of the Spud Mountain volcanics, and prove that the plutonic rock is intrusive into the Alder group. Along the Shylock fault, however, the quartz diorite is in fault contact with the Alder group; the small bodies of quartz diorite west of this fault are also in fault relationship to the Alder group.

The quartz diorite is in contact with quartz porphyry east of the Shylock fault along the southern margin of the Mingus Mountain quadrangle (near 420,000 E.), but both rocks are altered and no diagnostic evidence was found as to their age relationship. In the most northerly exposures of quartz diorite, west of the Verde fault, an eastward-trending dike of quartz porphyry is clearly cut off by the younger quartz diorite (pl. 1, 1,325,700 N.; 455,800 E.).

Several gabbro bodies are in contact with the quartz diorite, but in most places the contact is sharp and straight, and not conclusive as to age relationship. However, one small body of gabbro, west of Ash Creek (pl. 1, 1,300,800 N.; 437,000 E.) clearly contains apophyses of quartz diorite, proving the younger age of the latter rock. In addition, the widespread distribution of gabbro throughout the Ash Creek terrane and absence in the quartz diorite terrane, strongly suggests that the gabbro is older. The elongate body of gabbro cut off by quartz diorite (1,292,000 N.; 447,000 E.) has a pattern that also suggests a later age of the quartz diorite. The available evidence practically proves that the quartz diorite is younger than the gabbro.

The dike swarms of granodiorite porphyry in the Mingus Mountain quadrangle clearly cut the quartz diorite.

The contact of the quartz diorite and Ash Creek group from the Shylock fault (1,316,700 N.; 419,200 E.) southeastward is essentially parallel to the strike of the Ash Creek group and the Tex syncline. Northward from 1,292,000 N.; 447,700 E., this contact largely transgresses the structure of the group except at the north where the contact is again almost parallel to its strike. This broad relationship indicates that the folding in the Ash Creek group preceded the intrusion of the quartz diorite, a conclusion substantiated by the evidence that the older quartz porphyry was intruded during the period of major folding.

The quartz diorite west and southwest of the Iron King mine occurs as elongated blocks parallel to the Chaparral and Spud faults. To account for the dike-like masses of quartz diorite along the faults, either intrusion during deformation, or fault slivers are required.

In the northeastern part of the quartz diorite terrane, the Tapeats sandstone(?) and Martin limestone rest upon an eroded surface carved from the quartz diorite, proving on stratigraphic evidence that the quartz diorite is older than the Devonian. H. W. Jaffe of the U. S. Geological Survey (written communication, 1954) has determined the age of zircon from the quartz diorite, using the Larsen method, as 1,050 million years. The quartz diorite was collected east of the bridge where the Cherry Creek road crosses Ash Creek. This age determination of the zircon proves that the quartz diorite is Precambrian. Jaffe (written communication, 1954) determined the age of zircon from granodiorite, south of Prescott, as 910 million years. The granodiorite is one of the intrusive bodies that makes up the complex of Bradshaw granite.

The Tertiary lava flows and sedimentary rocks were deposited on an irregular erosional surface cut in the quartz diorite; the valleys were filled first.

General characters

Mingus Mountain quadrangle.—The quartz diorite is essentially a massive rock containing very few inclusions of foreign material. Most of the observed inclusions are equidimensional and average between 3 and 4 inches in diameter. Locally, a few as large as 1 foot across are present. Where hornblende is a major constituent, a crude orientation of the hornblende prisms can be measured; in large part this orientation produced a planar structure, and in part it is linear. In some exposures, the biotite crystals also show a crude alinement, and in others, the sporadic inclusions are alined parallel to the observed orientation of the hornblende. In many exposures, no positive orientation could be determined. Measurements of the observed orientation are plotted on plate 1, and this structure is interpreted to be due to magmatic flow.

Flow structure generally is more common in the western part of the quartz diorite body in the Mingus Moun-

tain quadrangle. This may be due either to better exposures, or to larger hornblende crystals. Adjacent to the Shylock fault, the orientation is commonly northwest, essentially parallel to the northeastward-trending contact with the Ash Creek group, but near the end of the prong of older quartz diorite rock, northeast trends are present in the flow structure. In the eastern part of the quartz diorite, fewer observations of flow structure could be made; these, however, also largely trend northwestward, except east of Cherry near the Verde fault, where northeast trends are dominant. Along the western contact of this eastern part of the quartz diorite, a few north trends of flow structure may be related to the northward-trending quartz diorite contact.

The quartz diorite in the western part of the quadrangle is coarser grained and contains a higher content of orthoclase than the quartz diorite in the eastern part. The western facies has a seriate texture and the average grain size of the rock ranges from 3 to 4 millimeters. Scattered stout hornblende prisms commonly are 8 to 10 millimeters long and the associated plagioclase crystals range from 4 to 5 millimeters in length. Most of the hornblende, plagioclase, and biotite crystals in the rock are about the average grain size. The quartz grains are smaller, ranging from 1 to 2 millimeters across.

Microscopic studies of the western facies reveal that the plagioclase is zoned; most of the plagioclase ranges in composition from andesine to sodic labradorite and the rims are albitic. In many specimens, the calcic core has been altered to saussurite, an aggregate of clinozoisite, epidote, sericite, and calcite, separated by an albitic base. Orthoclase in part forms large poikilitic crystals, containing all the other minerals as inclusions; much of the orthoclase is interstitial. Myrmekite occurs at the margins of some plagioclase where it is in contact with orthoclase. The hornblende is pleochroic in shades of green to yellowish brown. Biotite in part is altered to chlorite. Quartz occurs commonly as an interlocking aggregate of granules. Accessory minerals include magnetite and apatite, closely associated with the mafic minerals. Sphene is erratically distributed. Epidote is a common secondary mineral, flooding some specimens or appearing as veinlets. The modal composition of the western facies is given in table 13, showing that the rock is quartz-biotite-hornblende diorite.

In the eastern facies, the texture of the quartz diorite is also seriate. The larger slender hornblende and some plagioclase crystals range from 4 to 5 millimeters in length, but the average grain size ranges from 1 to 2 millimeters. The mineralogy of the eastern facies is essentially the same as the western although the orthoclase content is lower. The modal composition is given in table 13.

TABLE 13.—*Modal composition of quartz diorite*

	1	2	3
Orthoclase	9	4	19
Plagioclase	56	61	53
Quartz	27	26	14
Hornblende	2	4	11
Biotite	5	4	2
Accessory minerals	1	1	1
	100	100	100

1. Western facies, average of 4 thin sections.
2. Eastern facies, average of 5 thin sections.
3. Local marginal facies, 1 thin section.

In the area between 440,000 E. and 450,000 E., and south of the prong of Ash Creek rocks, there is a broad transition zone from the coarser textured western facies to the finer textured eastern facies. In this zone, the larger hornblende prisms decrease in length and are more slender in habit than those to the west. The larger plagioclase crystals gradually decrease in size to the east, and the orthoclase content decreases. No sharp boundary between the two facies can be found and the prevailing northwest trend of the flow structure in both the eastern and western parts of the quartz diorite indicate that the quartz diorite came in almost simultaneously, and that the coarse- and fine-textured parts are simply facies of one major intrusive body.

Where the westernmost dike swarm cuts the quartz diorite contact (1,307,000 N.; 425,000 E.), patches of a porphyritic facies of quartz diorite are exposed, cut by dikes of normal quartz diorite. The plagioclase and hornblende phenocrysts in the porphyritic facies range from 4 to 5 millimeters in length, and the fine-grained (average 1 mm) groundmass consists of hornblende needles, plagioclase, and quartz. Presumably this porphyritic facies represents an earlier intrusion of the quartz diorite magma.

Southeast of the Shylock mine (1,315,000 N.; 421,500 E.) a border facies of the quartz diorite is more equigranular (average grain size 4 mm) and richer in hornblende and orthoclase than most of the quartz diorite (table 13, analysis 3). This border facies contains enough orthoclase to classify the rock as granodiorite. A border facies east of the prong of Ash Creek rocks (1,296,500 N.; 445,000 E.) is hornblende-rich and contains no orthoclase, and biotite is rare.

A facies of quartz diorite is exposed near 1,286,000 N.–418,500 E. (pl. 1), in which conspicuous large euhedral plagioclase crystals indicate a porphyritic texture, but close examination reveals a seriate texture. This facies contains less quartz and biotite and more hornblende than the typical western facies of quartz

diorite, which is intermixed with this seriatic facies. A primary flow structure is parallel to the contacts of the included masses of Spud Mountain volcanics.

The southernmost outcrops of quartz diorite in the Mingus Mountain quadrangle (east of the Shylock fault) are lighter colored than the outcrops of the main body, in part because of alteration and in part owing to a lower content of mafic minerals. The texture of this quartz diorite is seriate and the average grain size is about 3 millimeters. A few scattered hornblende crystals were observed in the outcrops, but biotite, altered to chlorite, is the only mafic mineral observed in thin sections. The plagioclase is completely altered to saussurite. The ratio of plagioclase : orthoclase : quartz is essentially the same as in the main body of quartz diorite. No sphene was noted, but zircon is present as an accessory mineral.

Adjacent to the Shylock fault, the quartz diorite is changed to buff and orange-colored foliated rock; thin sections reveal that the quartz is granulated, plagioclase sheared and partly replaced by sericite, and some calcite is present. The mafic minerals are largely destroyed. The smaller bodies of quartz diorite west of the Shylock fault are similarly altered and foliated, particularly along the margins where the original texture is completely destroyed, and it is only in the centers of these bodies that the rock can be recognized as quartz diorite. Northeastward-trending shear zones well within the main body of quartz diorite show similar foliaiton and alteration; many of these shear zones are occupied by quartz-carbonate veins, locally containing stains of copper carbonate.

West of the Iron King mine.—The larger body of quartz diorite, exposed in the southwest corner of the Jerome area (pl. 1), is similar in texture and mineral composition to the finer grained facies exposed in the eastern part of the Mingus Mountain quadrangle. Locally along the margins, the rock has a weak foliation. The elongated bodies between Chaparral and Spud faults are strongly sheared, and locally, highly schistose. In the foliated facies, chlorite has formed from the mafic minerals, and sericite from the feldspar. Toward the southwest, the degree of shearing gradually decreases, and relicts of quartz diorite prove the original character of the highly foliated rock.

GRANODIORITE PORPHYRY DIKES

Distribution

Dikes of granodiorite porphyry, trending northward to north-northeastward, appear east of the Shylock fault in the Mingus Mountain quadrangle (pl. 1). The most westerly dikes form a system that extends from near the southern contact of the quartz diorite (1,283,000 N.) northward to well within the Grapevine Gulch formation (1,325,000 N.), a distance of more than 8 miles. A second dike system, more than 12 miles long, essentially cuts through the middle of the quadrangle, and reaches within 3 miles of the northern boundary. Two isolated dikes crop out in Black Canyon and extend northward; the south end of this minor dike system is on the south wall of Black Canyon (1,325,000 N., 452,000 E.).

Relation to other rocks

The dikes of granodiorite porphyry are the youngest of the Precambrian igneous rocks, for they intrude the quartz diorite, quartz porphyry, and gabbro, and the Ash Creek group. Between Gaddes Canyon and Oak Wash, and southeast of the Brindle Pup mine (1,324,500 N.; 441,000 E.) the dikes are overlain by the Paleozoic rocks, proving conclusively that the dikes are older than the Tapeats sandstone(?). In the south-central part of the quadrangle, the dikes are overlain in part by the Hickey formation.

Southward from Gaddes Canyon (1,332,500 N.; 442,500 E.), one dike extends along a northward-trending fault that displaces the lower formations of the Ash Creek group.

General characters

The dikes of granodiorite porphyry generally are more resistant to erosion than the intruded host rocks. The most northerly dikes east of Mingus Mountain are more deeply weathered than the southern dikes, and do not form prominent outcrops. In the southern half of the quadrangle, where quartz diorite is the host rock, the dikes have a tendency to crop out on northward-trending ridges. Wider dikes in the southern part of the quadrangle have rounded boulderlike outcrops. The dike rocks are light gray on fresh surfaces, and weather buff.

The dikes range in width from about 25 feet as the minimum to 300 feet as the maximum; most are between 50 and 100 feet. Wider dikes may have screens of older rocks within them. Near the Brindle Pup mine (1,327,000 N.; 441,000 E.) one of the dikes widens locally to 1,200 feet; this bulbous part is more than 2,500 feet long. Northwest of the Brindle Pup mine (1,327,000 N.; 439,000 E.) a stubby dike of porphyry, 300 feet wide and 2,000 feet long, is well exposed in Black Canyon.

Many dikes are almost leucocratic, and consist of quartz and dulled plagioclase phenocrysts in a pale-gray aphanitic groundmass. Other dikes contain abundant hornblende phenocrysts in addition to the quartz and feldspar, and mafic minerals are recognizable in the groundmass. Some of the wider dikes, and particularly the bulbous dike at the Brindle Pup mine

and the stubby dike in Black Canyon, have a seriate texture. The quartz and plagioclase crystals range from 0.5 to 2 millimeters.

At Johnson Wash ranger station (pl. 1, 1,291,000 N.; 463,000 E.) the dike is multiple; the western part contains phenocrysts 3–4 millimeters long, whereas the eastern part has phenocrysts only 1–2 millimeters long. Where this dike divides (1,294,000 N.; 464,700 E.), the western part is finer textured.

The thin sections of granodiorite porphyry all show that it is appreciably altered. The plagioclase is poorly saussuritized, and most of the hornblende is altered to epidote and chlorite. No biotite was observed, but chlorite or magnetite aggregates suggest pseudomorphs after biotite. The groundmass ranges from microcrystalline to finely crystalline where quartz and alkalic feldspar can be recognized. In some of the coarser textured groundmass, local micrographic intergrowths of quartz and alkalic feldspar are present. Accessory minerals recognized are sphene and apatite. Secondary minerals include epidote, calcite(?), and sericite.

The classification of these rocks is difficult because of the intense alteration. The chemical analysis of one of the dikes from the east side of Mingus Mountain (table 14), indicates a silica content higher than the average quartz diorite; solely on this basis, the dikes are classified as granodiorite porphyry. Obviously there is much variation, chemically, and some of the dikes may be quartz diorite porphyry and others may be quartz monzonite porphyry.

TABLE 14.—*Chemical composition of granodiorite porphyry*

[Dike exposed on Copper Chief road (1,349,000 N.; 448,800 E.). Analysis furnished through courtesy of Phelps Dodge Corp.]

	Percent
SiO_2	69.37
Al_2O_3	15.94
FeO [1]	3.26
MgO	2.17
CaO	3.00
K_2O	2.01
Na_2O	3.10
H_2O + 100°C	1.14
CO_2	.53
TiO_2	.20
Total	100.72

[1] Total iron reported as FeO.

The general spatial relationship to the quartz diorite and similarity in mineral composition indicate that the dikes are genetically related to the quartz diorite, and represent late injections from the quartz diorite magma chamber.

ALASKITE AND APLITE

Distribution

The alaskite and aplite are in small discrete masses chiefly in the southwest corner of the Jerome area, but a few aplite dikes are in the south-central part of the Jerome area in the tongue of Grapevine Gulch formation and gabbro that extends into the quartz diorite, and one lenticular mass is near 1,318,200 N.–451,100 E. The alaskite and aplite dikes in the southwest corner occur in the Chaparral and the Green Gulch volcanics. Those in the Chaparral volcanics consist of two elongate lenses, which are 2 and 2½ miles in outcrop length. The complete horizontal extent of one of these lenses is exposed, but only the south end of the other crops out, the northern limit is overlain by the Tertiary fill in Lonesome Valley. In the Green Gulch volcanics a short distance west of these two masses, there are seven small lenses or pods of alaskite and aplite; the longest lens crops out for about half a mile, and the widest is about 400 feet across.

Relation to other rocks

The alaskite and aplite in the southwest corner of the Jerome area intrude the Chaparral and the Green Gulch volcanics. In each of these formations, the shape of the alaskitic and aplitic masses ranges from dikelike to lenticular pods, and all masses are oriented with their long dimension parallel to the trend of the regional foliation. In the intensely deformed Chaparral volcanics the contact of the alaskitic and aplitic masses, in all probability, is faulted, so that it is doubtful that the contiguous rocks were those actually adjacent to the aplite and alaskite at the time of intrusion. The dikes in the south-central part of the Jerome area clearly cut the Grapevine Gulch formation and gabbro, and the dike at 1,318,200 N.; 451,100 E. cuts the quartz diorite. In other areas, other dikes of aplite, too small to illustrate on the regional maps, cut the quartz diorite.

The spatial relationship of the aplite dikes surrounding the periphery of the eastern mass of quartz diorite and of the dikes that cut it imply a genetic relationship to the quartz diorite. The age and genetic relationship of the alaskite and aplite in the southwest corner of the Jerome area are more difficult to determine. Certainly, the aplite dikes cut gabbro and quartz diorite, and may be genetically related to the latter, like the aplite dikes to the east. Perhaps the aplite dikes are related to the alaskite, or both aplite and alaskite are genetically related to the quartz diorite. The foliation in the aplite and alaskite clearly demonstrates that both are older than the youngest known period of Precambrian deformation, which resulted in the metamorphic structures in the Chaparral volcanics.

General character

The unweathered alaskite and aplite are pale red or grayish-orange pink; locally they are nearly white owing to alteration. Dynamic metamorphism induced a pronounced foliation in much of the alaskite and aplite

in the southwest corner of the Jerome area. Relicts of the original rock particularly in the northern part of the alaskite mass reveal the original texture and mineralogic composition. Shearing along the margins of the eastern mass of alaskite in the Chaparral volcanics was especially intense; so intense that the narrow southern whiplike prong of this mass may be entirely due to deformation. The contacts of this mass are zones of mechanical mixing of alaskite and rhyolitic tuffaceous rock. Where intensely sheared, these rocks are schistose or gneissose. The quartz grains have been drawn out into "ribbons," and the feldspar partly recrystallized into elongated aggregates.

Locally residual orthoclase augen interrupt the even flow of the metamorphic layering. Mafic minerals were largely removed through shearing; but locally metamorphic biotite flakes occur concentrated along certain planes, and locally sparse disseminated chlorite gives a greenish cast to the rock.

The texture of the less sheared alaskite relicts suggests a medium-grained granitic rock. Most of the alaskite was probably equigranular, but the size of some of the augen indicates coarse-grained or porphyritic facies. The texture of the aplite apparently was normal, that is aplitic or saccharoidal.

The alaskite consists chiefly of orthoclase, microcline, perthite, albite, and quartz. The minor constituents are biotite, chlorite, sericite, epidote, sphene, zircon magnetite, and ilmenite. Small amounts of hornblende, muscovite, and pyrite were observed, but they probably resulted from contamination or hydrothermal alteration. The aplite was not studied in thin section. Megascopically it consists of quartz, pink feldspar (presumably chiefly potash feldspar), and locally a trace of chlorite. The aplite dikes associated with the eastern mass of quartz diorite are granular, but locally they contain pegmatitic centers from 1 to 2 inches wide.

QUARTZ AND JASPER VEINS

Under the heading of Quartz and jasper veins, we include all deposits formed through introduction of appreciable amounts of silica. We have also included on plate 1 the alteration zone at the Iron King deposit and the Silver Belt-McCabe and Kit Carson veins. Introduced silica is not abundant in these two veins, which are included here in order to avoid use of a separate symbol. The alteration zone at the Iron King deposit contains much introduced sericite, pyrite, and probably some ankerite in addition to the silica. This alteration zone and the Silver Belt-McCabe and the Kit Carson veins are described under the section on ore deposits, and will therefore not be considered at length here.

Distribution

Quartz and jasper veins are most abundant in the rocks of the Alder group. In the Chaparral volcanics and the Texas Gulch formation most, if not all, the veins are quartz; only the larger ones are plotted on plate 1. Veins and pods of quartz and jasper are particularly abundant in the andesitic tuffs of the Spud Mountain and the Green Gulch volcanics. In these formations, jasper is more abundant than quartz, and only jasper occurs in the largest mass of andesitic tuffaceous rocks (Spud Mountain volcanics) enclosed by the Texas Gulch formation.

In the rocks of the Ash Creek group, jasper is abundant in three areas; in the Precambrian rocks between the United Verde and Copper Chief deposits, in the Deception rhyolite near 1,318,000 N.; 449,000 E., and in the Deception rhyolite along and near the contact with the overlying Grapevine Gulch formation southeast of Kendall Peak (1,330,000 N.; 430,000 E.).

Occurrence

The quartz and jasper in the rocks of the Alder group were introduced into sheared zones or perhaps locally in joints. In many places the sheared zones parallel the foliation, but commonly they cut the foliation at low angles, and a few cut the foliation at high angles. In a few places, the sheared zones were arranged en echelon, resulting in short, relatively thick jasper veins, each slightly offset from the adjacent one.

The jasper between the United Verde and Copper Chief deposits is closely associated with quartz porphyry, either occurring along the contact of the quartz porphyry with the adjacent rocks, or in the rocks adjacent to the quartz porphyry. We do not know whether this association is genetic, structural, or both. The jasper near Kendall Peak apparently has been chiefly localized by the contact between the Deception rhyolite and the Grapevine Gulch formation. Movement may have occurred along this contact, judging by the structural discordance between the two formations, although the lenticularity of the units in both formations prevents a firm conclusion. The jasper veins near 1,318,000 N.; 449,000 E. appear to be controlled by sheared zones of uncertain origin.

Local spots in the jasper veins were fractured and mineralized with pyrite and presumably chalcopyrite that produced enough copper stain to encourage the prospector into fruitless exploration; many jasper veins are pockmarked with shallow prospect pits and adits. So far as we know, no jasper vein has yielded sufficient ore to justify the exploration.

One jasper vein is overlain by the Tapeats sandstone(?). Inasmuch as jasper veins do not cut the

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Tapeats(?) and younger rocks, most of the jasper, and presumably also the quartz veins, are probably Precambrian. If so, they probably formed after the regional deformation, for they show no signs of deformation other than crackling, except for the alteration zone associated with the Iron King deposit. This zone was intensely sheared subsequent to the introduction of quartz and other constituents that together comprise the alteration.

REGIONAL CORRELATION OF THE OLDER PRECAMBRIAN ROCKS

The Precambrian rocks of Arizona, at times, have been termed Archean and Algonkian, and the same rocks have been placed in both categories by different writers. Butler and Wilson (1938, p. 11) subdivided the Precambrian rocks of Arizona into older and younger Precambrian, a preferable designation, which has been adopted for this report. It should be emphasized that there is no assurance that all the rocks grouped as older Precambrian are necessarily of the same age. These are terms used in a provincial sense to indicate that for the particular area under discussion some evidence is available to indicate that the rocks are early or late in the Precambrian sequence.

Rocks that have been designated as younger Precambrian in Arizona include the Grand Canyon series and the Apache group. The Grand Canyon series, exposed in the Grand Canyon of the Colorado River north of the Jerome area, has been divided into two groups, the Unkar and Chuar, that contain more than 10,000 feet of unmetamorphosed sedimentary rock at an angular unconformity below the basal Cambrian sandstone (Darton, 1925, p. 23–27). The Apache group consists of about 1,000 feet of sedimentary rock and is exposed in the Mazatzal Mountains, near Globe, and in southeastern Arizona. The Apache group is separated from the overlying Cambrian quartzite or Devonian limestone by a disconformity (Peterson, and others, 1951, p. 13). Darton (1925, p. 36) and Ransome (1932, p. 6) suggested that the Apache group is in general equivalent to the Grand Canyon series, because of lithologic similarities.

The Grand Canyon series and the Apache group unconformably overlie metamorphic and granitic rocks; this major unconformity has been used as the boundary between the older and younger Precambrian rocks in Arizona.

At the Grand Canyon, the older Precambrian rocks include the Vishnu schist, first described by Walcott (1890) and modified to include schist and gneiss by Noble and Hunter (1917) who suggested that at some future time it might be desirable to restrict Vishnu schist to mica schist and give new names to the gneiss. Campbell and Maxson (1938) proposed that the term "Vishnu schist" be discarded and that "Vishnu series" be used for the 25,000 feet of metasedimentary rocks exposed in the Grand Canyon. Originally fine grained argillaceous sandstone and sandy shale have been metamorphosed to quartzite, sericite-quartzite, and quartz-mica schist. In addition, they recognized a sequence of basaltic lava and tuff, now represented by amphibolite, in which relict pillow and amygdaloidal structures proved the volcanic character. No name was given by Campbell and Maxson for the metavolcanic rocks. The metasedimentary and metavolcanic rocks were intruded by plutonic rocks that range from quartz diorite to granite.

In the Globe area, Ransome (1903) first described and defined the Pinal schist, derived in part from quartzose sediments and intruded by granitic rocks; the Pinal schist and younger intrusive rocks are in turn unconformably overlain by the Apache group. During his later work in the Ray–Miami area, Ransome (1919, p. 35–37) recognized, in addition to metamorphosed impure sandstone or graywacke and shaly sediments, a metarhyolite or quartz latite, the composition of which he verified by a chemical analysis.

Cooper (Anderson, 1951, p. 1334) recognized the Pinal schist in the Little Dragoon Mountains, 60 miles east of Tucson, unconformably below the Apache group. The Pinal schist here includes (1) greenstone composed of amphibolite that probably represents a series of mafic flows or intrusives or both; (2) rhyolite, probably representing a thick flow; (3) a complex of quartz-feldspathic sedimentary rocks (arkose) and interbeds of sericite schist.

In the vicinity of Mazatzal Peak, Wilson (1939) divided the Yavapai series into three units: (1) intermediate to mafic flows, tuff, and interbeds of sedimentary material; (2) rhyolitic flows; (3) slate, argillaceous grit, conglomerate, and quartzite. The argillaceous grit is similar to the rhyolitic tuff unit in the Texas Gulch formation containing interbedded conglomerate, as exposed in the Jerome area. Wilson also recognized three formations younger than the Yavapai series: Deadman quartzite, Maverick shale, and Mazatzal quartzite. The Yavapai series and younger formations were folded, faulted, and intruded by granite before the deposition of the overlying Apache group.

At Bagdad, about 40 miles west of Prescott, metamorphic rocks similar to the Yavapai series in the Jerome area have been tentatively correlated with the Yavapai (Anderson, 1951, p. 1336). Three lithologic units were recognized: (1) andesitic-basaltic flows and minor sedimentary interbeds; (2) rhyolitic and rhyo-

litic-andesitic tuffaceous sedimentary rocks and shale interbeds; (3) shale, sandy shale, and sandstone, recrystallized to quartzite and mica schist. These rocks were folded and faulted before the intrusion of a diverse series of igneous rocks culminating in granite.

The repetitious and lenticular character of the flows and volcanic fragmental rocks in the Jerome area demonstrates that correlation of these types of rocks for more than a few miles is not justified. The distinct lithology of the Texas Gulch formation, consisting of alternating purple slate and rhyolitic tuff-conglomerate beds, may be useful for distant correlations; the similarity of the Texas Gulch formation to Wilson's (1939) series (unit 3 above) in the Mazatzal Mountains is so striking that we believe correlation is reasonable. On that basis, the Yavapai series in the Jerome area is older than the Apache group and can be assigned confidently to the older Precambrian. Even without this correlation, we would have no hesitation in placing the Yavapai series in the older Precambrian because of the complex structural and metamorphic history accompanied and followed by the invasion of plutonic rocks. At no place in Arizona is there evidence that the Grand Canyon series and Apache group were involved in a Precambrian orogeny comparable to that involving the Yavapai series in the Jerome area.

Lindgren (1926, p. 15) suggested that the Vishnu, Yavapai, and Pinal schists are essentially the same formation, and there is some merit to this suggestion. All three are definitely older Precambrian, all three have somewhat similar lithologic, and structural features; and the Vishnu and Pinal were deformed and intruded by granitic rocks before deposition of the Grand Canyon series and Apache group. The Yavapai is only known to be pre-Paleozoic on stratigraphic evidence, but only because no younger Precambrian rocks are associated with it. The only certain method of determining similarity in age of these nonfossiliferous Precambrian rocks is by tracing similar rocks from one area to the next, obviously impossible in Arizona with the widespread cover of younger rocks. Possibly such distinct lithologic units as the Texas Gulch formation can be used, but so far, this distinct unit has been found only in the Jerome area and in the Mazatzal Mountains. Precise correlation of basalt, rhyolite, tuff, and other rocks, from one region to another is impossible. Until some more accurate methods are found for age determinations of these nonfossiliferous rocks, no precise correlations of the Vishnu and Pinal schists with the Yavapai series are possible.

Some general statements can be made that provide evidence for regional interpretations. The older Precambrian rocks have been folded in the Little Dragoon and Mazatzal Mountains, and Bagdad and Jerome areas (Anderson, 1951, p. 1345). The trend of the folds generally is northwest, north, or northeast indicating general east-west compressive forces during the orogeny but Gilluly (written communication) reports that in the Dragoon Mountains, the trend is nearly east. In the areas where sufficient detailed work has been done to give certainty to the conclusions, only one period of orogeny is clearly demonstrated. In the Little Dragoon and Mazatzal Mountains, the orogeny is pre-Apache, whereas in the Jerome area, the orogeny can be dated on stratigraphic evidence only as pre-Paleozoic, but on the zircon age determination of the quartz diorite, the orogeny is older than 1,050 million years. The age of the granite at Bagdad, which is younger than the orogeny, is about 1,600 million years, determined by the radioactive age of potassium—argon—rhubidium on muscovite and lepidolite (L. T. Aldrich and G. L. Davis of the Carnegie Institution, Department of Terrestrial Magnetism and Geophysical Laboratory, oral communication, 1955). The age determinations of the quartz diorite from the Jerome area and granite from Bagdad indicate that the orogeny at these localities occurred well within the Precambrian. At the Grand Canyon, it has been established that orogeny followed by the intrusion of granitic rock caused the deformation and metamorphism of the Vishnu schist; the orogeny was older than the Grand Canyon series.

It is tempting to assume that the orogenies in these five separate areas in Arizona occurred at the same time, particularly because of the general parallelism of the folds where trends were determined. Wilson (1939) termed this probable widespread orogenic disturbance, the Mazatzal revolution. From a purely academic view, one might question this conclusion, for the Precambrian covers an immense period of time, and it would be surprising if only one period of orogeny occurred in Arizona during early Precambrian time. Hinds (1936) has suggested that two periods of orogeny and two periods of granitic intrusion occurred in Arizona before deposition of the younger Precambrian Grand Canyon series and Apache group, the Mazatzal quartzite in the Mazatzal Mountains marking the period of sedimentation between these orogenies. Because no positive angular unconformities have been found between the Mazatzal quartzite and the Yavapai series, some doubt exists as to the validity of this older period of orogeny and granitic invasion. Until more precise age determinations can be made, we have no basis of correlating the orogenies in these five separate areas.

UNCONFORMITY AT THE BASE OF THE PALEOZOIC ROCKS

A profound unconformity separates the older Precambrian rocks from the overlying Paleozoic sedimentary rocks. The acute deformation of the Yavapai series, accompanied and followed by the intrusion of a series of igneous rocks culminating in granitoid rocks, batholithic in scope, implies appreciable orogeny that marked the end of older Precambrian sedimentation and volcanism in the Jerome area. Erosion of the mountains formed by this orogeny exposed the metamorphosed Yavapai series and granitoid rocks underlying the basal Paleozoic sandstone.

The basal Paleozoic sandstone was deposited on a surface of low relief, but scattered hills as high as 400 feet rose above this surface. At two localities, 45 and 80 miles northwest of Jerome (see Darton and others, 1924) basal Paleozoic sandstone is similarly exposed, resting on granite. The extensive outcrops of sandstone also indicate a surface of low relief carved from the underlying granite.

At the Grand Canyon, the younger Precambrian Grand Canyon series was also deposited on an erosional surface of low relief carved from the older Precambrian Vishnu schist and associated granitic rocks. Subsequently, the Grand Canyon series was tilted and eroded appreciably before the deposition of the basal Cambrian sandstone. Near Globe, the younger Precambrian Apache group was deposited on an erosional surface cut from the older Precambrian Pinal schist and associated granitic rocks; the basal Paleozoic rocks are only disconformable to the underlying Apache group.

No sedimentary rocks equivalent to the younger Precambrian Grand Canyon series or Apache group have been recognized in the Jerome area. The record merely indicates long erosion, that generally is represented at the Grand Canyon and at Globe by a profound unconformity beneath the younger Precambrian rocks. Probably in the Jerome area, the unconformity separating the Yavapai series and granitoid rocks from the basal Paleozoic sandstone represents a longer period of erosion than the unconformities at the Grand Canyon and in the Globe area beneath the younger Precambrian rocks.

PALEOZOIC SEDIMENTARY ROCKS

The Paleozoic rocks, limited chiefly to the northern half of the Jerome area (pl. 1), have an aggregate thickness of about 1,200 feet (fig. 2). They range in age from Cambrian, to Pennsylvanian and Permian. The lowermost Paleozoic rocks are placed in the Cambrian with reservation, for they may be the basal part of the local Devonian system. The uppermost Paleozoic rocks are represented by the lower part of the Supai formation, which may be either Pennsylvanian or Permian.

The oldest Paleozoic formation is mainly a sandstone, that ranges in thickness from a few inches to 100 feet and is tentatively correlated with the Lower Cambrian Tapeats sandstone in the Grand Canyon. It unconformably overlies older Precambrian metamorphic rocks and underlies, with apparent conformity, the Martin limestone of Devonian age. The Martin, which is about 450 feet thick, comprises at least three and possibly four units. The Mississippian Redwall limestone overlies the Martin. It comprises three units that have an aggregate thickness of about 290 feet. The Supai, in turn, overlies the Redwall. Only about 370 feet of the basal part of the Supai formation occurs in the Jerome area, but north and east of Jerome the full thickness of Supai is exposed superbly in the brilliantly hued cliffs of the Mogollon Rim.

TAPEATS SANDSTONE(?)

Distribution

The Tapeats sandstone(?) underlies the entire summit area of Mingus Mountain, though it crops out only around the periphery owing to the cover of younger rocks. This summit area occupies about 150 square miles in the north-central part of the Jerome area.

Additional small areas underlain by Tapeats(?) occur as outliers surrounding the margin of the summit area and, to a lesser extent, as inliers in the younger Paleozoic rocks displaced by the Verde fault, which bounds the Black Hills on the east. The small outlying masses of Tapeats(?), some of which are capped by younger rocks, mainly lie south of the summit area. Commonly they form the crests of the higher peaks and ridges, such as those northwest of Cherry. The most southerly exposure of Tapeats(?) is in the upper drainage area of Buckbed Wash (pl. 1, 1,292,000 N.; 472,000 E.).

Small areas underlain by Tapeats(?) occur on the east side of the Verde fault. At Jerome a small mass of highly deformed Tapeats sandstone(?) is within the Verde fault zone, and about 1,200 feet east-northeast of Jerome a continuous outcrop of Tapeats(?) is exposed for about 3,000 feet in the opencuts of the Verde Tunnel and Smelter Railroad, which connected the smelter at Clarkdale to the United Verde mine (pl. 1, 1,368,000 N.; 444,000 E.). Between Blowout Creek and Copper Chief road (1,348,000–1,353,500 N.; 454,000 E.) the Tapeats sandstone(?) crops out in several places within an area about 1 square mile. Tapeats sandstone(?) crops out continuously along the west side of Mingus Mountain as far north as Coyote Springs and sporadically from Coyote Springs northward nearly to the northern boundary of the Mingus Mountain quadrangle.

System	Series	Formation	Map symbol	Thickness (feet)	Section	Kind of rock
QUATERNARY AND TERTIARY	*Pleistocene*	Younger gravels	Qg	20–50		Heterogeneous unconsolidated accumulation of boulders, cobbles, pebbles, and some finer grained sediments
	Pliocene or Pleistocene	Verde formation	QTvg and QTvl	1,400	QTvg, QTvl, QTvg, QTvl, QTvg, QTvl	Lenticular alluvial fan deposits, QTvg, interbedded with lake deposits, QTvl. The alluvial fans consist chiefly of boulder gravel and cobble gravel, but contain minor amounts of sand and silt. The lake deposits comprise intercalated buff marls and limy silts and white fine-grained tough limestone. A few thin basaltic flows are intercalated in the lake deposits
	Pliocene (?)	Hickey formation	Ths and Thv	1,200	Ths, Thv, Ths, Thv, Ths, Thv, Ths	Interbedded basaltic volcanic rocks and gravel. The volcanic rocks comprise chiefly olivine basalt and lesser amounts of olivine andesite, agglomerate, and basaltic sedimentary rock. The gravel beds consist of boulders and cobbles derived chiefly from local bedrock. Locally the gravel is crudely bedded and cemented by lime
PENNSYLVANIAN AND PERMIAN		Supai formation	PIPs	368±		Red beds comprising chiefly red sandstone and siltstone cemented by lime and subordinate amounts of light-colored limestone and red shale. Locally a zone of breccia and rubble occurs at the base of the formation
MISSISSIPPIAN		Redwall limestone	Mr	255–286		A basal lenticular unit of slabby arenaceous limestone overlain by a massive cliff-forming bluish limestone unit, which is overlain by a light-colored coarsely crystalline unit characterized by abundant conspicuous light-colored chert. An upper unit of white coarsely crystalline crinoidal limestone
DEVONIAN	*Middle(?) and Upper Devonian*	Martin limestone	Dm	440–465		Impure dolomite and dolomitic limestone characterized by diverse lithology and by thin interbeds of shale and mudstone. Divisible into four distinct units
CAMBRIAN		Tapeats sandstone(?)	Ct	0–100		An upper unit of red to yellow limy siltstone and marl grading downward into red well-bedded coarse sandstone and pebble conglomerate separated by thin siltstone partings

FIGURE 2.—Stratigraphic column of Paleozoic, Tertiary, and Quaternary rocks in the Jerome area.

Thickness and stratigraphic relationship

The thickness of the Tapeats sandstone(?) ranges from a few inches to about 100 feet, and averages about 40 to 50 feet. Resistant Precambrian hills that rise into the overlying Martin limestone caused the uneven surface upon which the Tapeats(?) was deposited, and resulted in the great variation in thickness.

The Martin limestone of Devonian age overlies the Tapeats sandstone(?). The contact between them was arbitrarily placed at the base of the first massive limestone bed, which overlies yellowish to greenish limy siltstone and marl.

Lithology

Wherever the Tapeats(?) is well exposed, it consists of two lithologic units: a lower unit, which generally comprises two-thirds of the total thickness, consisting chiefly of sandstone, and an upper unit consisting chiefly of limy siltstone or marl. Both units are generally present, but locally the lower unit wedges out against low hills on the old Precambrian surface. In a few places where erosion has removed the overlying Martin limestone but not the Tapeats(?), only the lower part of the Tapeats(?) is present, for the soft upper unit offers little resistance to erosion.

The lithology of the lower unit consists of medium- to coarse-grained sandstone, granule sandstone, a few pebble conglomerate beds, and minor amounts of siltstone. These rocks are in beds that range from about 1 inch to 10 or more feet. The color generally ranges from dusky red to dark-reddish brown, but the most ferruginous beds are very dusky red. Local zones are variegated, presumably owing to local leaching of iron. Crossbedding, cut and fill channels, and ripple marks are widespread throughout the lower unit.

In some beds sorting is poor but in others it is uniform. In the former, the grain size ranges from medium-grained to granule or locally to pebble size. Crossbedding and cut and fill channels are common in these poorly sorted beds, in which thin irregular-shaped zones that are well sorted may be intercalated. Commonly adjacent to these ill-sorted beds is a zone of well-sorted, even-bedded, medium-grained sandstone. Such zones are succeeded by poorly sorted, crossbedded facies. The entire lower unit of the Tapeats(?) comprises such successions.

Thickness of individual beds appears to be controlled by abrupt changes in grain size and in amount of ferruginous cement, and by change from well to poorly sorted rock. Siltstone and perhaps claystone partings are the most common cause of bedding. In a few places, beds are marked by an abundance of powdery ferruginous cement, and the rock can be crushed with the hand.

Quartz, chert, and jasper constitute nearly all the mineral grains in the lower unit of the Tapeats(?). The minor constituents are feldspar and lithic fragments that are dense, hard, and siliceous. In part at least, these fragments are rhyolitic. The larger sand grains and pebbles in the conglomerate appear rounded.

The upper unit of the Tapeats sandstone(?), generally 15–20 feet thick, comprises siltstone, claystone, marl, and locally from one to three thin sandstone beds near the top. Locally the upper unit appears somewhat fissile, resembling shale. Bedding is crude; it consists of discontinuous irregular planes. Siltstone directly overlying the lower unit is generally dark reddish brown. These beds grade upward into rocks of light-colored hues, yellowish gray, pale olive, and dusky yellow. Color changes abruptly along and across the strike.

The thin sandstone beds, intercalated in the upper 15 feet of the unit, are medium to coarse grained and a light cream in color. These beds rarely crop out so that their extent is not known, but their presence did influence the decision to place these siltstone and marl beds in the Tapeats standstone(?) rather than in the overlying Martin limestone.

Age and correlation

What is the age of the rocks herein called Tapeats(?) and what rocks in adjacent areas are stratigraphically equivalent to them? In the past there has been no consensus of opinion of the age of the Tapeats sandstone(?) in the Jerome area. Reber (1922), Ransome (1932), Stoyanow (1936), and McKee (1951) believed that it is equivalent to the Lower Cambrian Tapeats in the Grand Canyon section. In opposition, Fearing (1926) and McNair (1951) suggest that these beds are the basal part of the Devonian Martin limestone. We believe that the admittedly weak evidence favors a tentative correlation with the Tapeats sandstone. A sample of the siltstone from the upper unit was examined by W. H. Hass of the Geological Survey for conodonts, but none was found. The lack of diagnostic fossils precludes a positive age designation.

The Tapeats(?) in the Jerome area resembles the Tapeats in the Grand Canyon area more than the basal part of the Martin limestone. McKee (1945, p. 16) writes that the upper part of the Tapeats sandstone in all but the eastern part of the Grand Canyon consists of alternating sandstone and shale beds, which is like the rock in the Jerome area. In contrast, shale, siltstone, or marls have not been described from the clastic arenaceous zone at the base of the Martin limestone. The basal part of the Martin limestone is from 10 to 20 feet thick (Huddle and Dobrovolny, 1945), whereas the Tapeats(?) in the Jerome area is as much as 100 feet thick.

McKee (oral communication) has traced fossiliferous (Cambrian trilobites) Tapeats sandstone from the Grand Canyon to a point south of Peach Springs, Ariz. From the Jerome area to Peach Springs, 70 miles northwest of Jerome, outcrops of lithologic similarity occur sporadically, so that correlation based on lithology is tenable.

In the Jerome area, the lithologic gradation and conformity between lowermost Tapeats(?) and the basal Martin limestone argues for a Devonian age for the Tapeats sandstone(?). In these beds the lithology changes from coarse-grained sandstone, through limy siltstone and marl with minor amounts of intercalated sandstone into impure limestone. This gradation and the apparent conformity fail to support the concept of an age hiatus between the two formations. During the assumed hiatus from Lower Cambrian to Middle or Upper Devonian, the upper part of the Tapeats(?) could not have been above wave base, for the friable upper unit would be susceptible to erosion.

Stoyanow (1926; 1936, p. 499–500; 1942, p. 1268) found arthrodiran fish remains in a sandstone at the base of the Devonian in the East Verde River, about 70 miles southeast of Jerome. The sandstone ranges from 50 to 75 feet in thickness and is similar to the Tapeats(?), according to Stoyanow. Fearing (1926, p. 759) correlates the Tapeats(?) of the Jerome area with the Devonian sandstone of the East Verde River. He believed that the apparent conformity between the Tapeats sandstone(?) and the Martin limestone in the Jerome area necessitated this correlation. Stoyanow (1936, p. 497), however, placed the Tapeats(?) of the Jerome area in the Cambrian. Stoyanow (1936, p. 497, 499) believes that beds correlative to the basal Martin limestone in the Jerome area underlie the sandstone in the headwaters of the East Verde River that contains the arthrodiran fish remains.

McNair (1951), who studied 11 Paleozoic stratigraphic sections in northwestern Arizona, traced the Tapeats standstone from the Grand Canyon area southward to Simmons, about 20 miles northwest of the Jerome area. McNair believes that between Simmons and the Jerome area the Tapeats sandstone is overlapped by the Martin limestone and that the Tapeats sandstone(?) of the Jerome area is basal Martin. McNair cites the abrupt overlap of the Cambrian system between Fort Rock, Ariz., and Simmons in support of his thesis. He does not include any shale in the upper part of the Tapeats sandstone; but where he measured sections, the Bright Angel shale overlies the Tapeats sandstone in all but one locality. McNair considered that all shale overlying the coarse sandstone of the Tapeats belonged in the Bright Angel shale. Similar treatment in the Jerome area would place the lower sandstone and pebble conglomerate unit in the Tapeats and the thin upper marl and siltstone unit in the Bright Angel shale. In the absence of diagnostic fossils in the Jerome area, our interpretation is preferred to that of McNair.

MARTIN LIMESTONE

Distribution

The Martin limestone is limited chiefly to the north-central (Mingus Mountain) part of the Jerome area. In the Mingus Mountain area, which covers about 150 square miles, younger rocks conceal the Martin limestone in all but the peripheral zone, and here the platformlike summit area falls away into dissected terrane on the east, south, and west. Northward this summit area joins the Colorado Plateau and only younger rocks crop out.

Small erosional remnants of Martin limestone also occur south of the Mingus Mountain area in the Cherry Creek district, and faulted segments of the Martin occurs in patches east of the Verde fault from the northeast corner of the Jerome area to a point 15,000 feet south of Black Canyon. Most of the Martin east of the Verde fault, however, is north of 1,348,000. Erosion of the overlying Verde formation locally exhumed the small patches of the Martin and the other Paleozoic rocks along the Verde fault.

Thickness

The thickness of the Martin limestone near Haynes gulch 1,366,600 N.; 434,000 E.), north of Jerome, is 441 feet; near Little Coyote Canyon on the western flank of Mingus Mountain (1,353,000 N.; 413,000 E.) it is 465 feet. Stoyanow (1936, p. 497) measured a section 505 feet thick at Jerome. In the headwaters of the East Verde River, about 70 miles southeast of the Jerome area, Huddle and Dobrovolny (1945) measured 390 feet of the Martin. At Simmons, about 20 miles northwest of the Jerome area, McNair (1951, p. 518) measured 381½ feet of the Martin.

Lithology

Diverse varieties of limestone and much contamination by silt and clay characterize the Martin limestone, not only in the Jerome area, but also in southeastern Arizona. The lithology of the Martin is so obviously different from that of the overlying Redwall limestone that the two can be distinguished at a glance. Certain beds in the Martin contain abundant invertebrate fossils, and Stoyanow (1936, p. 495–500) has listed and located the fauna in the section near Jerome.

The Martin limestone comprises four units. These units were distinguished only in measured sections, because detailed mapping of the Martin was not necessary

to meet our objectives. For convenience of description, however, each of the units will be described individually. Listed in order of decreasing age, they are: lower unit, lithographic unit, middle unit, and upper unit.

All of the Martin, and particularly the lower unit, is dolomitic. The coarser the grain size, the more dolomitic is the rock. The dolomite is in large grains set in a fine-grained matrix of calcite.

In two places where hills on the old erosinal surface prevented deposition of the Tapeats sandstone(?), thin lenticular masses of sugary-grained, somewhat vuggy quartz separates the Martin limestone from the older rocks. These two places are south of Jerome: one on the west bank of the north fork of Mescal Gulch (1,355,000 N.; 435,500 E.) and the other on the south wall of Black Canyon (1,327,000 N.; 444,000 E.). The quartz appears to be a replacement of the Martin, but its significance is unknown.

The lower unit ranges from 34 to 53 feet in thickness and is nearly uniform lithologically. A few thin beds, however, contrast in color, texture, and bedding. The lower unit is dominantly light brownish gray, and subordinately yellowish gray. Color depends partly on grain size and partly on the amount of impurities; the finer grained rocks generally are of lighter hues, and the impurities, presumably clay, locally color irregular zones and beds pinkish and yellowish. Medium- to coarse-grained beds are most abundant, but some are fine, and others, coarse. Beds range in thickness from ⅛ inch to 5 feet. The clastic origin of some beds is shown by etched out crossbedding and channeling. Other beds are massive; whereas a few are distinctly fragmental. The massive beds are always coarse grained, perhaps owing to crystallization.

In the lower unit, all the coarse-grained thick beds, and some thin ones are dolomite in which dolomitic limestone forms thin interbeds. The coarse-grained dolomite commonly is spotted with "augen" of coarsely crystalline white calcite, the origin of which is obscure. Conspicuous light-colored chert is nodules and lenses marks the uppermost 2 feet of the unit.

Interrupting the sombre gray sequence of the lower unit, is a series of light-colored, thin-bedded, dense fine-grained limestone beds from 80 to 90 feet thick, aptly called the lithographic unit. These rocks are pinkish gray on fresh fracture and white to very light gray on weathered surfaces. A few beds are slightly darker, and in Haynes Gulch, two beds near the base are dark gray. Beds range in thickness from 3 inches to about 3 feet, and are separated by shaly or silty partings, from a thin film to 3 inches thick, which make the individual beds conspicuous. Most beds are massive but a few show flat or gently undulating internal bedding. Disseminated grains of sand stand in relief on some weathered surfaces. The chief impurity is yellow, gray, and black chert, which occurs in concentric nodules, lenses, and thin layers parallel to bedding; no reason for its localization is apparent.

Near the top of the lithographic unit is a distinctive pale-red sandy zone ranging in thickness from a few to 10 feet. Locally the beds are wholly sandstone, elsewhere they are sandy limestone. This zone is an excellent key bed, locally known as the red marker bed (fig. 27). It has been used to determine offsets of small faults.

The middle unit ranges from 65 to 78 feet in thickness. Its lithology is conspicuously uniform, but abrupt changes from fine- to medium-grain size distinctly mottle the rock and give the false impression of heterogeneity. The mottling appears to be due to recrystallization.

The beds range in thickness from less than 1 inch to 4 feet. Individual beds are characteristically not laminated but locally there are exceptions. The middle unit is dolomitic, and although the ratio of dolomite to calcite is unknown, dolomite probably predominates.

Diverse lithology distinguishes the upper unit, which comprises 250 feet of beds of a total of 440–465 feet for the Martin limestone. It consists of alternating zones, the smallest of which are single beds, perhaps 1 foot thick, whereas the thickest is about 50 feet. For detailed mapping, the upper unit could be further subdivided, and the uppermost 50 feet might be mapped over the entire Jerome area.

The lithology of the upper unit varies in texture, structure, color, and composition; it includes all the types found in the underlying units. It consists chiefly of dolomitic limestone and probably some dolomite, with thin interbeds of limy siltstone or shale. A few pure limestone beds may be sandwiched in the dolomitic rocks, but they were not recognized.

The lower 60 feet of the unit comprises interbedded (1) light-gray, very fine grained, dense dolomitic limestone, (2) light-olive gray, medium- to coarse-grained dolomitic limestone or dolomite, and (3) very light gray, thin-bedded, finely laminated dolomitic limestone. The overlying beds are generally similar to those just described, except for clay and silt impurities and interbeds of claystone or siltstone. Clay mottles the dolomitic limestone purplish to reddish and commonly purple siltstone or claystone interbeds separate these impure beds. The zone from 100 to 140 feet above the base of the upper unit comprises alternating thin-bedded impure slabby dolomitic limestone and poorly exposed purplish limy siltstone that crop out between cliffs of massive dolomitic limestone. These impurities re-

appear sporadically in lesser amounts in the overlying beds. The uppermost 50 feet of the unit consists of grayish-orange-pink, medium-grained dolomitic limestone beds; it is uniformly and evenly bedded, and crops out in uniform steplike ledges.

Two similar key or marker beds occur from 40 to 42 and from 55 to 57 feet above the base of the upper unit. These beds are light-olive gray, medium- to coarse-grained dolomitic limestone. But the distinguishing feature is many inclusions of fine-grained, moderate red sandstone in irregular-shaped masses from a fraction of an inch to 9 inches long. Another excellent, but local, marker bed, distinguished by corals, occurs from 150 to 157 feet above the base of the upper unit. Locally the abundance of both solitary and colonial types indicates a reef, but elsewhere, the bed was not recognized.

Age and correlation

Middle and Upper Devonian sedimentary rocks, chiefly limestone and dolomitic limestone, are widespread in Arizona. McKee (1951, pl. 2) has shown that Devonian seas probably covered all the State except for the Defiance uplift, a small area in notheastern Arizona.

The Martin limestone crops out in several mountain ranges northward from the Bisbee area in southeastern Arizona, where it was first described by Ransome (1904, p. 33), to the Globe–Miami area. Huddle and Dobrovolny (1945) traced the Martin from near Globe to the headwaters of the East Verde River, southeast of Jerome. They state that the Martin probably can be traced into the Temple Butte limestone in the Grand Canyon area. There is little doubt that the Martin described by Huddle and Dobrovolny at the East Verde River is the same as the Devonian system in the Jerome area, and we believe the name Martin for the limestone in the Jerome area is justifiably applied. This correlation already has been made by McNair (1951, p. 516) who designated the Devonian system from Jerome to the western part of the Grand Canyon at Hurricane Cliffs as the Martin.

In contrast to Huddle and Dobrovolny and to McNair, Stoyanow (1936, p. 495) placed the northernmost limit of the Martin limestone near Roosevelt Reservoir. Stoyanow believes a land mass—Mazatzal land—existed during the Devonian in the area between Roosevelt Reservoir and the Pine–Payson area. The Devonian seas north of Mazatzal land were not connected to the Martin sea to the south, according to Stoyanow. He cites lithologic and paleontologic differences to support his thesis. Stoyanow (1936, p. 499) proposed the name Jerome formation for the Devonian system in the Pine–Payson and Jerome areas, and correlated the Martin limestone with only the upper part of his Jerome formation.

Huddle and Dobrovolny (1945) point out that Devonian rocks were deposited in the area Stoyanow called Mazatzal land, although in places the strata are only 30 feet thick. They maintain that the designation of the Devonian system in the Pine–Payson area as the Martin limestone is justified by general lithologic similarities and by the indications of original continuity with the Martin in the Globe area.

Huddle and Dobrovolny (1952, p. 67) assign a Late Devonian age to the Martin limestone. They state (1952, p. 86) that their admittedly inadequate collections from the Martin may contain equivalents of some of the faunal zones of the Devils Gate formation in central Nevada (Merriam, 1940) and indicate a somewhat older age for the lower part of the Martin. Their middle member of the Martin contains abundant stromatoporoids and corals, and they believe that their lower and middle members may represent part or all of the *Stromatopora* zone of the Devils Gate formation and thus be Middle Devonian. The *Phillipsastrea* (*Pachyphyllum*) zone (Late Devonian) of the Devils Gate formation (Huddle and Dobrovolny, 1952, p. 86) is probably present in their upper member of the Martin, and they conclude that the available faunal evidence indicates that the uppermost beds of the Martin are Late Devonian.

REDWALL LIMESTONE

Distribution, thickness, and stratigraphic relationship

The Redwall limestone, like the Martin, crops out only in the peripheral zone of Mingus Mountain, for the lava flows of the Tertiary Hickey formation conceal all but the edges of the older rocks. Southeast of Mingus Mountain (1,322,500 N.; 455,800 E.) there is also a small outlier of Redwall resting on Martin. Several areas of Redwall occur east of the Verde fault northward from Jerome, and from Jerome south to near the Green Monster mine (1,348,500 N.), several small inliers of Redwall occur in the Verde formation.

Near Haynes Gulch (pl. 1, 1,366,600 N.; 434,000 E.), the Redwall limestone is 286 feet thick; on the west side of Mingus Mountain, near Little Coyote Canyon (1,353,000 N., 413,000 E.) it is 256 feet thick, but here erosion has removed an unknown but presumably small part of the formation. About 1½ miles north of Little Coyote Canyon, Gutschick (1943) measured an incomplete section of the Redwall 270–275 feet thick, and about 50 miles northwest of Jerome in Chino Valley, McNair (1951, p. 519) measured 344½ feet of the Redwall, which indicates a general thickening of the Redwall northwestward.

The Redwall limestone, of early Mississippian age, disconformably overlies the Devonian Martin lime-

stone, and the Supai formation disconformably overlies the Redwall. The surface of the Martin formation upon which the Redwall was deposited was somewhat irregular, for 30 feet or more of basal Redwall is cut out between Jerome on the eastern flank of Mingus Mountain and Little Coyote Canyon on the western flank.

Erosion has stripped the Supai from most of the Mingus Mountain area underlain by Redwall, so that Tertiary lava and gravel of the Hickey formation, in part, rest directly on the Redwall. The central part of the northern boundary of the Mingus Mountain quadrangle approximately marks the erosional edge of the Supai formation. The best exposures of the Redwall-Supai contact are along the road in the Clarkdale quadrangle to Perkinsville; they show at least 6 feet of coarse limestone rubble at the base. In other places, the basal breccia contains detrital chert as well as fragments of limestone. In Sycamore Canyon, 9 miles north-northeast of Jerome, McKee (oral communication) found 3 feet of conglomerate overlying the Redwall. Jackson,[1] who studied the Supai formation in the Fort Apache area southeast of the Jerome area, found that the erosional unconformity at the top of the Redwall had from 15 to 20 feet of relief; and northeast of the Jerome area in Chino Valley, Hughes (1952, p. 640) found thick lenses of basal conglomerate of the Supai resting on the Redwall. There is little doubt that a significant break separates the Redwall from the Supai, which is further emphasized by the lithologic differences.

Lithology

The Redwall limestone is typical of the Mississippian crystalline pure limestone formations in Arizona. From a distance, it appears white, but on close inspection is seen to be light gray and pink. Most of the rock is distinctly granular and coarse grained; only thin local zones are fine grained. Abundant fossiliferous beds alternate with beds of fossil-poor, crystalline limestone; a few beds appear almost wholly crinoidal. Road cuts and cliffed stream canyons reveal channels and small caves marked by brick-red claylike sediment, presumably residual from solution of the limestone. From a distance, the trace of these red solution openings appear as a haphazard, red-on-white pattern. Parts of the Redwall are so massive that they stand as spectacular white cliffs. These cliffs, owing to the near horizontal attitude of the Redwall, contour the rugged topography. Without doubt, the Redwall limestone is a handsome rock, boldly displayed.

[1] Jackson, R. L., 1950, The stratigraphy of the Supai formation along the Mogollan Rim, Central Arizona: Unpublished thesis, Univ. Ariz., Tucson, Ariz.

Gutschick (1943) studied the Redwall limestone in the area from Black Mesa in the northwest end of Chino Valley southeastward to Jerome, and subdivided it into four members. We recognize only three members, the lower two of which coincide with Gutschick's. These units were not mapped separately except locally where needed to decipher the structure; none are shown on the maps. For descriptive purposes, however, we recognize the three members.

The lower member of the Redwall consists of two distinct parts: a lower unit, which locally wedges out, of slabby limestone, sandy limestone, and limy sandstone, and an upper unit of massive, thick-bedded, oolitic limestone. On the eastern flank of Mingus Mountain, the lower unit is about 35 feet thick; on the southern flank it is much thinner; and on the western flank, it is missing, presumably being cut out against a "high" on the erosional surface of the Martin. The lower beds are light gray, and some have a light-purplish cast resulting from impurities. They are distinctly slabby, and range in thickness from about 1 inch to 2 feet. Commonly the bedding planes are wavy, irregular, and discontinuous; some beds thin and lens out, whereas others end unexpectedly. These beds are sandy and contain significant amounts of dolomite, and appear to be reworked Martin limestone, a belief supported by a local thin basal conglomerate composed of fragments of the Martin.

The upper unit of the lower member is a massive, thick-bedded, light-blue to gray, oolitic limestone, 50–55 feet thick, that generally forms spectacular cliffs. In places the cliffs rise vertically for 50 feet, and elsewhere the oolitic limestone combines with the overlying beds to form still higher vertical drops. In most places the rock is a mass of conspicuous oolites, but locally it has recrystallized into a coarse-grained, granular rock.

The oolitic limestone passes conformably upward into the middle member, which ranges from 55 to 74 feet in thickness. Chert, in abundant pods, lenses, layers, and irregular-shaped masses, is diagnostic of the middle unit. The limestone is thick bedded (2–5 feet), and medium to coarse grained; where medium grained, it is nearly white, and where coarse grained is light cream. The chert generally ranges in color from white to yellow, but some is dark gray or, uncommonly, nearly black. The middle member is fossiliferous; solitary corals are most conspicuous.

The upper member, about 140–150 feet thick, is a yellowish-gray, massive, thick-bedded crystalline limestone. It is generally coarse grained, but in local areas it is medium to fine grained. It abounds in fossils; some beds appear to be derived entirely from crinoid remains. Gutschick (1943, p. 7) found fossil brachio-

pods, crinoids, corals, bryozoa, blastoids, gastropods, trilobites, and fishes.

Age and correlation

The Mississippian limestone in northern Arizona was called the Redwall by Gilbert (1875, p. 162, 177–186), who included rocks older and younger than the Mississippian. Noble (1922, p. 26, 54), as a result of studies in the Grand Canyon, restricted the name Redwall to strata of Mississippian age. The Mississippian limestone in the Bisbee district was named Escabrosa by Ransome (1904, p. 42–44), and this term has been widely used in southeastern Arizona. Huddle and Dobrovolny (1952, p. 86) traced the two formations into each other, and used Redwall limestone in their report dealing with central Arizona. The term "Redwall" has been used earlier in the Jerome area by Ransome (1916, p. 162), Lindgren (1926, p. 9), Stoyanow (1936, p. 512–514), and Gutschick (1943). Fossils collected by Ransome (1916, p. 162) and Wooddell (Stoyanow, 1936, p. 514) in the Jerome area, indicate an early Mississippian age, corresponding to parts of the Kinderhook and Osage groups.

SUPAI FORMATION

Distribution

The Supai formation crops out in the northern part of the Jerome area, the southern limit of the formation nearly coinciding with the southern boundary of the Clarkdale quadrangle (pl. 1). The Supai underlies the north margin of Woodchute Mountain and part of the eastward-dipping slopes that extend northeastward from Woodchute Mountain to the Verde Valley. Volcanic rocks of the Tertiary Hickey formation cover much of the Supai formation, especially on Woodchute Mountain.

The Supai underlies the entire northeast corner of Arizona. The south edge of the Supai extends from about the center of the eastern boundary to the approximate northwest corner of the State. The Jerome area is about in the center of this line marking the southernmost extent of the formation.

Thickness and stratigraphic relationship

In the Jerome area, a maximum of only 370 feet of the basal part of the Supai formation remains; this thickness represents the southernmost beveled edge of Supai, which erosion is forcing to retreat northward. From the Jerome area northward to the Mogollon Rim, for 10 to 12 miles, only partial sections of Supai are present. In the escarpment of the Mogollon Rim, however, the overlying Coconino sandstone assures a complete section. Here McKee (oral communication) measured 1,665 feet of Supai formation in Sycamore Canyon which includes 332 feet of Pennsylvanian rocks at the base. Northwest of the Jerome area near Black Mesa on the east side of Chino Valley about 10 miles south of Ashfork, Ariz., Hughes (1949, p. 33) measured 1,155 feet of Supai. About 70 miles southeast of the Jerome area on Fossil Creek, which drains off the Mogollon Rim into the Verde River, Huddle and Dobrovolny (1945) measured 2,200 feet of Supai from which they reassigned the basal 470 feet as the Pennsylvanian Naco formation.

The disconformable relationship of the Supai to the underlying Redwall limestone has been described in the section on the Redwall. In the Jerome area, the Supai has been beveled and covered by the Tertiary Hickey formation.

Lithology

The Supai is a red-bed formation, consisting chiefly of sandstone and siltstone, but it contains subordinate amounts of limestone and shale. In the Jerome area, the color ranges from moderate red to pale reddish brown. The Supai is well bedded, and forms a cliff and slope topography. Some beds are slabby, and others distinctly laminated. Many of the sandstone and siltstone beds are crossbedded, and crossbedded epiclastic limestone has been reported from the Supai. Noble (1922, p. 60) found that the inclined laminae in the crossbedding dipped southward, and McKee (1940, p. 882), who made an exhaustive study of structures in the Supai, found that the laminae inclined south and southeast, indicating the source of the sediment lay to the north. Hughes (1952, p. 648) suggested that sweeping cross lamination and mud cracks indicate the Supai is deltaic.

In the Jerome area, the Supai comprises a basal zone of rubble or breccia overlain by an impure limestone of uncertain origin. Above the limestone are well-bedded, moderately red siltstone beds, a few thin beds of limestone, and possibly some claystone or shale. The basal zone is commonly poorly exposed, and probably varies in thickness and character from place to place. It is most prominent north of Jerome near the road to Perkinsville. Here it consists of 6 feet of rubble composed of blocks of limestone, ranging from about 1 inch to 3 feet in diameter, cemented by moderately red siltstone. On the hills above Haynes Gulch west of Jerome and also about 12,000 feet north of the United Verde pit, a most distinctive pebble bed, 3–20 feet thick, occurs about 20–30 feet above the base of the Supai. It consists of pebbles of chert, limestone, and siltstone in a dark-maroon limy and sandy matrix. Overlying the basal rubble bed is a siltstone and limestone zone, which north of the road to Perkinsville, consists of about 50 feet of impure gray limestone and lesser amounts of intercalated red siltstone.

Along the Perkinsville road about 45 feet above the base of the Supai, a 5-foot limestone-breccia bed occurs in which the fragments are as much as 4 inches in diameter. Above the breccia bed is a zone about 70 feet thick composed of unsorted sedimentary detritus derived mostly from limestone. Patches of deep red siltstone are also abundant. Above the head of Haynes Gulch west of Jerome, this zone is only 20–40 feet thick, and though poorly exposed, consists of gray limestone, partly to completely replaced by chert. Chert nodules and lenses form a small part of these limestone beds in other places, but nowhere else does it almost completely replace entire beds. In all exposures, this sedimentary chaos of limestone, siltstone, and breccia passes abruptly into siltstone, the characteristic facies of the Supai.

The siltstone facies forms a cliff and slope topography, which is somewhat spectacular because of the bright red hue of the rocks. A maximum thickness of about 250 feet of these rocks remains in the Jerome area. They are lithologically similar throughout, except for one or two thin beds of gray limestone. The variations are in degree rather than kind. Siltstone is the most abundant rock type, but fine-grained sandstone, minor beds of limestone, and perhaps claystone constitute a small part of the section. The siltstone consists of small quartz grains, calcareous cement, and iron oxide. The iron oxide occurs chiefly as a cement and pigment. The relative amounts of these materials vary from bed to bed; it was often debatable whether a rock was a calcareous siltstone or a silty limestone. Bedding ranges in thickness from about 1 inch to as much as 4 feet. The thin slabby beds are generally friable and commonly are a deeper red than the thick beds, perhaps because pulverulent iron oxide comprises a greater part of the matrix. These beds form the slopes where outcrops are sparse. The thicker beds are generally in cliffs and ledges and are lighter red and limy. Crossbedding marks many of the cliff-forming beds. The sweep of the inclined laminae ranges from 1 or 2 inches to perhaps 3 feet. In summary, the features that characterize the Supai in the Jerome area are the prevailing red hue, the cliff and slope topography, the uniform grain size of the siltstone, and the crossbedding.

Sporadically certain beds contain minor structures for which we have no ready explanation and no attempt was made to investigate them. Chief among these were a "pimply" structure, a pseudobreccia, and an odd mottling resulting from variations in texture and composition. The "pimply" structure, a product of weathering, appears as small raised surfaces liberally scattered on the surface of the rock. The pseudobreccia appears on the bedding planes as many, irregular-shaped, light-pinkish-gray areas, some of which are slightly raised, set in a matrix of the normal red siltstone. Perpendicular to the bedding planes, however, they are tubular or branching, like the trunk and main limbs of a tree. The only apparent lithologic difference between these odd-shaped masses and their matrix is in the relative abundance of iron oxide. The mottled rock appears to be a haphazard mixture of small irregular-shaped masses of light-pinkish-gray impure limestone and of red calcareous siltstone. There is not the slightest tendency for these two rock types to be segregated.

Age and correlation

The Supai formation was originally defined by Darton (1910, p. 25–27) from exposures in northern Arizona. It included about 800 feet of red sandstone and shale beds that underlie the Coconino sandstone and overlie the Redwall limestone. Later, Noble (1922, p. 64) redefined the Supai formation by reassigning from the top about 300 feet of red sandy shale and fine-grained sandstone, which he named the Hermit shale, and by adding to the base about 250 feet of sandy shale, limestone, and calcareous sandstone, which formerly had been included in the Redwall limestone. Noble redefined some of the Permian beds from the top of the Supai, and added Pennsylvanian(?) rocks to the base. As thus defined, the Supai is of Permian and Pennsylvanian(?) age.

In recent years, there has been a tendency among some to reassign the Pennsylvanian beds from the Supai. But the lack of an unconformity and the lithologic similarity of the Pennsylvanian and the Permian beds and the scarcity of diagnostic fossils render the wisdom of this division questionable. Huddle and Dobrovolny (1945) in their study of the Paleozoic stratigraphy of central and northeastern Arizona redefined the Pennsylvanian rocks from the Supai and called them the Naco formation.

Ransome (1904, p. 44) originally described the Naco formation from the Bisbee area, Arizona; following Girty, he indicated the Naco to be of Upper Carboniferous age and to contain a Pennsylvanian fauna. Subsequently, Huddle and Dobrovolny (1945) recognized the Naco formation in the Globe-Miami area, Arizona. From there, they extended the limits northward into the Fossil Creek-East Verde River area of central Arizona, which is about 70 miles southeast of the Jerome area. Following Huddle and Dobrovolny, Jackson[2] and S. S. Winters, whose stratigraphic work in the Fort Apache area of Arizona has not been published but is quoted by Jackson, reassigned the Pennsylvanian beds from the Supai and called them Naco formation.

[2] Jackson, R. L., op. cit.

The lithologic similarity of the Permian and Pennsylvanian beds influenced these workers to choose some arbitrary plane for the contact between the Permian and Pennsylvanian beds placing it above the highest bed in the section that contained a marine Pennsylvanian fauna. This treatment resulted in Winters' placing the Naco-Supai contact about 570 feet higher in the section than Jackson, for Winters found upper Pennsylvanian fossils (Virgil age), but Jackson found only middle Pennsylvanian fossils (Des Moines age).

Hughes (1952, p. 656) did not differentiate the basal fossiliferous limestone facies from the Supai formation as exposed in Black Mesa, 30 miles to the northwest of the Jerome area. Many fossils collected by Hughes were too fragmental for positive identification or the geologic range of others was too great for precise age determination. Hughes concluded that the available evidence indicated that the basal Supai at Black Mesa is of early Pennsylvanian age, and related to the limestone of the Nevadan geosyncline to the west.

No fossils have been found in the Supai of the Jerome area. Although probably it is stratigraphically equivalent to beds called Naco by Jackson, Winters, and Huddle and Dobrovolny, we believe that too little is known about the lateral continuity of proposed subdivisions and about the stratigraphic relationship of the subdivisions offered by different workers, for the Supai in the Jerome area to be subdivided into other formations. Hence we shall call these rocks the Supai formation of Pennsylvania and Permian age.

UNCONFORMITY AT THE BASE OF THE TERTIARY ROCKS

Much of central Arizona southwest of the Colorado Plateau contains only Precambrian metamorphic and granitic rocks, locally covered by Cenozoic volcanic and sedimentary rocks, as in the southern part of the Jerome area. Several lines of evidence, summarized by McKee (1951, p. 486), indicate that Paleozoic sedimentary rocks formerly covered the central Arizona area. The thickness and lithologic character of the Paleozoic rocks, except the Cambrian, give no indication of deposition on the margins of a basin. The Devonian rocks thicken toward this area from the north and east. The most southerly exposures of Mississippian strata are as free of detrital sediment as to the north, and although they are thinner than in the Grand Canyon section, they thicken toward the Jerome area from the east and reach about 300 feet in thickness in that area. The Permian rocks are 2,000 feet thick at their southernmost margin. Isopach maps prepared by McKee (1951, pls. 1 and 2) clearly indicate that Devonian, Carboniferous, and Permian rocks formerly extended from the Colorado Plateau across the central part of Arizona to the southern and southwestern parts of the State. Lower and Middle(?) Triassic sedimentary rocks of the Moenkopi formation are 300 feet thick at the south edge of the Colorado Plateau, and they may have extended for an unknown distance into the Jerome area.

McKee (1951, p. 494) has also summarized the evidence for extensive uplift in the central Arizona region in Late Triassic time. The Shinarump conglomerate of Late Triassic age was deposited as a broad sheet of gravel throughout much of northeastern Arizona and adjacent areas. The large size and the high degree of rounding of the detritus in the gravel indicate vigorous erosion and long transportation from the source area. Among the pebbles in the Shinarump conglomerate McKee (1937) found distinctive lithologic types and diagnostic fossils that prove a source to the south for much of the sediment.

Except for the possible deposition of a thin sheet of Moenkopi the regional evidence indicates that during Triassic, Jurassic, and Cretaceous time (McKee, 1951, pls. 2 and 3) the Jerome area was eroded. The lack of early Tertiary deposits indicates that erosion probably continued until the accumulation of the Hickey formation in the late Tertiary. During this period 3,000 to 3,500 feet of Paleozoic sedimentary rocks and an unknown thickness of Precambrian rocks were eroded.

Sometimes during this period of active erosion, the Paleozoic rocks in the Jerome area were tilted northward a few degrees so that the Precambrian rocks are exposed to the south of the Paleozoic. There is no evidence to date the period of tilting: (1) it may have been at the time of the Late Triassic uplift, or (2) during the Late Cretaceous and early Tertiary period of faulting discussed on page 78) or (3) as a result of recurrent uplift at different periods.

The erosional surface on which the Hickey formation accumulated had a relief of 500 feet or more, indicating perhaps, some regional uplift just before the time of accumulation of the Hickey. This might favor the possibility of recurrent uplift during the Mesozoic and Tertiary period of erosion.

The evidence seems compelling that the unconformity in the Jerome area separating the Supai and older rocks from the late Tertiary Hickey formation is a major unconformity in central Arizona, and represents a long period of erosion.

CENOZOIC ROCKS

In the Jerome area, Cenozoic rocks comprise the Hickey formation of probable Pliocene age, the Verde formation and the older Quaternary gravel both of which are younger than the Hickey formation, younger

Quaternary gravel, and Recent riverwash, talus, and terrace deposits. The Hickey and Verde formations consist of lava flows intercalated with sedimentary rocks ranging in size from boulder conglomerate and gravel to clay and mudstone. The younger formations are predominantly unconsolidated gravel, ranging in size from boulder gravel to pebble gravel and coarse sand.

HICKEY FORMATION

Distribution

The Hickey formation here defined is named from mediocre exposures on Hickey Mountain, a subsidiary upland area contiguous on the northwest to Mingus Mountain. The most typical exposures are on Mingus Mountain and logically this area should be the type locality; however, the name Mingus is preoccupied.

In the Mingus Mountain quadrangle, the Hickey formation covers Mingus Mountain and the subsidiary mountains—Hickey and Woodchute—lying northward. This formation occurs in many scattered patches, some of which crown peaks, from the summit area of Mingus Mountain southward to the road between Dewey and Cherry; the largest patch is along Ash Creek and Tex Canyon just north of their junction. A mantle of the Hickey overlies much of the area south of the road between Dewey and Cherry, and farther southward, it increases in extent and thickness, covering fully half the adjoining Mayer quadrangle on the south. Most of the south end of Lonesome Valley is also cut in the Hickey. To the north in Lonesome Valley, small strip-like exposures of the Hickey crop out on the walls of some of the larger gulches, separating the riverwash from the Quaternary gravel that mantles the pediment surfaces.

At one time, the Hickey formation must have covered the Black Hills and at least part of Lonesome Valley. But to what extent the Hickey extended eastward in the area now occupied by the Verde Valley, and to what extent it was once contiguous with the Tertiary lava and gravel of the Colorado Plateau, have not been determined.

Thickness and stratigraphic relationship

Erosion has removed the top of the Hickey formation so that only partial sections occur in the Jerome area. Along the northern boundary of the Mingus Mountain quadrangle, about 1,400 feet of the Hickey, of which all but 50 feet is basalt, occurs at Woodchute Mountain. In the south end of Lonesome Valley about 1½ miles north of the Precambrian bedrock at Humboldt, drilled wells penetrated 735 feet of Hickey formation. In addition, at least 300 feet of strata crops out to the east in the adjacent ridges, and along the southeast margin of Lonesome Valley, remnants of the Hickey formation lie at altitudes 300 and 750 feet above the valley floor. These perched remnants clearly attest to the former presence of a greater thickness of the Hickey, but uncertainties of initial dips and fault offsets preclude determination of the excess above the 1,035 feet now present. Presumably the thickness increases northward from the south end of Lonesome Valley because the surface of the Precambrian rocks beneath the Hickey formation slopes northward, whereas altitudes on the Hickey increase.

The Black Hills were uplifted along the Coyote and Verde faults after the accumulation of the Hickey formation, and the Verde formation was deposited in the newly formed Verde Valley to the east. The unconformity between the Hickey and Verde formations is exposed in the area east of the Verde fault and north 1,350,000 N. (pl. 1).

In Lonesome Valley along the front of the Black Hills, erosion carved a pediment on the Hickey formation, which was masked by a mantle of gravel—the tool of erosion. Subsequently, accelerated erosion produced by capture of the drainage of the south end of Lonesome Valley by the Agua Fria River stripped the gravel mantle from the pediments, and dissected the Hickey formation.

Lithology

The relative abundance of the sedimentary and volcanic facies of the Hickey formation varies. On Mingus Mountain, volcanic rocks compose all but a small part of the formation, but along the south-central margin of the area, gravel and lava flows are intercalated. Here in one place four distinct gravel beds are sandwiched between five flows. In the south end of Lonesome Valley, however, sedimentary rocks predominate, for only one zone of volcanic rocks occurs south of Dewey.

The volcanic facies comprises chiefly flows and subordinate amounts of basaltic sediments and breccia. The breccia beds are probably the blocky, fragmental, and scoriaceous tops of flows, but some may be true pyroclastic rocks. The basaltic sediments comprise material derived entirely from the volcanic rocks; they are well bedded, and are fairly well sorted. In addition in one locality in the south-central part of the area, a thin bed of rhyolitic tuff is intercalated between flows (1,305,000 N.; 436,800 E.).

The thick pile of lava on Mingus Mountain comprises many flows, some 50 feet or more thick. The lava in the southern part of the area, however, apparently comprises all thin flows from 10 to 20 feet thick. Fragmental basalt or basalt breccia occurs intimately associated with flows in all areas.

Basaltic sediments are widespread; they occur on Mingus Mountain, in the southeast corner of the area,

and under the lava flows exposed at Humboldt in the southwest corner of the area. On Mingus Mountain, the roadcuts along Highway 89A near the summit area of Mingus Mountain display well-sorted and bedded material. Small topographic sags or basins on Mingus Mountain, such as the Potato Patch campground, are believed to have formed by sluicing out of soft basaltic sediments by flood action, although all the exposures in these areas are masked by soil. In the southeastern part of the area, Quaternary terraces and stream cuts expose about 40 feet of bedded volcanic sand and fine gravel, composed partly of small red to pink lapilli. Similar material occurs under the lava near Humboldt in the stream canyon that dissects the lava. All these well-bedded basaltic sand and gravel beds probably collected in local depressions during lulls in volcanism.

The physical character and mineralogic composition of the basalt appears typical of many plateau basaltic flows. The interior parts of flows are massive, except for vesicles, whereas the tops of flows are blocky from brecciation. Vesicles, ranging from a pin point to 1 inch in diameter, are abundant, even in dikes and related lenses or pods. Columnar joining is common, and magnificently displayed in the peripheral escarpment around the summit area of Mingus Mountain. The basalt is porphyritic; phenocrysts of olivine and augite are set in a very fine grained or microcrystalline felty (pilotaxitic) groundmass, composed of needles of labradorite in a base of mafic material—presumably augite and olivine. In many places, the olivine was partly altered to iddingsite. Locally the augite appears to be limited to the groundmass, but on Crater Mountain south of Cherry, augite phenocrysts are one-eighth to one-half inch long. Local patches of biotite-bearing lava, probably andesite or latite, were noted east of the Verde fault south of Jerome. Similar rocks may occur elsewhere in the Jerome area, for the lithology of the lava was not checked in detail. Medora H. Krieger differentiated biotite and hornblende andesite in the Tertiary volcanic rocks in the Prescott quadrangle.

The sedimentary facies of the Hickey formation ranges from fine marl or silt to coarse gravel and conglomerate. The coarse sediments accumulated (1) in the area now marked by Mingus, Woodchute, and Hickey Mountains, (2) in the southern part of the Jerome area south of the road between Dewey and Cherry; (3) in the bedrock margins near Humboldt; and (4) presumably in the intervening areas where erosion has removed all but scattered gravel beds. Some of the boulders and cobbles, especially those derived from the Paleozoic limestone, are rounded to subrounded; others, especially those derived from siliceous Precambrian rocks, show little evidence of abrasion. As might be expected in deposits of this type, the degree of sorting varies greatly from bed to bed and from one locality to another. It is not uncommon to see a bed or lens of well-sorted coarse-grained sand intercalated in a heterogeneous aggregate of boulders, cobbles, pebbles, and finer grained sedimentary rocks.

The finer sediments generally accumulated away from the bedrock margins, and are best exposed in the south end of Lonesome Valley. Presumably, local basins were favorable for quiet-water deposition; these basins may have been structural in origin or the result of lava flow barriers in the general drainage pattern.

The fragments composing the gravel were derived from the Precambrian, Paleozoic, and Tertiary rocks—the same type of older rocks that are exposed in the Jerome and adjacent areas. Most of the gravel contains fragments of all three major rock types. But local gravel beds comprise Precambrian fragments to the exclusion of others and conversely. In Gaddes Canyon and on the adjacent ridge to the northeast, the well-consolidated gravel beds in stream channels abound in cobbles and boulders derived from Paleozoic rocks, and have only a few fragments of Precambrian rocks. The gravel beds, however, that underlie the basalt of Kendall Peak originated solely from Precambrian rocks, even though Paleozoic rocks crop out about 2,000 feet northward. Elsewhere in the Mingus Mountain area, the gravel contains abundant fragments of all the older rocks. Of special interest is the presence of basalt fragments in some of the gravel beds below the lowest lava flows on Mingus Mountain.

In places, gravel beds at different levels in the same locality had different sources, as shown by the exposures in Tex Canyon and northeast of Jerome. In Tex Canyon, the older cemented gravel comprises chiefly Paleozoic rocks, but the overlying unconsolidated gravel contains mostly Precambrian rocks. The exposures northeast of Jerome are similar, except that the older gravel contains only fragments of Paleozoic rocks. Thus the drainage shifted from time to time during accumulation of the gravel, though perhaps only on a local scale. In many places, however, the lithology of the gravel reflects the lithology of the adjacent bedrock. This is well displayed in the northwest corner of the Mingus Mountain quadrangle. Here the gravel overlying the thin wedge of Supai contains much of that formation, some in angular blocks several feet in diameter. Similarly, fragments of rocks of the Alder group and others derived from the Grapevine Gulch formation of the Ash Creek group predominate in the gravel in the south end of Lonesome Valley, but other types are present in minor amounts.

In a few places diagnostic rocks in the gravel beds reveal considerable information on source, and hence on the general direction of drainage. Of particular interest is the presence of alaskite and gneissic alaskite fragments in gravel along the east margin of Lonesome Valley from 1 to 2 miles north of the road between Dewey and Cherry and in the gravel farther east at the same latitude. The source of these alaskite fragments is in the southeastern part of the Prescott quadrangle; they are not known to occur in any other place in this vicinity. Drainage thus at one time was from the west or southwest. Similarly, north of Jerome, some of the unconsolidated gravel contains fragments of breccia from the Spud Mountain volcanics and other rock types belonging to the Alder group. Rocks of the Alder group lie only to the west and southwest of where the fragments were found; drainage must have been from that direction. Drainage from the south is also indicated for the gravel that underlies the basalt of Kendall Peak; for, as mentioned earlier, these beds comprise only Precambrian rock types and Paleozoic rocks predominate to the north. Drainage from the north would have yielded abundant detritus derived from Paleozoic rocks.

Age and correlation

Throughout Arizona, Cenozoic rocks are widespread; they consist of (1) sedimentary rocks ranging in size from coarse gravel to clay; (2) extensive basaltic flows; and (3) intercalations of varying amounts of basaltic flows in sedimentary rocks. The Hickey formation is similar lithologically to the other Cenozoic rocks found in Arizona. On Mingus Mountain, basaltic flows are dominant; near Humboldt, sedimentary rocks are dominant; and along the south edge of the Mingus Mountain quadrangle, about equal amounts of lava flows and sedimentary rocks are intercalated. In southeastern Arizona, the widespread Gila conglomerate is dominantly sedimentary, but Gilbert (1875, p. 540) reported that in places, basaltic flows predominate over the conglomerate.

Age assignments for the Gila vary, but generally Pliocene is the favored choice, based chiefly on vertebrate fossils. Pliocene vertebrate fossils were found in Gila conglomerate by Gidley (1922; 1926) in the San Pedro Valley and by Knechtel (1936, p. 87) in the Gila and San Simon Valleys, although Knechtel (1936, p. 86) states that some of the deposits along the upper Gila River might be Pleistocene. A recent discovery of early Pliocene (J. F. Lance, oral communication) vertebrate fossils has been made 20 miles south of Prescott in the Mount Union quadrangle along Milk Creek, in a sequence composed of gravel, sand, clay, rhyolitic tuff, and intercalated dacitic and basaltic flows.

The country near Flagstaff to the north and northeast of the Jerome area contains extensive basaltic flows that cover part of the Colorado Plateau. Robinson (1913) established three stages of eruption, (1) widespread basalt, (2) central eruptions of andesite and rhyolite, (3) basalt from scattered vents, in part Recent in age. On indirect evidence, Robinson concluded that the first-stage basaltic eruptions occurred in late Pliocene time, and this conclusion is widely quoted. However, the facts do not permit such assurance, as they are not based on stratigraphic or fossil data. Childs (1948) on physiographic evidence suggested that the first-stage basaltic eruptions are very late Pliocene or early Pleistocene. In the Hopi Buttes to the east of Flagstaff, vertebrate fossils indicate middle and late Pliocene age (Childs, 1948, p. 378) for a formation consisting of intercalated sedimentary rocks and lava flows. This middle and upper Pliocene formation is truncated by a pediment that Childs traced from the Hopi Buttes to northeast of Flagstaff where Robinson's first-stage basaltic flows are above the pediment. Although this may prove that the first-stage basaltic flows are very late Pliocene or Pleistocene northeast of Flagstaff, it does not necessarily prove that the basaltic flows north of Jerome on the edge of the Colorado Plateau are of the same age.

In the Verde Valley, Mahard (1949, p. 118) reported lava flows intercalated between sedimentary rocks of the Verde formation that are younger than the Hickey formation. The Verde formation is clearly younger than the basaltic flows on the edge of the Colorado Plateau to the north that are believed to be the same age as the Hickey.

Lava flows and associated sedimentary rocks are not all the same age in Arizona, and age assignments to such rocks is difficult owing to the lack of fossil evidence. Unfortunately no fossils were found in the Hickey formation in the Jerome area. In the Prescott quadrangle, in rocks of similar lithology, K. K. Kendall of the U. S. Geological Survey, in 1946, found fragmentary antelope and llamalike camel bones. C. L. Gazin (1947, written communication) of the National Museum studied these fossils and has stated that there is nothing in the material to preclude the possibility of a Pliocene age, but the material is too fragmentary for certainty.

In this report the Hickey is tentatively assigned to the Pliocene epoch. The reasons are as follows: (1) much of the satisfactory dating of comparable Cenozoic rocks in Arizona has been to divisions of the Pliocene epoch; (2) the vertebrate fossils found in the sedimentary rock near Prescott do not preclude the possibility of Pliocene age; (3) the tectonic activity subsequent to the deposition of the Hickey formation followed by the

deposition and much erosion of the Verde formation do not favor a Pleistocene age for the Hickey formation; (4) bias on our part because of the accurate early Pliocene dating of comparable rocks 20 miles south of Prescott.

VERDE FORMATION

Distribution

The Verde formation is a unique sequence of limy lake deposits confined to the upper Verde Valley, an area that covers about 300 square miles and whose maximum dimensions are about 40 miles long by 15 miles wide. The upper Verde Valley is bounded on the east and north by the Mogollon Rim and on the west by the Black Hills, which include the Jerome area. To the south, hills formed of older rocks enclose and restrict the Verde Valley to a narrow canyon. About 25 square miles along the west margin of the Verde Valley encroaches on the northeast corner of Mingus Mountain quadrangle. Here the marginal facies of the Verde formation occurs; but is only locally well displayed, for much of it is masked by Quaternary gravel.

Thickness and stratigraphic relationship

Jenkins (1923), who named the Verde formation from its occurrence in the Verde Valley, first described in detail its character and origin. Jenkins (1923, p. 71) states that in the vicinity of Clarkdale, surface exposures plus well data indicate at least 1,400 feet of Verde formation. Farther south, about 2 miles south of Camp Verde, a well, drilled in search of oil, penetrated 1,650 feet. This thickness in addition to that exposed in adjacent hills totals 2,000 feet or more (Jenkins, 1923, p. 71). Mahard (1949, p. 104) measured two incomplete sections: one, 2 miles north of Clarkdale; the other, near the place where Highway 89A crosses the Verde River near the eastern boundary of the Mingus Mountain quadrangle. The section near Clarkdale exposes 733 feet. Similarly, the section near the highway crossing of the Verde River exposed 1,040 feet. In both places an unknown part of the section lies below the level of the Verde River. About 300 feet of Verde formation is exposed in the Mingus Mountain quadrangle. The actual thickness in this area is somewhat greater, but probably not more than 500–600 feet.

In the Jerome area after mid-Triassic time degradation prevailed until the accumulation of the late Tertiary Hickey formation. Subsequent deformation displaced the Hickey formation and, in part at least, initiated the present drainage pattern and topography. The Verde formation, unconformably overlying the Hickey formation, accumulated in a topographic basin, the Verde Valley, that formed largely during the present erosion cycle. Whether an ancestral Verde Valley existed before accumulation of the Hickey formation, as suggested by Lindgren (1926, p. 14) is not easily determined. Price (1950) found gravel beds along the rim of Sycamore Canyon that underlie Tertiary lava and that contain Precambrian rock types similar to those in the Jerome area. Sycamore Creek, which is a large tributary flowing into the Verde River from the northeast, forms a canyon that lies on the opposite side of the Verde River from the Jerome area. If these gravel beds on the rim of Sycamore Canyon can be correlated with the gravel separating volcanic rocks from older rocks on Mingus Mountain, an ancestral Verde Valley did not exist. In support of this correlation, some channels localizing gravel on Mingus Mountain trend northward. It must be remembered, however, that the evidence in support of the correlation of the two gravel beds is only suggestive. The gravel units along the rim of Sycamore Canyon may be older than the Hickey formation.

On the assumption that faulting along the east and west margins of the Black Hills was contemporaneous, the older gravel beds located north of Highway 89A along the western flank of Mingus Mountain accumulated at the same time as the coarse gravel member of the Verde formation adjacent to the Verde fault. Both of these gravel units accumulated as the result of elevation of the Black Hills along boundary faults. Quaternary terrace deposits and riverwash are the only rocks in the Jerome area younger than the Verde formation.

Lithology

The Verde formation is diverse lithologically, and the beds are lenticular. In the central parts of the basin, fine-grained clastic rocks and limestone prevail, but toward the west margin, these sedimentary rocks intertongue with coarse gravel beds and thin out, so that only coarse clastic rocks accumulated along the Verde fault. Jenkins (1923, p. 73) describes the Verde formation as, "hard impure reprecipitated, cavernous limestone interbedded with sand, gravels, and clays * * *. Near the border or shoreline of the old Verde lake, conglomerate beds are in evidence." He (1923, p. 71) states that the finer grained rocks lie near the south end of the Verde Valley, where apparently the water was deeper in the old lake. Here salt deposits formed, presumably by fractional crystallization from saturated solutions during times when evaporation exceeded influx of rain water. Studies by Mahard (1949, p. 105) near his measured sections clearly indicate the lenticularity of the lake beds. Although his two sections, are only 5 miles apart, he was unable to make any correlations between them, except for a general preponderance of limestone in the upper part of both. He (1949, p. 118) also describes two thin lava flows, 23 and 34 feet

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thick interbedded with lake sediments about 6 miles northeast of Camp Verde, which is near the south end of the Verde Valley. These flows, he believes, originated at House Mountain, which lies eastward.

Of all the lithologic types in the Verde formation, the limestone beds are most conspicuous, for they are aphanitic, tough, cliff- and ledge-formers that terrace the low hills and protrude from the cuts of the larger stream canyons. The view from the eastern slope of the Black Hills toward the north, east, and southeast displays the brilliantly hued Supai cliffs towering above the flat-lying white Verde lake beds. On the weathered surface, they are pure white, owing to a chalky film, and commonly rough or pitted from differential solution. On fresh fracture, they are light pink or light yellow. Small tubelike structures mark many of the limestone beds, and have been interpreted as the result of plant roots. Possibly they indicate abundance of plant life during or before lithification.

Zones of marl separate the beds of limestone, limy siltstone, and fine-grained limy sandstone. These rocks constitute the greater part of most sections, but they are friable and form topographic slopes, which mask their relative abundance. In contrast to the limestone, these rocks are pinkish gray, yellowish gray, and grayish-orange pink on both fresh and weathered surface. The marl is aphanitic, but sporadically sand and small lithic fragments contaminate some beds. In the Verde formation lying within the Mingus Mountain quadrangle, sandstone is rare, but Mahard (1949, p. 105) noted much sandstone in his two sections farther east along the Verde River.

The following is a partial section measured on the south side of the low hill in the center of sec. 5, T. 15 N., R. 3 E. This section represents the best exposure of the limestone and fine-grained clastic facies of the Verde formation lying within the Mingus Mountain quadrangle.

Partial stratigraphic section, Verde formation

	Feet
Marl, pinkish-gray, soft, friable interbedded with similar but indurated marl. All are very limy	23
Limestone, yellowish-gray on fresh fracture, white on weathered surface; fine-grained, ledge-forming. Weathers to a rough, pitted surface	3
Marl and limy siltstone, pinkish-gray, soft, friable. Forms topographic slope	11½
Limestone, yellowish-gray, ledge-forming; appears fragmental. Weathers to a rough surface	2
Marl and limy siltstone, pinkish-gray, soft. Some fine-grained limy sandstone	7
Limestone, grayish-orange pink, thin bedded (12 inches). Contains sand grains and small lithic fragments	2½
No outcrops. Soil indicates friable limy marl and siltstone	8
Limestone, grayish-orange pink, aphanitic, ledge-forming	3

Partial stratigraphic section, Verde formation—Continued

	Feet
Limestone or limy siltstone, some pinkish-gray, sandy; probably some marls. Poorly exposed	16
Limestone, white aphanitic, cavernous, ledge-forming. Weathers to a rough surface	2½
No outcrops. Probably soft, friable marl	13½
Limestone, white, aphanitic, ledge-forming. Contains small gastropodlike tests	4
Marl and limy siltstones, probably friable. Poorly exposed	13
Limestone, white, massive, thick-bedded (2 feet), aphanitic. Weathers to a rough, pitted surface	4½
Marl, pinkish-gray, alternating; limy siltstone; and silty limestones. Thin bedded	13
Base not exposed.	
Total	126½

Along the west margin of the Verde Valley and directly east of the Verde fault, coarse to fine gravel and intercalated sand layers comprise the gravel facies of the Verde formation. The cobbles and boulders were all locally derived. Changes in bedrock along the eastern front of the Black Hills are reflected by a corresponding change in the cobbles and boulders lying downslope. Basalt, the different Precambrian rocks, and the rocks of the Supai, Redwall, and Martin are all represented locally. The matrix of the gravel comprises coarse- to very fine-grained sand, silt, and perhaps clay; these constituents also form local beds or zones in the gravel.

The gravel facies of the Verde formation consists of a series of coalesced alluvial fans that accumulated at the base of an elevated fault block mountain. The distal ends of these fans apparently extended far out into the old Verde Lake so that accumulation of fine-grained lake deposits was interrupted periodically by rapid accumulation of coarse clastic deposits resulting in intertonguing of fan and lake deposits.

There is little doubt that the Verde formation is a lake deposit, and that the lake deposits accumulated throughout an area about 40 by 15 miles in maximum extent. Concerning the origin of the lake, Jenkins (1923, p. 69) writes, "It (the Verde formation) is apparently the result of sedimentation produced both mechanically and chemically through damming of the Verde River by surface lava flows." Six miles north of Camp Verde, Ariz., Mahard (1949, p. 118) noted only two lava flows, the combined thickness of which is 57 feet, interbedded in the Verde formation; whereas 6 miles south of Camp Verde, the section appears to contain nearly equal amounts of lava interbedded with the lake deposits.

Age

Diagnostic fossils have not yet been found in the Verde formation; Jenkins (1923, p. 76–77) found a few fresh-water(?) gastropods on the east side of Island

Mesa (Black Mountain), about 10 miles northeast of Jerome, and Mahard (1949, p. 119) found some small gastropods belonging to the family Helicidae north of the smelter at Clarkdale. The writers noted a few small gastropod-like tests in the limestone cropping out in the low hills southwest of the Clemenceau airport.

Both Jenkins (1923, p. 77) and Mahard (1949, p. 126) postulate a late Pliocene or early Pleistocene age for the Verde formation, mainly on the relationship of the Verde formation and the basalt near the confluence of the Verde River and Sycamore Creek, north of Jerome. Here the Verde formation overlies the basalt, which is confined to the walls of Sycamore Canyon, and they believe that the basalt is equivalent to the lava on the plateau near the San Francisco Peaks. Robinson (1913) subdivided the lava flows in the region near the San Francisco Peaks into three stages whose ages ranged from late Pliocene to Pleistocene with minor eruptions occurring in Recent time. Jenkins and Mahard, accepting Robinson's age for the basalt, date the Verde formation as late Pliocene and (or) early Pleistocene. But as discussed in the section on the age of the Hickey formation, page 58, the evidence for the age of the basaltic lava on the edge of the Colorado Plateau north of Jerome is not conclusive.

The absolute age of the Verde formation probably will not be determined satisfactorily until someone finds diagnostic fossils. The relative age of the formation, which is most important, is clear enough. The Verde formation is younger than the Hickey; and in large part at least, younger than the movement on the Verde fault that elevated the Black Hills to their present position. In part, it is probably contemporaneous with the older Quaternary gravel that occurs along the western front of Mingus Mountain north of Highway 89A, for both bear the same relation to the Black Hills. The younger Quaternary gravel covers pediments cut on the Verde formation.

OLDER QUATERNARY GRAVELS

Distribution, thickness, and stratigraphic relationship

North of Highway 89A in Lonesome Valley west of the boundary fault along the Black Hills, ridges and low hills of coarse gravel rise above the general level of the pediment surface. Pediment gravel beds largely mask the extent of these older gravels.

The thickness of the older Quaternary gravel is unknown, for exposures are erosional remnants. Several of the larger ridges, however, expose about 70 feet of gravel, which can safely be considered as a minimum thickness.

The older Quaternary gravel is older than the gravel that caps the pediments and younger than the Hickey formation. The older Quaternary gravel accumulated after the faulting that displaced the Hickey and elevated the Black Hills. These gravel facies bear the same relation to the Black Hills on the west as the coarse gravel facies of the Verde formation does on the east. On the reasonable assumption that faulting occurred simultaneously on both margins of the Black Hills, the two gravel facies accumulated simultaneously. In part, the pediments are cut on the older Quaternary gravel, as they are on the Verde formation.

Both south and east of the Coyote Spring Ranch, the older Quaternary gravel rests directly on Precambrian rocks. The Hickey formation probably had been stripped from these areas before deposition of the gravel.

Lithology

The fragments composing the older Quaternary gravel were derived almost entirely from Paleozoic rocks and Tertiary basalt. Locally, however, the gravel contains a few scattered cobbles and boulders of Precambrian rocks. South of Coyote Spring Ranch, these fragments of Precambrian rock may have come from bedrock. But north of the ranch, no Precambrian bedrock crops out at altitudes higher than the base of the gravel. Here the source of the Precambrian rock fragments must be the sedimentary facies of the Hickey formation, which contains an abundance of such fragments and is well exposed in this vicinity in Martin Canyon.

YOUNGER QUATERNARY GRAVELS

Distribution

The younger Quaternary gravel facies constitutes the surface mantle in most of the Verde Valley and much of Lonesome Valley, exposed in the Jerome area. In the Verde Valley, the younger Quaternary gravel mantles most of the surface lying between the Verde fault and the Verde River, with the exception of the nearly continuous belt of coarse gravel of the Verde formation that crops out adjacent to the Verde fault. Additional small, irregularly shaped areas of Verde formation, chiefly limy silt, marl, and fresh-water limestone, crop out in inliers eroded through the gravel beds. In Lonesome Valley, younger gravel beds crop out west of the Black Hills and north of 1,300,000. Indian Hills and similar isolated hills lying northward protrude through the gravel, to break the monotonous even surface.

Thickness and stratigraphic relationship

The younger Quaternary gravel beds, which are essentially gravel veneers on pediment surfaces, range from less than a foot to 30 and perhaps as much as 50 feet. They are overlain by terraces along the Verde

River and by Recent riverwash, which collects along the major streams and gulches.

The younger Quaternary gravel in large part overlies the Hickey and Verde formations with a small angular unconformity. Along the west side of the Black Hills, however, this younger gravel overlaps the Hickey onto Precambrian and Paleozoic rocks. The overlap of the gravel on Precambrian rocks is well displayed along the south side of the Indian Hills, and the overlap on Paleozoic rocks in the vicinity of Little Coyote and Martin Canyons north of Highway 89A.

Lithology

The younger Quaternary gravel in large part is a heterogeneous accumulation of gravel composed of boulders, cobbles, pebbles, and some finer grained sediment. Sorting is characteristically poor, but sporadic individual beds that are somewhat sorted stand out in marked contrast. Color ranges from red, through pink, to gray. With the exception of quartz diorite, fragments of Precambrian rocks produce red gravel, and a preponderance of fragments of Paleozoic rocks and Precambrian quartz diorite results in a gray color. Mixtures of the two, give intermediate colors, but red hues predominate.

The composition of these gravel beds reflects the composition of the types of bedrock found uphill. In most places, the gravel contains fragments of Precambrian and Paleozoic rocks and Tertiary basalt. In the southern part of the Jerome area, fragments of Precambrian rocks predominate, but north of Highway 89A in Lonesome Valley, the gravel was derived almost completely from Paleozoic rocks and Tertiary volcanic rocks.

RECENT RIVERWASH, TERRACE DEPOSITS, AND TALUS

The bulk of the Recent riverwash occurs in the Agua Fria and Verde Rivers and their flood plains. The major tributaries of these two rivers also contain variable but significant amounts of riverwash, and the Recent riverwash in the largest of these tributaries has been indicated on plate 1. The smaller gulches and washes contain narrow "ribbons" of Recent riverwash, especially where they cross Lonesome and Verde Valleys, but the amount is too small to show on the scale used for illustration.

Terrace deposits formed along the flanks of the Agua Fria and Verde Rivers, especially the latter, and in local places along Ash Creek. These deposits along the Verde River have been distinguished in mapping; they are used for farming through irrigation and for the townsite of Cottonwood. The terraces along the Agua Fria are not so well defined and have been only used for farming south of Dewey. They are included with the riverwash. The small terrace deposits along Ash Creek are near 1,311,500 N.; 446,000 E.

Talus deposits occur in two localities: one along the escarpment west of Mingus Mountain between 1,340,000 N. and 1,350,000 N., and the other, east of the Black Hills on the southern bank of Black Canyon near 1,330,000 N.; 458,000 E. Beneath the lava on Mingus Mountain, other talus patches whose extent was sufficient to show on the regional map, occur sporadically, but were not mapped because scattered inliers permitted recognition of the underlying bedrock.

STRUCTURE OF THE OLDER PRECAMBRIAN ROCKS

Throughout Arizona, Anderson (1951) recognized a broad similarity in the structure of the older Precambrian rocks. Folds trending northward or northeastward occur in the Little Dragoon and Mazatzal Mountains, and at Bagdad, and similar trends plus northwest trends are present in the Jerome area. In the Mazatzal Mountains, low-angle southeastward-dipping thrust faults occur.

The degree of metamorphism in the older Precambrian rocks is variable. Intense dynamic metamorphism resulting in local boudinage structure is present in the Little Dragoon Mountains. Wilson (1939) reported weak dynamic metamorphism in the Mazatzal Mountains. In the Bagdad area, dynamic metamorphism is only of local importance, and thermal metamorphism related to granite is strong. A brief description by Ransome (1903, 1904, 1919) of the fissile character of the Pinal schist at Globe, Miami, Ray, and Bisbee, and the description of the same schist at Morenci, by Lindgren (1905) indicate regional metamorphism throughout a widespread area. Gneissose and schistose structures are present in the Vishnu schist at the Grand Canyon (Noble and Hunter, 1917) and Campbell and Maxson (1938) noted folds and minor drag folds in the Vishnu.

In the Jerome area, the Shylock fault separates the Alder and Ash Creek groups. Their stratigraphic relationship is unknown and their structure is different. In the Ash Creek group, open folds (B-tectonites) are the dominant structure. Foliation, which developed erratically, ranges from pronounced to imperceptible. In places the foliation generally parallels the axial planes of the folds and in other places it does not. In the Alder group, foliation, strike faults, isoclinal folding parallel to the foliation (S-tectonites) are the dominant structural features.

STRUCTURE OF THE ASH CREEK GROUP

The structure of the Ash Creek group is dominated by two major folds: the Tex syncline and the Mingus

anticline (fig. 3). The Tex syncline, in the southwestern part of the area, strikes northwest, whereas the Mingus anticline to the northeast and partly covered by younger rocks, has a sigmoid pattern, owing to deformation after folding.

Minor folds are common on the flanks of the major folds. Near Jerome the foliation is intense and is essentially parallel to the axial planes of minor folds. The term "axial plane foliation" is used in this report to indicate the geometric relation of foliation to axial planes of folds formed during a common period of deformation. We do not wish to imply that the foliation is "flow cleavage" of Leith (1923, p. 113–135) developed at right angles to the compressive stress. In the zones of intense deformation where foliation and tight folds formed simultaneously, continuous deformation rotated the early foliation (shear) planes to positions more nearly parallel to the axial planes of the folds. Later shear planes of the same period of deformation intersect the earlier ones at acute angles, so that the result is a family of planes whose average strike and dip parallels the axial plane of the folds. This is the megascopic foliation seen in outcrops. Once the approximate parallelism between the average strike of the foliation and the fold axes has been established, the attitude of the foliation indicates the attitude of the major fold axial planes, and the lineation, chiefly bedding intersections on the foliation planes, indicates the plunge of the folds. This approximate geometrical relationship between foliation and axial planes of the fold provides a means of untangling the major structure in the areas where the metamorphism has obliterated nearly all of the bedding.

In the southern part of the Ash Creek terrane, weak and sporadic foliation is discordant to the axial planes of the folds and bears no relationship in time to the folding (pl. 1).

Two sets of faults are generally present; the older set strikes approximately east, and the younger set strikes north. The eastward-trending faults, in part, are occupied by quartz porphyry dikes. The northward-trending faults, in part, are occupied by granodiorite porphyry dikes.

TEX SYNCLINE

The Tex syncline is present in the southwestern part of the Ash Creek terrane, and the southwest limb is exposed along Tex Canyon, suggesting this name for the fold (pl. 1). The northwestern part of the syncline contains the coarse breccia unit of the Grapevine Gulch formation in the trough of the syncline, and on the southwest side of the breccia unit beds face and dip northeast, whereas on the northeast side of the breccia, beds face and dip southwest. In the southeastern part of the syncline, the trough of the syncline is difficult to determine because bedding is obscure in the thick dacitic flows in the Grapevine Gulch formation.

The Tex syncline plunges southeastward throughout as shown in the trough of the fold by many southeast dips ranging from 20° to 50° but averaging about 30° in the Grapevine Gulch formation. The trough of the syncline is broad and irregular, and is marked by many minor flexures revealed by the trace of some of the jasper-magnetite beds and minor variations in the attitude of bedding. The exact location of the trace of the axial plane in such a complex syncline is difficult to determine.

The Tex syncline is faulted at the eastern contact of the coarse breccia unit of the Grapevine Gulch formation (1,320,000 N.; 422,400 E.), but this fault cannot be traced for any great distance and may turn northwestward into a bedding fault. Whether the trough of the syncline is actually displaced by this fault as much as is indicated on plate 1 might be questioned. East of the breccia unit the disposition of the bedding in the Grapevine Gulch formation was the basis for determining the location of the trough, and it is conceivable that drag along the fault has obscured the true position of the trough.

To the east of the coarse breccia unit, a northward-trending fault can be traced for a much longer distance, and the northward offset of the trough of the Tex syncline is demonstrated more clearly.

Just east of the Shylock fault, which cuts off the Tex syncline, the plunge of the syncline must flatten appreciably and the northeasterly limb is diverted to a more northerly strike along the Shylock fault. This change in plunge is indicated by the failure of the units in the Grapevine Gulch formation to curve around the coarse breccia unit. Perhaps the flattening of plunge is related to movement along the Shylock fault or it may represent regional deformation before faulting.

A minor anticline south of Yaeger [3] Canyon plunges southeastward and is cut off by a northward-trending fault. To the east a minor anticline revealed by the curvature of a dacitic flow, plunges northward. Possibly drag along the fault is responsible for this reversal of plunge.

North of the south end of the Tex syncline and adjacent to the quartz diorite (fig. 3) a subsidiary syncline plunges northwestward; to the north it is cut off by the quartz diorite. Clearly the southeastern extension of the Tex syncline is cut off by the quartz diorite, whereas the southwestern limb is intruded by quartz diorite without any apparent disturbance of the synclinal structure.

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[3] Incorrect spelling on the Mingus Mountain quadrangle.

MINGUS ANTICLINE

The Mingus anticline is well exposed southeast of Mingus Mountain and passes beneath the Paleozoic and younger rocks in the southern part of Mingus Mountain, suggesting this name for this important structural feature (pls. 1 and 2).

The Mingus anticline is essentially domical, plunging southeastward in the vicinity of Black Canyon and

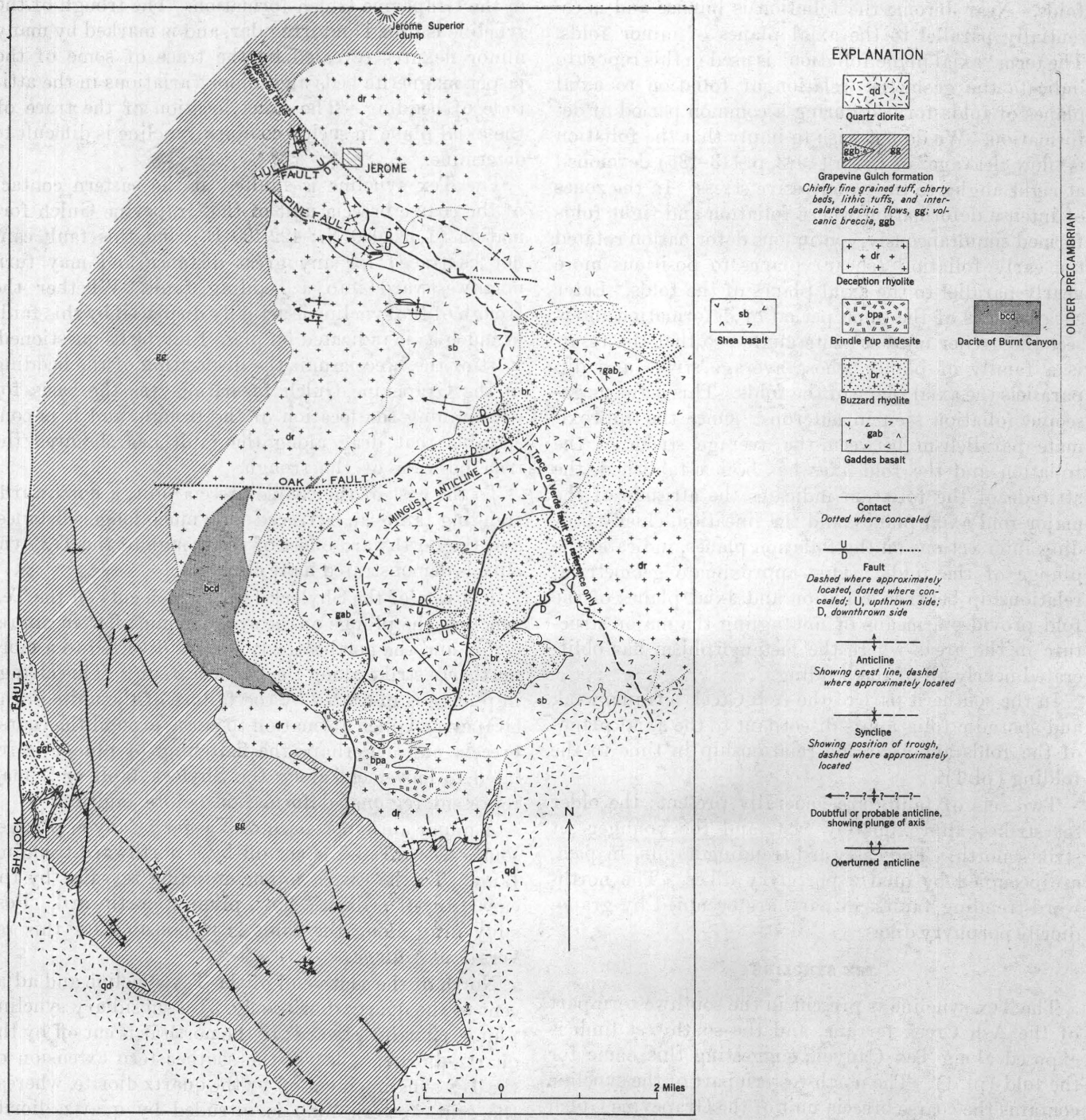


FIGURE 3.—Interpretative structural map of Ash Creek group. Quartz porphyry and gabbro intrusive bodies deleted.

northwestward at Jerome. It is our conclusion that the Mingus anticline has been warped to a sigmoid pattern and displaced by the Oak fault (fig. 3).

The southeastward-plunging nose of the Mingus anticline is well exposed in Black Canyon where the Buzzard rhyolite overlies the older Gaddes basalt on three sides.

Although beds cannot be traced around the nose of the fold, the northwesterly plunge of the northern nose of the Mingus anticline at Jerome (fig. 3) is indicated by several lines of evidence. Fragments of Grapevine Gulch formation on the Jerome Superior dump east of the Verde fault suggest a regional fold in the Grapevine Gulch formation (fig. 3), because the Grapevine Gulch formation is only exposed west of Jerome where the beds face west and strike north-northeast, and where drag fold patterns indicate a northward-plunging anticline to the east. Furthermore the rocks exposed in the United Verde Extension mine are chiefly Deception rhyolite (pl. 5), and the beds in this formation in the mine probably face east, indicating that they are on the eastern limb of the Mingus anticline. Additional details are given in the section on the United Verde Extension mine page 140. By placing the plan of the geology of the 1,400-foot level of the United Verde Extension mine adjacent to the plan of the geology of the surface of the United Verde mine (pl. 5), the geology of the mine area is depicted essentially on the same horizontal plane as before Tertiary displacement on the Verde fault. The structural discordance between the Grapevine Gulch formation, beds facing west and striking north, and the Deception rhyolite, beds facing east and striking northwest, indicate a northward-plunging anticline cut by a fault essentially parallel to the axial plane of the fold (fig. 3).

Observations made in the pit of the United Verde mine at Jerome show that the foliation in the Grapevine Gulch formation is essentially parallel to the axial planes of the small folds, indicating axial plane foliation that is presumably generally parallel to the axial plane of the major fold (Mingus anticline). The strike of the megascopic foliation is N. 20° W. near the United Verde mine, but to the south, the foliation and the crestlines and troughs of the smaller flank folds both swing to a more easterly trend (pl. 1), indicating that the axial plane of the Mingus anticline is bent in the same direction (fig. 3).

The evidence for the warping of the south end of the Mingus anticline is less secure, and an alternate structural interpretation is possible, if only the Gaddes basalt-Buzzard rhyolite contact south of the Oak fault is considered. The trace of this contact from Black Canyon (pl. 1) 1,330,000 N.; 450,000 E. northwestward, northward, and east-northeastward to the Verde fault indicates that the Buzzard rhyolite here is in a southeastward-plunging syncline. A southeastward-plunging anticline might be expected to the east (east of the Verde fault), which would parallel the exposed southern part of the Mingus anticline. Clear-cut bedding data are not available in the exposures of Gaddes basalt south of Oak Wash to prove or disprove this possibility.

This alternate structural interpretation, however, is challenged, and the interpretation of a warped Mingus anticline is strengthened by north-facing flows in the Gaddes basalt just south of the Oak fault and by the presence of Shea basalt and Buzzard rhyolite north of Oak Wash. Along Oak Wash and near the Verde fault (fig. 3) the pattern of pillows in one of the flows suggests that the flows are facing north. The Gaddes basalt clearly faces south along the Gaddes basalt-Buzzard rhyolite contact south of 1,340,000. If the evidence from the pillows in Oak Wash is reliable, a northeast anticlinal structure is proven south of Oak Wash, and warping of the Mingus anticline seems very likely. The evidence from the pillow lavas, however, is too uncertain to use as the main link in the chain of argument; for this we turn to the Shea basalt and Buzzard rhyolite north of the Oak fault.

In Black Canyon, the Buzzard rhyolite and Shea basalt, in that order, clearly overlie the Gaddes basalt. The basalt and rhyolite north of Oak fault dip and face northwesterly which indicates that the basalt is younger than the rhyolite. The basalt and rhyolite north of Oak fault can be correlated with the basalt and rhyolite of similar lithology and sequence in Black Canyon. The basalt and rhyolite north of Oak fault therefore must be younger than the Gaddes basalt. However, the alternate structural interpretation, that is, a southeastward-plunging anticline to the east of the Verde fault and south of the Oak fault would require that the basalt and rhyolite north of the Oak fault be older than the Gaddes basalt and face to the southeast. Assuming our correlation of the Shea basalt and Buzzard rhyolite north of Oak fault and in Black Canyon is correct, and we are convinced that the evidence requires this correlation, the most plausible explanation for northwestward-facing Shea basalt and Buzzard rhyolite north of Oak Wash is that the Mingus anticline was warped from a northwest to a northeast trend (fig. 3).

Additional evidence for the interpretation of a warped anticline is found in the trend of the contacts on the southern limb of the Mingus anticline westward from Black Canyon. The Buzzard rhyolite-dacite of Burnt Canyon contact trends northwestward near 1,333,000 N.; 434,000 E. in Burnt Canyon where it

changes strike to north-northeast. In similar manner, the contact between the Grapevine Gulch formation and older volcanic formations trends west-northwestward in the area south of Black Canyon, but near 1,334,000 N.; 429,000 E. the contact changes strike from northwest to northeast. Both of these areas where the trends of formations change from northwest to northeast are in the correct spatial position to be on the axis of the crossfold that warped the Mingus anticline.

Though no single item is conclusive, the whole of the available evidence suggests warping of the south end of the Mingus anticline to a northeast trend south of the Oak fault. The bending of the axial plane of the fold from Jerome southward permits joining the two anticlinal structures in a general way to produce the sigmoid pattern (fig. 3). Our interpretation, although admittedly not proven, is consistent with the known facts, as is no other interpretation that has occurred to us.

The Mingus anticline is cut by three westward-trending faults; the Hull, Pine, and Oak. Only the Oak fault will be discussed in this section as the other faults are discussed below. The Oak fault truncates the Shea basalt, Buzzard rhyolite, and the Gaddes basalt in the Oak Wash drainage basin. The only marker that can be used is the Buzzard rhyolite-Shea basalt contact. On the south side of the Oak fault, this contact is buried beneath younger rocks, therefore no measurement of horizontal separation can be made. Figure 3 indicates a horizontal separation of 7,000 feet, but this amount is only an approximation based on the inferred outcrop widths and trends of the formations beneath the younger rocks. If the structural interpretation given on figure 3 is essentially correct, vertical displacement on the Oak fault would give some horizontal separation because the contacts dip northwest on the north side of the fault. Our structural interpretation requires some horizontal displacement in order to provide the necessary space for the Buzzard rhyolite north of the Oak fault, as there is no evidence of steep northeast plunge of the Mingus anticline where it crosses the trace of the Verde fault (fig. 3).

Two northward-trending faults cut the Mingus anticline in the southern part of the area (1,320,000 N.; 441,000 E.; 1,330,000 N.; 449,500 E.), and the displacements along these faults may be entirely vertical. The structure between these faults is essentially horstlike, the blocks on both sides dropped relative to the central block. These northward-trending faults are younger than the eastward-trending quartz porphyry dikes and in part are older than the northward-trending granodiorite porphyry dikes. Both faults die out north and south, and do not displace the Paleozoic rocks.

North of Black Canyon (1,333,300 N.; 454,000 E.) a northeast-striking normal fault with southeast dips ranging from 35° to 45°, brings younger Shea basalt into fault contact with older Buzzard rhyolite. This fault dies out to the southwest and is truncated by the Verde fault to the northeast. Presumably the displacement along the fault is small; no evidence is available on the time of movement.

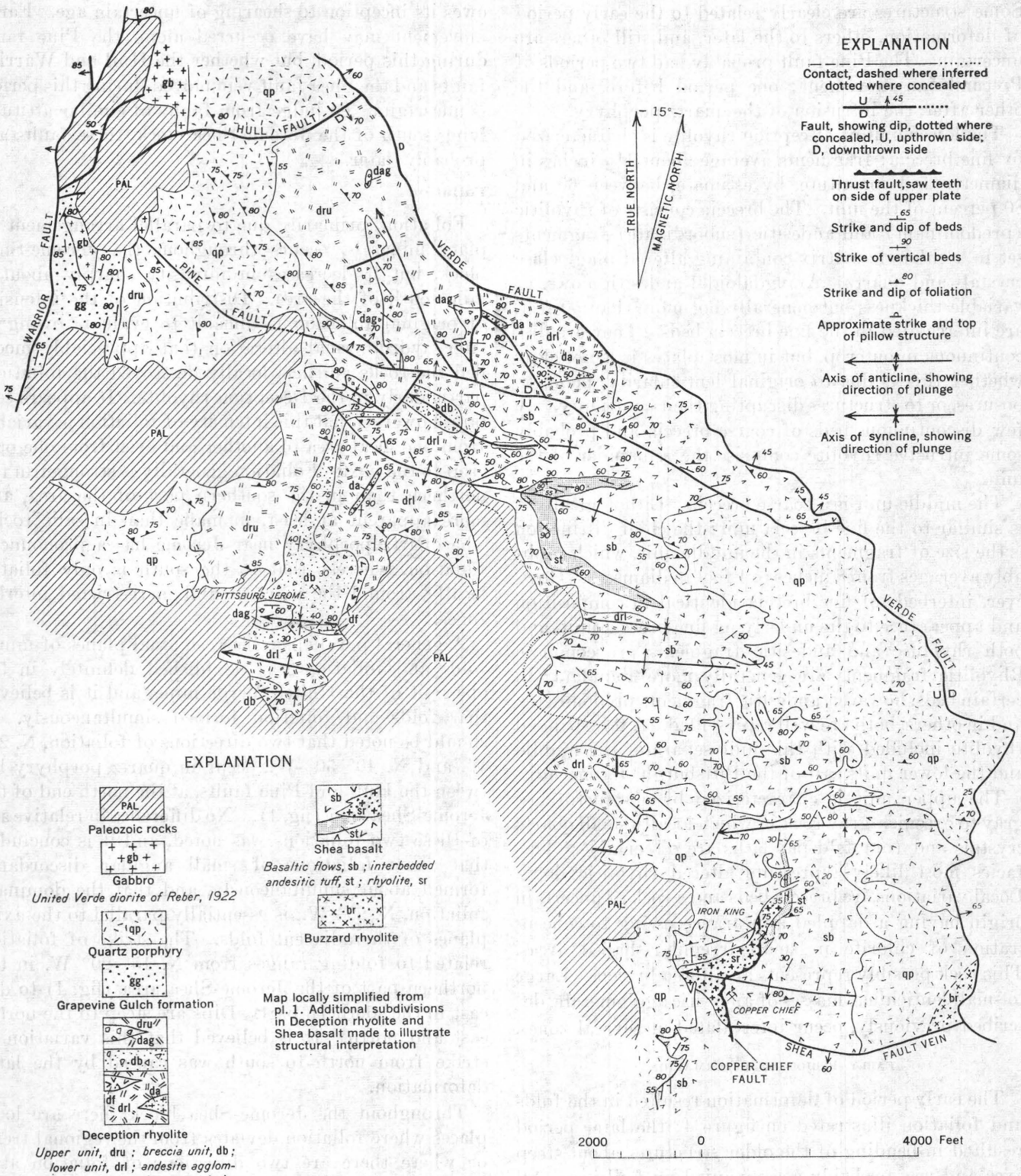
The Ash Creek group in the drainage of Black Canyon contains a weak foliation, largely striking east-northeast, although locally west-northwest. This foliation clearly cuts across the southern nose of the Mingus anticline and must have formed after the Mingus anticline. In the Oak Wash drainage where the Mingus anticline appears to trend northeast, this younger foliation is essentially parallel to the northeast trend of the crest of the Mingus anticline. Perhaps the east-northeast foliation is related in time to the deformation that warped the Mingus anticline into a sigmoid pattern, which would imply that the regional trend of the Mingus anticline was northwest, parallel to the present trend of the Tex syncline. The Mingus anticline was deformed before or during the faulting along the Oak and Pine faults, and the stresses warping the Mingus anticline may have produced the eastward-trending faults.

JEROME–SHEA BLOCK

The area from Jerome to the Shea vein has been chosen for more detailed description because of the complex structure and because of the strong hydrothermal alteration that may be related to ore deposition. Hydrothermal alteration has greatly modified the composition of the rocks and has destroyed relict textures and structures that survived regional metamorphism; for this reason structures were more difficult to determine and chances for error, especially in local structural details, were greater. Consequently, we spent more time in the field compiling data for this area than for any other of comparable size in the Ash Creek group.

The Hull fault is the northern boundary of the Jerome–Shea block, and the Shea fault-vein the southern boundary (fig. 4). The contact between Paleozoic and Precambrian rocks forms the western boundary and the Verde fault the eastern.

The block is underlain by Buzzard rhyolite, Shea basalt, Deception rhyolite, Grapevine Gulch formation, and intrusive rocks, chiefly quartz porphyry. To more accurately determine the local structure, the Deception rhyolite was subdivided into three units, termed: upper, middle, and lower units. These units are only local; other areas of Deception rhyolite were not amenable to a similar subdivision. The rocks contain structures that indicate two periods of deformation; the earlier probably more intensely deformed these rocks than the later.



EXPLANATION

PAL — Paleozoic rocks

+gb+ — Gabbro

United Verde diorite of Reber, 1922

qp — Quartz porphyry

gg — Grapevine Gulch formation

dru, dag, db, da, df, drl — Deception rhyolite

Upper unit, dru *; breccia unit,* db *; lower unit,* drl *; andesite agglomerate in upper unit,* dag *; andesite flows,* da *; and rhyolite flows,* df *; in lower unit*

sb, sr, st — Shea basalt

Basaltic flows, sb *; interbedded andesitic tuff,* st *; rhyolite,* sr

br — Buzzard rhyolite

Map locally simplified from pl. 1. Additional subdivisions in Deception rhyolite and Shea basalt made to illustrate structural interpretation

FIGURE 4.—Interpretative structural map of Jerome–Shea block.

Some structures are clearly related to the early period of deformation, others to the later, and still others are uncertain. The Pine fault probably had two periods of Precambrian movement: one period before, and the other after, the intrusion of the quartz porphyry.

The lower unit of Deception rhyolite is characterized by fine breccia; fragments average about 1½ inches in diameter and constitute, by estimate, between 60 and 80 percent of the unit. The breccia consists of rhyolitic (predominant) and andesitic (subordinate) fragments set in a chloritic matrix containing altered plagioclase crystals and quartz. Amygdaloidal andesitic flows, of variable thickness but generally not more than 50 feet, are intercalated in the fine breccia beds. They are discontinuous in outcrop, but in most places it is not clear whether this is due to original lenticularity, poor exposures, or to structures disrupting their continuity. A few discontinuous beds of coarse breccia and probably some intrusive rhyolite comprise the remainder of the unit.

The middle unit is a coarse breccia. Lithologically it is similar to the fine breccia unit; the chief distinction is the size of fragments in the middle unit, which probably averages from 8 inches to 1 foot in diameter. However, interbeds of fine breccia identical in composition and appearance to the underlying fine breccia unit, and both rhyolitic and andesitic fragments are common. Rhyolitic fragments are generally more abundant; in certain beds, however, andesitic fragments predominate, and in others, both are well represented. A flow-banded rhyolite included with this unit separates the middle and the lower units east of the Pittsburgh–Jerome adit.

The upper unit is characterized by a structureless gray to greenish-gray rock in which locally small quartz crystals and (or) feldspar laths are discernible. This facies most likely represents altered rhyolitic flows. Local variations are breccia beds and a rock of uncertain origin having a dappled appearance owing to concentration of chlorite in small irregular-shaped areas. This rock possibly represents an altered breccia. Large to small lenticuluar masses of andesitic agglomerate, described previously, occur intercalated at several zones.

EARLY PERIOD OF DEFORMATION

The early period of deformation resulted in the folds and foliation illustrated on figure 4; the later period resulted in bending of the older structures about steep axes and was probably accompanied by faulting. All the folds whose crests and troughs are shown on figure 4 originated during the early period. The dominant foliation formed simultaneously with folding and is essentially parallel to the axial planes of the folds. Locally some foliation discordant to folds probably owes its inception to shearing of uncertain age. Early movement may have occurred along the Pine fault during this period, but whether the Hull and Warrior faults and the Shea fault vein moved during this period is uncertain. To judge from faults of similar attitude lying south of the Jerome-Shea block, these faults are probably later.

Foliation

Foliation, primarily due to parallel arrangement of platy minerals, occurs throughout the Jerome-Shea block, but it decreases in intensity sharply about a mile south of the area. Distribution of the intensity of original foliation is difficult to evaluate owing to later hydrothermal alteration that appreciably modified the rocks and has locally destroyed the foliation, particularly in the Deception rhyolite. It appears certain, however, that this intensity of foliation is variable from one local area to another and even from one outcrop to the next. Foliation generally is stronger in the northern than in the southern part of the area, and more apparent at least, in mafic than in salic rocks. The quartz porphyry near Jerome has a pronounced foliation, whereas that to the south is only foliated locally around the margins and in restricted interior zones.

A general parallelism between axial planes of minor folds and foliation was established definitely in the open pit of the United Verde mine, and it is believed that folds and foliation formed simultaneously. It should be noted that two directions of foliation, N. 20° W. and N. 40°–50° W., occur in quartz porphyry between the Hull and Pine faults, at the north end of the Jerome-Shea area (fig. 4). No difference in relative age of these two foliations was noted, and it is concluded that these foliations of small angular discordance formed about simultaneously, and that the dominant foliation, N. 20° W., is essentially parallel to the axial planes of the adjacent folds. The strike of foliation related to folding ranges from N. 15°–20° W. in the northern part of the Jerome-Shea area (fig. 4) to due east in the southern part. Dips are steep to the northeast and north. It is believed that this variation in strike from north to south was caused by the later deformation.

Throughout the Jerome-Shea block there are local places where foliation deviates from the regional trend or where there are two directions of foliation at a marked angular discordance. Some of these are clearly due to zones of shearing, and commonly quartz with or without sulfide minerals occurs in them; the origin of aberrant foliation in other places is obscure. Foliation in a small area from 2,000 to 3,000 feet east of the Pittsburgh-Jerome mine appears to be from 50° to

60° discordant to the local trend of the fold axis. Whether this discordance is real or only apparent owing to local unrecognized complexities of structure is uncertain.

Folds

Three anticlines and three synclines occur within the Jerome-Shea block. The Pine fault has probably offset the northern anticline, and there is probably a small offset in the southern anticline along the Copper Chief fault not indicated in figure 4. The magnitude of the folds ranges from about 2,500 to about 6,000 feet. The central anticline and the northern syncline are tight but the other folds are much more open. Folding after deformation modified the strike of the axial plane so that there is a general swing from about N. 20° W. in the northern part of the block to about due east in the central and southern parts. To judge from foliation and bedding, axial planes are nearly vertical or dip steeply northward.

For some of the folds, such as the northern and southern anticlines, an exact location of the crestline or trough is difficult; bedding commonly is sparse in critical areas, and minor folds, faults, and overturning have complicated the structure. The extent to which individual attitudes do not reflect the major trends is well illustrated by attitudes in the tuffaceous rocks intercalated in the Shea basalt, figure 4.

South of the Pine fault, the position of the northern anticline is well established by such marker beds as the breccia unit of the Deception rhyolite and the andesitic tuff unit of the Shea basalt. North of the Pine fault, however, it is not quite so certain owing in large part to the more uniform lithology of the upper unit of the Deception rhyolite. The disposition of the andesitic agglomerate, which is the only large mass of contrasting rock type in the upper unit of Deception rhyolite, indicates duplication on opposite limbs of an anticline. This fold, to the north of Pine fault, probably is the faulted segment of the anticline to the south of the Pine fault (fig. 4).

In the southern part of the block, a zone containing bedded tuffaceous sedimentary rock was traced for about 2,000 feet south and southwest to the intersection of the Copper Chief fault. Dips are steep to the west, and locally small folds are present. A short distance west of the intersection of the Copper Chief fault and the bedded zone, another zone characterized by bedding (perhaps the same zone) occurs. Here bedding strikes east, but within a short distance it swings toward the south and in some places can be recognized adjacent to the quartz porphyry southward to about the southern boundary shown on figure 4. Small folds appear to be especially abundant where the zone changes from an easterly to southerly trend. Apparently part of this zone is missing owing to the intrusion of the quartz porphyry mass; and locally within the quartz porphyry, inclusions of bedded rock occur not far removed, if any, from their original position. These bedded zones and an intercalated mass of rhyolite in the Shea basalt (pl. 1) establish the southern anticline. The fold is open and plunges westward.

Pine fault

The Pine fault strikes about N. 65° W. and dips 60°–75° N. It is cut by the Verde at the south end and Paleozoic rocks overlie it on the north; probably the Pine terminates against the Hull beneath the Paleozoic cover. Along part of its course, the Pine forms the boundary between quartz porphyry and Shea basalt and between Deception rhyolite and Shea basalt. An early period of normal movement offset the northern anticline and the contact between Deception rhyolite and Shea basalt. The footwall (south side) of the fault moved northward with respect to the hanging wall, offsetting the anticlinal crest about 1,500 feet and the Shea basalt Deception rhyolite contact about 2,200 feet. A short distance northwestward from the anticline, however, the Deception rhyolite-quartz porphyry contact is offset about 100 feet. Two periods of movement on the Pine took place, an early period that occurred after folding but before intrusion of the quartz porphyry and a later period that slightly offset the quartz porphyry.

LATER PERIOD OF DEFORMATION

The later period of deformation warped the earlier structures about steep axes forming broad curves to foliation trends and to crests and troughs of folds. The poorly defined traces of axial planes of these warps strike about northeast, and the strike of early structures progressively change from N. 15°–20° W. in the northern part of the block to about due east in the central and southern areas. The axes of these flexures must plunge steeply, for there is no significant range to the dips of the deformed foliation. The Hull, Shea, and the Copper Chief faults do not appear to be deformed; hence movement probably occurred during the latter part or after this period of deformation. Possibly the renewed movement along the Pine fault occurred during this period. If there is a component of Precambrian movement on the Warrior fault, it probably occurred at this time.

FAULTS OF UNDETERMINED AGE

Hull and Warrior faults

The Hull and Warrior faults are alike in that movement occurred on them after the deposition of Paleozoic rocks. It is probable that the Hull also had a Precambrian period of movement, but the small extent of

Precambrian rocks exposed against the Warrior prevents determination of whether or not it also moved during the Precambrian. Movement on the Warrior fault later than Paleozoic time is described elsewhere.

Strike of the Hull fault ranges from due east to N. 55° E. In the only place where the dip was observed, it was steep toward the south. Quartz porphyry, Deception rhyolite, and Grapevine Gulch formation are the Precambrian rocks on the footwall (north side) of the fault (pl. 5); quartz porphyry and Deception rhyolite occur on the hanging wall side. The Hull terminates against the Verde fault on the east and against the Warrior fault on the west. Displaced Paleozoic rocks show about 100 feet of normal vertical separation and about 700 feet of left lateral offset. This offset is no more than would result from a dip slip of about 100 feet. A Precambrian component of movement is indicated by the relative displacement of the quartz porphyry-Deception rhyolite contact; north of the fault (pl. 5, 1200 S., 3200 E.) this contact is offset eastward about 1800 feet in relation to its position south of the fault (pl. 5, 1680 S., 1400 E.). The contact between quartz porphyry and Deception rhyolite south of the fault, as exposed in the walls of Deception Gulch, dips nearly vertically. This suggests an appreciable component of strike slip movement in Precambrian time.

Verde fault

The displacement on the Verde fault is discussed at length in the section on the "Relationship of the United Verde deposit to the United Verde Extension deposit." Here, it suffices to summarize the conclusions: (1) the maximum vertical separation of the Precambrian movement is about 1,000 feet, and may be less, and later displacement is about 1,500 feet; (2) the maximum total net slip, which includes displacement during Precambrian and later time is about 3,300 feet, and its approximate bearing and plunge are S. 85° E. at 50° E.; the rake is about 66° S. These maximum calculations are based on the assumption that the intersection of a small mass of Deception rhyolite and the Verde fault define a point that can be recognized on both the hanging wall and footwall of the Verde fault.

Shea fault vein

The Shea fault vein occurs about 1,500 feet south of the Copper Chief mine and forms the southern boundary of the Jerome-Shea block (fig. 4). The average strike is about due east and the dip is southward. Reber (1938, p. 58) reports a dip of 42° S. in the Shea mine and a dip of 60° S. was recorded from surface exposures about 1,000 feet east of the mine. On the west, the fault vein is cut off by the Copper Chief fault, and on the east it probably continues to the Verde fault. Reber reports that the fault vein probably has a throw of more than 100 feet. Surface exposures indicate an offset of quartz porphyry contact of about 200–250 feet with the north side moving eastward in relation to the south.

The age of the Shea fault vein is uncertain. Movement occurred later than the granodiorite porphyry dikes, which are later than the foliation, but earlier than movement on the Copper Chief fault. The Shea fault vein has not been recognized in the hanging wall of the Copper Chief fault.

Copper Chief fault

The Copper Chief fault is in the south-central part of the Jerome-Shea block (fig. 4). It is well exposed at the portal of the Iron King adit, and cuts the adit of the Copper Chief mine (fig. 30). It strikes northward and dips at a low angle westward. Reber (1938, p. 58) reports a dip of 13° W. for the fault in the underground workings in the Shea mine. Northward the fault cuts quartz porphyry but is lost soon thereafter. Southward it is traceable without difficulty for only about 2,000 feet south of the Shea mine, where the fault is largely within the Shea basalt, but locally it cuts granodiorite porphyry dikes and quartz porphyry. From detailed studies of the Copper Chief and Iron King mines, Reber (1938, p. 56) estimated the offset to be about 300 feet in a general direction more nearly east than southeast.

STRUCTURE OF THE ALDER GROUP

The discernible Precambrian structural history of the Alder group embraces an early period of deformation followed by at least two later periods. Locally distinct structural trends permit separation of the early structures from the later. No indication of the time interval separating the periods was recognized, so it is problematical whether or not they represent phases of one long period of nearly continuous deformation.

The earliest period of deformation is revealed by the rocks near Spud Mountain and the Indian Hills (pl. 6). This deformation produced tight folds and concomitant foliation, both trending from north to north-northeast. Whether or not any significant faulting occurred during this period is unknown. The later periods of deformation resulted chiefly in distributive shearing, and perhaps small folds. Intrusion of quartz diorite and gabbro occurred after the older of the later periods of deformation.

The later periods of deformation produced two broad zones of distributive movement; one is occupied by the Chaparral volcanics and bounded by the Chaparral and Spud faults, and the other is occupied by the Texas Gulch formation. These two wide "fault" zones divide

the rocks of the Alder group into three blocks. Each block has structural continuity, and this continuity permitted us to propose a tentative stratigraphy for the central block. But attempts at stratigraphic correlation between the rocks in these blocks and fault zones were largely unsuccessful. The stratigraphy of the Alder group is thus in doubt, and much of the structure still remains to be unraveled.

Henceforth, for the purposes of clarity and ease of expression the different planar structures will commonly be referred to as $s1$, $s2$, $s3$, and $s4$—$s1$ refers to bedding; $s2$ to the foliation or cleavage produced during the early period of deformation; $s3$ to the foliation produced during later deformation; and $s4$ to the fracture cleavage and late foliation that locally cuts $s3$ and $s2$.

EARLY PERIOD OF DEFORMATION

Within the map area the early period of deformation produced steeply dipping beds ($s1$), foliation ($s2$), and a mineral assemblage containing actinolitic hornblende. The steeply dipping beds reflect isoclinal or nearly isoclinal folds. The foliation, which formed concomitant with the tight folds, dips 70° or more, and strikes north, north-northwest and north-northeast. We ascribe considerable significance to the local predominance of actinolitic hornblende as the dominant mafic mineral, for retrograde metamorphism apparently accompanied later deformation and altered the actinolitic hornblende to chlorite.

Intersecting foliation clearly separates the early period of deformation from later periods of Spud Mountain in the low hills southwest of Spud Mountain and in the Indian Hills. All the structures in the low hills extending through the alluvial fill of Lonesome Valley northwest of Indian Hills possibly resulted from this early period of deformation. It also is possible that all the structures in the Green Gulch volcanics lying northwest of the Chaparral fault were produced from the stresses of this early period, although the structures are difficult to relate to those in the remainder of the area underlain by the Alder group because of the large movement postulated for the zone bounded by the Spud and Chaparral faults. In other areas underlain by the Alder group, separation of structures clearly related to the early period of deformation from those related to later periods is virtually impossible. In large part, later stresses probably accentuated the existing structures. Commonly two or more slightly divergent cleavages can be distinguished in outcrops, but we were not able to recognize any age difference, although one may exist. It thus appears most probable that the early period of severe deformation established the major structural trends, except those in the Chaparral volcanics, which lie between the Spud and Chaparral faults, and possibly those in the Texas Gulch formation.

LATER PERIODS OF DEFORMATION

The later periods of deformation apparently resulted largely in distributive shearing, except for local areas of small-scale folded cleavage in the purple slate unit of the Texas Gulch formation. These are only local features, formed in response to local conditions. So far as known, no large-scale folds resulted from this later deformation. It does not necessarily follow, however, that the stresses were radically different from those that caused the earlier structures, and both the earlier and later structures may have formed during a long, continuous period of orogeny marked by several surges or peaks when stresses were at a maximum. The differences between the structures produced—that is the difference between (1) isoclinal folds and foliation whose average attitude is nearly parallel to the axial plane of the folds, and (2) wide zones characterized by distributive shear—could be a function of the fabric of the terrane under stress. Any further structural shortening across a terrane consisting of isoclinally folded rocks that exhibit a well-defined, steeply dipping foliation, probably would result in rifts or thrust faults. We ascribe to some such process for the later structures in the rocks of the Alder group.

Separation of the structures formed during the later periods of deformation from those formed earlier is easy in those areas where the two intersect, such as in Spud Mountain, in the Indian Hills, and in the zone occupied by the Chaparral volcanics. Only one structural trend—north to north-northeast—is found on rocks of the Alder group in the rest of the area. Multiple deformation is difficult to prove in these places, but it is equally difficult to disprove. The possibility of multiple periods of deformation will be discussed in the sections on the structures of these areas.

STRUCTURE OF ROCKS BETWEEN CHAPARRAL VOLCANICS AND TEXAS GULCH FORMATION

The block lying between the Chaparral volcanics and the Texas Gulch formation including the Indian Hills and the low hills north of the Indian Hills, so far as we know, is not cut by any major fault zones, so that all units lie in their normal stratigraphic position. The block thus is a structural unit. The chronology of the recognizable structural events consists of an early period of isoclinal folding accompanied by formation of axial plane foliation. A second period of deformation, manifest by distributive shearing, locally produced fractures, breccia zones, and foliated

zones that intersect the older structures. The local shearing of the quartz diorite in the southwest corner of the Jerome area, particularly along the Spud and Chaparral faults, is ascribed to this period. So far as known, this was the last period of sufficiently intense deformation to foliate rocks. Fracture cleavage, joints, scattered occurrences of folded cleavage, and faults marked by gouge testify to later deformation, but the ages of most of these structures are uncertain. In comparison to the earlier periods of deformation, they indicate mild deformation of no great significance.

Although we had little difficulty in dividing the rocks in this block into three formations each of which we further subdivided into lithologic units, the correct order of stratigraphic succession is not entirely certain. The beds are isoclinally folded, so that nearly all bedding dips steeply. Shearing during deformation and recrystallization have largely destroyed bedding and the primary structures indicative of the depositional sequence. In the few places where the tops of beds could be recognized, the probability of isoclinal folding restricts its use to a local area. The trustworthy data on the stratigraphic succession has been summarized on plate 6.

The symmetrical distribution of the formations and the rock units comprising them about an axis in the center of the Iron King volcanics strongly suggest that the major structure of the block is a fold, and the meager data on the direction that beds face suggest a syncline. Cleavage-bedding relationship and sedimentary features indicate that beds face east in the tuffaceous unit of the Spud Mountain volcanics from west of the Shylock mine northward to the Paleozoic rocks. Cleavage-bedding relationship and a flow breccia along the east side of the lava flow west of the Iron King mine suggests that the beds in the Spud Mountain volcanics west and southwest of the Iron King mine face east. Pillow structures in lava intercalated in the tuffaceous unit of the Spud Mountain volcanics southeast of Humboldt (1,267,500 N.; 404,000 E.) indicate westward-facing beds. It appears reasonable that the rhyolitic tuff zone lying about 1,500 feet east of these pillow lavas also faces west, for there is no duplication of the pillow lavas in the interval between the two units. If these rhyolitic tuff beds face west, then probably the bulk of the andesitic tuffaceous unit of the Spud Mountain volcanics lying east of the Iron King volcanics faces west.

The syncline apparently plunges southward at a low angle. The units exposed just east of the Indian Hills are only about 6,500 feet stratigraphically below the youngest beds exposed in the core of the syncline along the southern boundary of the map area; taking into account the difference in altitude between these two points, the average plunge of the syncline is only about 7° S. Along the south margin of the map area, lineation, manifest by mineral streaking on the foliation planes, also indicates a southward plunge for the fold, but it is questionable whether the deformation that produced the fold also produced the lineation, or whether the lineation is due to later deformation.

An anticline lies west of the syncline. It is based on unequivocal determination of the direction that beds face in the breccia of the Spud Mountain volcanics in Ticonderoga Gulch east of the quartz diorite tongue that extends into the southwest corner of the map area, and on poor determination of the direction that beds face near the contact with the tuffaceous unit of Spud Mountain volcanics to the east of the breccia. In Ticonderoga Gulch, beds dip and face westward. Here in two different places determination of the depositional sequence is based on gradation from breccia to fine-grained tuffaceous sedimentary rocks repeated in several beds in succession. Evidence for the eastward-facing limb was given in the preceding discussion on the syncline.

The planar structures in this block consist of $s1$ (bedding), $s2$ (early foliation), $s3$ (later folation, breccia zones, and fractures), and $s4$ (chiefly fracture cleavage). In general, $s1$ and $s2$ are parallel, but discordances mark outcrops scattered throughout the Spud Mountain volcanics. Most of these probably resulted from the intersection of the foliation with the "noses" of folds. $S2$ and $s3$ are discordant in Spud Mountain and in the hills lying southeast of Spud Mountain. Elsewhere $s3$ appears parallel to $s1$ and $s2$, although the extent to which $s3$ is superimposed on $s2$ in these other areas cannot easily be determined. Apparently the early deformation squeezed the rocks so intensely that the foliation, bedding, and fold axis became essentially parallel.

In the Green Gulch volcanics away from the Chaparral fault and in the breccia unit of the Spud Mountain volcanics in the southwest corner of the Jerome area, the bedding ranges in strike from N. 20° E. to N. 45° W., but the general strike is about north (pl. 1). On the southern flank of Spud Mountain and on the hill to the southeast, the strike of the bedding ranges from north to N. 20° E. On Spud Mountain, however, the strike swings from north to N. 45° W., and then to nearly north again, as if to parallel the Spud fault. Bedding as well as foliation makes the swing, which suggests folding of both structures. This folding is not entirely due to drag on the major fault zone bounded by the Spud and Chaparral faults and marked by the Chaparral volcanics, except where it parallels the Spud fault. $S3$ in the vicinity of Spud Mountain strikes N.

45°–65° E. and dips steeply northwestward. It clearly cuts $s2$, and comprises narrow foliated or brecciated zones and in some places, single planes along which shear has occurred. Drag folds occur along the margins of nearly all the $s3$ zones. Their pattern, similar to that along the Chaparral fault, indicates that the rocks on the north side of the $s3$ zones moved northeastward in relation to those on the south side. There is a possibility that any $s3$ zone also had a vertical component as well as the obvious horizontal component. This possibility becomes a probability for the zone occupied by the Chaparral volcanics and for the later movement that occurred in the andesitic tuffaceous unit of the Spud Mountain volcanics and in the western zone of basaltic flows in the Iron King volcanics. For in these three areas, mineral streaking on the foliation planes plunges steeply, and is believed to be the a direction of the structural coordinate system.

In the Indian Hills, foliation ranges in strike from north-northwest to north-northeast, and this range may be due to the presence of both $s2$ and $s3$. If so, the more westward-trending foliation is $s2$, and the more easterly $s3$. In general, actinolitic hornblende is associated with rocks (mostly andesitic breccia) whose foliation has the former trend, and chlorite with rocks (mostly andesitic tuffaceous rocks) whose foliation has the latter.

In both the Indian Hills and the vicinity of Spud Mountain, retrograde metamorphism accompanied the development of the $s3$ planar structure, so that chlorite is the dominant mafic mineral in the $s3$ zones, and actinolitic hornblende is characteristic of the rocks where only $s2$ is present. In other areas where there is only one structural trend, the presence of chlorite as the mafic constituent has been used as evidence for the superposition of $s3$ on $s2$.

The average trend of all the foliation in the rocks that are east of the breccia unit (Spud Mountain volcanics) in Spud Mountain and the Indian Hills strikes about N. 10°–35° E. Although there is considerable variation in the strike of foliation in these rocks, we cannot demonstrate two generations of foliation through intersection and deformation of an earlier foliation by a later one. As was pointed out earlier, this north-northeast structural trend is ascribed to the early period of deformation. But where and to what extent—if at all—subsequent deformation reactivated movement along the preexisting foliation planes is a problem that we have been unable to solve to our satisfaction.

Several lines of evidence suggest the superposition of $s3$ on $s2$ in part of the block east of Spud Mountain. Eastward from Spud Mountain, $s3$ becomes more and more prominent and the rocks become more and more chloritic; simultaneously, however, the angle between $s2$ and $s3$ becomes more acute, and finally $s2$ and $s3$ appear to merge. At this point the rocks are chloritic. If chlorite is indicative of $s3$, which is certainly true near Spud Mountain, the extent of the later period or periods of shearing can readily be determined. Chloritic rocks crop out from about the contact between breccia and tuffaceous facies of the breccia unit of the Spud Mountain volcanics, which is about 1,000 feet west of the Iron King mine, eastward about to the major fold axis near Humboldt. From this line eastward about to the contact between the Iron King volcanics and the tuffaceous unit of the Spud Mountain volcanics, actinolitic hornblende predominates. Chlorite again prevails from here eastward to the fault that bounds the Texas Gulch formation on the west. The rocks also become much more fissile near this fault, indicating that shearing was distributed through a wide zone. As is discussed in a subsequent section, the quartz diorite in the southwest corner of the map area may be younger than the early deformation. Narrow zones of rocks that appear to be foliated quartz diorite dikes occur near the Iron King mine. If these zones are quartz diorite dikes related to the large mass of quartz diorite to the southeast, then the andesitic tuff near the Iron King mine possibly was sheared during a later period of deformation.

Whatever deformation sheared the rocks near the Iron King mine also sheared the rocks as far east as Humboldt, for all have a common, northward plunging lineation (mineral streaking). It is perhaps more than fortuitous that the lineation in the Chaparral volcanics at the latitude of the Iron King mine also plunges northward, nearly parallel to that in the rocks around the Iron King mine. Certainly the lineation in the Chaparral volcanics is related to $s3$. The lineation in the central and western part of the Indian Hills and possibly that north of Highway 89A plunges northward, but that east of the Indian Hills plunges steeply north and south. The rocks east of the Indian Hills, however, were deformed by Paleozoic and later faulting, so that the variation in the plunge of the lineation might be due to this later faulting.

Lineations were observed only in the chloritic rocks. Where actinolitic hornblende is the predominant mafic mineral, the rocks are not fissile and break along the joints rather than along the foliation. The age and significance of the lineation will remain undetermined until the problem can be solved of whether movement occurred along the foliation in the presently chloritic rocks during the deformation that produced $s3$. The assumption that the lineation is related to $s2$ does not fit well with the southward plunge of the major syncline. The average plunge of the lineation is about

60°–70° N. Because the lineation is chiefly streaking of chlorite and sericite, both of which are platy minerals, it probably is the *a* direction of the structural coordinate system. Then, the *b* direction, which is parallel to the fold axes and perpendicular to *a*, would plunge southward at 20°–30°. If the lineation is related to *s*3, there is no reason to expect an agreement between the plunge of the folds and that of the lineation.

Deformation after foliation produced a lineation determined by crenulated foliation and by fracture cleavage. The axes of these crenulations plunge northward and southward at angles ranging from nearly horizontal to 80°. No systematic pattern was recognized and this lineation is not illustrated on any of the maps, except for plate 11, which shows fracture cleavage in the alteration zones associated with the Iron King deposit.

STRUCTURE OF THE ROCKS (CHAPARRAL VOLCANICS) BETWEEN THE SPUD AND CHAPARRAL FAULTS

We believe that the structures in the Chaparral volcanics and the Chaparral and Spud faults together constitute one of the major structures in the Alder group. Deformation in this zone was intense. The Chaparral volcanics and the associated intrusive material in the Spud and Chaparral faults are all finely laminated schist. Rocks in local zones become almost a mylonite and rock units grade into each other by mechanical mixing. The zone in which the Chaparral volcanics crop out probably can best be visualized as a wide fault zone, characterized by extensive distributive shear.

The structures in the Chaparral volcanics, including the Chaparral and Spud faults, trend about N. 40° E., athwart the trend of all the structures to the northward and to some of the foliate structures to the southeast. Toward the northeast, this zone passes beneath the Tertiary Hickey formation. It does not, however, reappear in the vicinity of the Indian Hills, as its strike trend indicates. We therefore believe that the strike of the zone, including the bounding faults, must curve to a more northerly direction (pl. 6) and pass west of the Indian Hills in the area covered by alluvium. It is possible that part or even all Chaparral volcanics might be cut out by faulting between Dewey and the Indian Hills, but it is unlikely that the fault zone, itself, would die out.

Foliation (*s*3) is the dominant structure in the Chaparral volcanics; the foliation strikes northeast, parallel to the faults bounding the formation and parallel to the lithologic units comprising it. The dips of foliation range from 65° W. to 80° E., but west dips are far more abundant. The lineation, chiefly mineral streaking and elongation of pebbles, south of 1,273,500 N., plunges steeply northward. North of this line, however, most of the lineation plunges gently southward, but outcrops sometimes occur in which the lineation plunges steeply northward.

The relationship of lineation to folds was not determined satisfactorily, owing to the multiple deformation of the Chaparral volcanics. The outcrop pattern of the rhyolitic tuff unit in the southeastern part of the Chaparral volcanics suggests a fold, but whether an anticline or a syncline is not known, owing to the lack of data on the direction that beds face. The two limbs of the fold meet at the "nose" near 1,272,500 N.; 386,500 E. From this point northeastward for about 4,000 feet, the rhyolitic tuff crops out as a single band, suggesting a flat plunge. Farther northeastward, the unit divides again with alaskite between the two "limbs." This bifurcation may or may not be due to a reversal of plunge of the fold.

Observations north of 1,275,000 N. established the parallelism between the gentle southward-pitching lineation and the plunge of some minor fold axes. Presumably this lineation is "*b*," and presumably it is related to one of the later periods of deformation. But whether or not these minor folds are related to the larger fold just described is conjectural.

A high-angle, northward-trending fault of uncertain age occurs west of the Iron King mine. It offsets the Chaparral and Spud faults, the granodiorite, and the Silver Belt-McCabe vein, but localized the mineral deposit in the Henrietta mine, which is just south of the Jerome area. It has a horizontal displacement of about 500 feet where it offsets the Chaparral fault and about 100 feet or less where it offsets the McCabe-Silver Belt vein; the east side moved southward relative to the west.

STRUCTURE OF GREEN GULCH VOLCANICS

The Green Gulch volcanics occur only west of the Chaparral fault, so that their stratigraphic position in the Alder group is not known. If the movement on the zone embracing the Chaparral and Spud faults and the Chaparral volcanics is measured in miles, which is a possibility, the Green Gulch volcanics may not be a part of the Alder group.

The area underlain by the Green Gulch volcanics is too small to determine the structural pattern; it is only a "fragment" of the whole, about which little is known. Foliation, bedding, lithologic units, and minor fold axes all trend northward, except where dragged into northeast trends along the Chaparral fault. Foliation and bedding generally dip steeply westward. Foliation, bedding, and the axial planes of minor folds generally are all parallel, although in scattered outcrops bedding

and foliation intersect along a northward-plunging lineation that appears to reflect a north plunge of the minor folds. Scattered determinations of the direction that beds face, though few and inconclusive, suggest that, except for local reversals due to minor drag folds, the formation faces west.

The incomplete structural data in the Green Gulch volcanics imply that the unit is on the east limb of a tight or isoclinal northward-plunging syncline. The pronounced foliation, stretched fragments in the breccia beds, and the parallelism of foliation, bedding, and fold axes indicate strong deformation. Possibly this period of deformation is the same as the early period of deformation recognized elsewhere in rocks of the Alder group.

STRUCTURE OF TEXAS GULCH FORMATION

Like the zone that includes the Chaparral volcanics, the one that includes the Texas Gulch formation can best be visualized as a broad fault zone characterized by distributive shearing. The distributive shearing accounts for the continuity of the lithologic units and structures within the formation, for although units are greatly thinned, they are rarely completely sheared out. The Texas Gulch formation is bounded by faults—each contact with the Spud Mountain volcanics is a surface dividing rocks of discordant structures—but distributive movement also occurred within the formation, for all internal contacts between rock units are mechanical. Some contacts separate rocks with discordant structures; others do not. The abrupt thickening and thinning of the purple slate units attests to the magnitude of the differential movement or rock flowage.

The differences in strike between the Texas Gulch formation, the contiguous formations and the faults separating them are readily apparent from plates 1 and 6. The Iron King and the Spud Mountain volcanics west of the Texas Gulch formation strike more easterly, and are cut out from north to south against the bounding fault. The Spud Mountain volcanics east of the Texas Gulch formation south of 1,280,000 N., strikes more westerly than the Texas Gulch formation, and is cut out against the eastern bounding fault. Along this latter fault north of 1,300,000, the Texas Gulch formation is partly cut out, because of a more easterly strike. About half a mile south of Highway 89A, the entire formation is cut out for 1,300 feet by the junction of the two faults that bound it.

The amount of displacement on the faults bounding the Texas Gulch formation cannot be determined because we do not know the stratigraphic relationship of the Texas Gulch formation to the Spud Mountain volcanics. Mapping of a marker bed, which is not illustrated on plate 1, did show, however, that about 2,500 feet of Spud Mountain volcanics is cut out against the western strand of the Shylock fault between the southern boundary of the Jerome area and a point on the fault 1½ miles to the north.

To obtain a minimum figure for the stratigraphic throw on the western boundary fault a seemingly reasonable assumption must be made: the approximate 7° plunge of the major fold in the block lying between the Texas Gulch formation and the Chaparral volcanics is the average for the entire block. On the basis of this assumption, an imaginary marker bed in the Spud Mountain volcanics will contact the fault at a point at least 10,000 feet higher at the last northern exposure of the fault than at the southern boundary of the Jerome area; or stated in another way, the structural relief measured on the imaginary marker bed is 10,000 feet between the northern limit of the Texas Gulch formation and the southern limit. This figure may be in error by as much as half, that is 5,000 feet. If the Texas Gulch formation underlies the Indian Hills volcanics, which is the closest stratigraphic position to the Spud Mountain volcanics that one can propose, to this figure of 10,000 feet must be added the thickness of the Indian Hills volcanics. This unit has a maximum outcrop width of about 6,500 feet, and because it presumably dips steeply, the stratigraphic thickness is probably not less than 5,000 feet. Adding the two figures together gives a minimum stratigraphic throw of 10,000 feet for the fault bounding the Texas Gulch formation at the point where it crosses the southern boundary of the Jerome area.

Much faulting also occurred within the Texas Gulch formation both between rocks of different lithology and within units of uniform lithology. The elongated mass of andesitic tuffaceous rock enclosed in the Texas Gulch formation in the southern part of the formation and the smaller elongated masses lying near 1,305,500 N.–416,000 E. and 1,328,000 N.–418,800 E. were all detached from the parent formation and squeezed or faulted into their present positions. The fault bounding the east side of the largest or most southerly mass of andesitic tuffaceous rock in places truncates folds and beds. The western fault conforms to the trend of the adjacent rocks, but is marked by a zone as much as 500 feet wide of fine-grained very fissile chlorite schist in which no vestige of the original character of the rock remains. The quartz diorite masses lying near 1,320,700 N.–414,000 E.; 1,341,000 N.–415,500 E.; 1,327,500 N.–418,000 E.; 1,298,000 N.–417,000 E. also have mechanical contacts. Indications of contact metamorphism are lacking in the purple slates alongside

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these quartz diorite masses, and the pronounced cleavage passes uninterrupted from slate into quartz diorite. Evidence clearly indicates that the purple slate and the quartz diorite were sheared simultaneously. Angular discordances between foliation or bedding in purple slate and foliation or bedding in adjacent rhyolitic tuff indicate differential movement between the two rock units; and the thickening and thinning of the purple slate units shows differential movement within and along the margins of lithologically uniform units. The sheared-out middle limb of the fold near 1,276,500 N.–414,700 E. reveals a relative concentration of distributive movement in local zones. But despite the large amount of distributive shearing evident in the Texas Gulch formation, there was no great loss of structural continuity.

In considering the gross structural pattern of the region, the zone defined by the Texas Gulch formation is a single fault zone. Under this concept, the andesitic tuffaceous unit of the Spud Mountain volcanics is in fault contact with the breccia unit. This implies that the stratigraphic throw on the zone is much less than that known to have occurred on either of the two faults bounding the Texas Gulch formation. Two possible explanations occur to us: one, the zone occupied by the Texas Gulch formation is essentially a rift; the other, the zone is a diapirlike structure. The rift theory presumes two separate periods of chiefly horizontal movement: the first period moved the Texas Gulch formation against the andesitic breccia, and the second moved the Texas Gulch formation and the breccia against the andesitic tuff. The chronology of the two periods of movement could be reversed. The diapir theory presumes an "intrusion" or squeezing up of a detached mass comprising the Texas Gulch formation from a lower altitude. Under this latter process, the breccia facies of the Spud Mountain volcanics that lies to the east of the Texas Gulch formation probably moved little, if at all, relative to the andesitic tuffaceous unit of the Spud Mountain volcanics east of the Texas Gulch formation; the presence of hornfelsed Spud Mountain in the quartz diorite east of the Shylock fault supports this theory.

The outcrop pattern of the purple slate and rhyolitic tuff in the Texas Gulch formation strongly indicates isoclinal folds, but the irrefutable evidence of much distributive shearing within the formation indicates a second process that could produce a similar pattern. The symmetry of the pattern of outcrops alone almost compels acceptance of isoclinal folding as the major process, but each individual case must be considered in view of the structures in the immediate vicinity and in view of the outcrop pattern near the particular exposure. Isoclinal folds were demonstrated by walking out conglomerate beds around the plunging noses of two folds, but it seems probable that some isolated masses of rhyolitic tuff were severed from the parent rock and moved to foreign positions. Such masses have lost their structural continuity; they are not too common.

The recognition of any individual structure as an anticline or syncline depends on local determination of the direction beds face. In many places these were not recognized, so the nature of the local fold was surmised from that of adjacent folds and from the outcrop pattern in relation to the lineation. Although lineation in the Alder group is an unreliable indicator of plunge of folds, owing to a possible second period of unrelated shearing, we proved that fold axes and lineation are parallel in the southern outcrops of the Texas Gulch formation; and we used similar interpretations elsewhere because the outcrop pattern indicates isoclinal folds, that is, there does not appear to be any significant loss of structural continuity. By using available lineation and top determinations and by extrapolating from these to other areas where they were sparse or absent, we were able to suggest a fold pattern, which is illustrated in plate 6. The plunge of the fold apparently oscillates from north to south throughout the formation (pl. 6), so that the structural relief between the northern and southern limits of the formation in the map area is probably small. This is in contrast to the progressive increase in structural relief from north to south in the block lying between the Texas Gulch formation and the Chaparral volcanics.

In the Texas Gulch formation the pronounced foliation essentially parallels the axial planes of the folds, except at the "noses" of the folds. Of all the rocks in the Alder group, the megascopic foliation in the purple slate can be seen most clearly to consist of a family of intersecting cleavage planes. Lineations on the cleavage planes are common; they consist of intersection of bedding and cleavage, intersection of two cleavages, and small puckers or wrinkles of uncertain origin.

The purple slate is particularly susceptible to later cleavage deformation, as shown by the many foliation symbols on plate 1 athwart the regional trend. Around the margins of the large block of quartz diorite, centering about 1,298,000 N.; 417,000 E., later deformation folded the older cleavage and induced a local new cleavage approximately parallel to the axial planes of the new folds.

The presence of sheared masses of quartz diorite enclosed in the Texas Gulch formation suggests that this formation was sheared during the youngest known period of Precambrian deformation.

STRUCTURE OF THE ROCKS EAST OF TEXAS GULCH FORMATION

The rocks east of the Texas Gulch formation consist of a wedge-shaped mass of Spud Mountain volcanics bounded by the eastern and western strands of the Shylock fault. North of the vicinity of 1,280,000 N., rock has been sheared out to a very fissile schist in which relict structures and textures are limited to widely scattered outcrops. South of this transition zone, foliation abruptly becomes less pronounced, and the veil produced by the regional metamorphism gradually lifts, revealing abundant relict textures and structures. In general, beds strike N. 20° E., and dip from 60°–80° W. and on the basis of graded bedding, beds face west, with the exception of the middle limb in the small draglike fold adjacent to the Texas Gulch formation.

SHYLOCK FAULT

The Shylock fault separates the Ash Creek and Alder groups. Unbroken Paleozoic rocks that cover the trace of the fault north of 1,341,000 N.; 419,000 E. show that the movement was wholly Precambrian. Because the Shylock separates rocks of diverse lithology and structure whose stratigraphic relationship is unknown, it is assumed to be a major fault.

The Shylock fault crops out from the mantle of Paleozoic rocks in the northern part of the Jerome area to the southern boundary. It strikes north and dips chiefly westward at high angles as far south as near 1,300,000 N.; 418,000 E., where a general zone of brecciation rather than any single plane or narrow zone marks the fault. This wide zone of brecciation may be due to splitting of the Shylock into two strands: the eastern strand trends a few degrees east of south and, near the southern boundary of the Jerome area, dips eastward at about 65°; the western strand trends a few degrees west of south and dips from vertical to 60° W. It also continues southward past the southern boundary of the Jerome area.

Over the northern part of its trace, the Shylock fault separates the Ash Creek and Alder groups. Here the fault is sharply defined, and marked by springs and by gouge derived from purple slate, which may be due to later movement. Farther south, the Shylock fault separates the quartz diorite from purple slate of the Texas Gulch formation. Here gouge is not common, but a straight contact, the pronounced cleavage in the slate alongside the quartz diorite, and the absence of contact metamorphism all compel a mechanical contact. The eastern strand separates intrusive rocks, chiefly quartz diorite, and hornfels on the east from the Spud Mountain volcanics on the west; the western strand separates the Spud Mountain volcanics from Texas Gulch formation on the west.

The displacement on the Shylock fault may decrease remarkably from north to south. The northern part of the Shylock divides the Ash Creek and Alder groups, and the stratigraphic throw is at least 20,000 feet, the approximate thickness of each group. The throw is probably much greater judging by the differences in metamorphic structures between the groups, and by each group being confined to its side of the fault with neither top nor bottom of either group exposed.

To the south, displacement on the combined western and eastern strands of the Shylock is somewhat less than the thickness of the Alder group. The western strand, which was considered in the section on the structure of the Texas Gulch formation, separates rocks of the Alder group throughout most of its length. The eastern strand separates Alder group (west wall) from quartz diorite (east wall) throughout most of its length. Near 1,283,000 N.; 418,000 E., however, hornfelsed Spud Mountain volcanics forms the east wall, and here the stratigraphic throw on the combined eastern and western strands must be somewhat less than the thickness of the Alder.

Large-scale movement along the eastern strand of the Shylock is suggested perhaps in the discordance between the trend of the primary flow banding in the quartz diorite and the trend of the fault. This implies that the margin of the quartz diorite was removed by the fault. This discordance, however, may be interpreted differently, for the planar structure in the quartz diorite adjacent to the crosscutting contact with the dacitic flows of the Grapevine Gulch formation is nearly as discordant. The hornfelsed Spud Mountain and a border facies of the quartz diorite just north of the hornfels are both marginal features which oppose a large-scale movement.

This discrepancy in the probable stratigraphic throw between the northern and southern parts of the Shylock fault could be explained by an older period of faulting that occurred before emplacement of the quartz diorite. North of the quartz diorite, the Shylock fault may mark the trace of an older fault; but to the south, this older fault may have been engulfed by the quartz diorite, and only the stratigraphic throw of the post-quartz diorite movement can be deciphered.

The gouge of crushed purple slate for several miles along the fault south of Highway 89A deserves consideration because it indicates localized younger movement. Nowhere in the Alder group did Precambrian deformation prevail under temperature and pressure conditions that permitted gouge; instead zones of more pronounced fissility or foliation formed. We thus ascribe the origin of this gouge to later deformation. Two periods of Paleozoic and later faulting displaced rocks along the western side of the Black Hills; either provides satisfactory conditions for the generation of gouge. In order not to have offset the Paleozoic rocks

north of Highway 89A, the displacement on the Shylock during this later movement must have been small and localized whether an independent period or a part of a major period of faulting that elsewhere offset the Paleozoic rocks.

STRUCTURE OF THE PALEOZOIC AND CENOZOIC ROCKS

The mild stress recorded by the Paleozoic and Cenozoic rocks in the Jerome area resulted in two distinct types of deformation: (1) simple uplift marked by gentle tilting, and (2) faulting accompanied by mild to severe tilting. Apparently the rocks yielded by faulting and tilting rather than folding; the only major structure resembling a fold is the gentle arching of the Black Hills from north to south (fig. 5), ascribed to differential uplift along the boundary faults of the Black Hills. The writers recognized several periods of faulting, in places accompanied by tilting, which were later than Paleozoic. The first period occurred before deposition of the Hickey and is referred to Late Cretaceous and early Tertiary time. The second followed the accumulation of the Hickey formation, and the third followed the accumulation of the Verde formation. Some faulting also occurred during accumulation of the Hickey, as shown by faulted lower beds of gravel overlain by undisturbed lava.

OLDER STRUCTURE

The older structures consist of gentle tilting and faulting of Late Cretaceous and early Tertiary age. The exact degree of tilting cannot be determined, owing to later faulting, but it must have been only a few degrees. That the Paleozoic rocks dipped gently north at the time the Hickey formation accumulated, is shown by the southward overlap of the Hickey over the beveled edge of the Paleozoic rocks on to the Precambrian. Perhaps this tilting accompanied the older period of faulting.

The oldest post-Paleozoic faults displace the Paleozoic rocks but not the Tertiary Hickey formation; a more precise age cannot be determined here. Deformation was common throughout much of Arizona, particularly in the southern part of the State, during Late Cretaceous and early Tertiary time. More germane to the Jerome area, is the evidence cited by Babenroth and Strahler (1945, p. 149) that the East Kaibab monocline in the Grand Canyon area clearly formed during Late Cretaceous and early Tertiary time. Some faulting was associated with the flexing of the south end of the East Kaibab monocline, and this has strengthened our tentative dating of the older faulting involving Paleozoic rocks in the Jerome area as of the same age (Late Cretaceous and early Tertiary).

Faults that displace Paleozoic rocks but not the late Tertiary Hickey formation occur in the area north of Mescal Gulch and east of the Verde fault. Many younger faults, including the Verde, may have moved earlier also, but exposures do not permit measurement of displacement of the Tertiary rocks independently from that of the Paleozoic rocks at the same locality. The stratigraphic throw on all these older faults is small. On the small fault at 1,373,500 N.; 436,600 E., it is less than 50 feet and on the fault at 1,379,700 N.; 435,500 E., it is about 150 feet. The fault about 2,300 feet north of the portal to Hopewell tunnel (1,373,000 N.; 441,800 E.) passes under lava of the Hickey formation at a point where the stratigraphic throw on the Paleozoic rocks is about 450 feet. Farther north (1,377,500 N.; 437,500 E.), this same fault cuts the Hickey formation; here the vertical separation on the base of the Hickey is about 250 feet. The earliest movement on the Bessie fault (1,363,000 N.; 446,000 E.), one of the faults bounding the east side of the Black Hills, occurred between Mescal and Deception Gulches during this period of deformation. Similar to the other faults described, the east side of the Bessie moved down relative to the west. The stratigraphic throw, however, cannot be determined, because the Bessie moved twice more (described in the section on the Bessie and related faults), and no measurement of the movement could be made for the intermediate period of faulting.

On the northwest flank of Mingus Mountain, data on this older period of faulting are more difficult to obtain because of recurrent movement (Coyote fault) of Tertiary age along the same trace. On the west side of the older fault, lava of the Hickey formation rests on Precambrian rocks, but on the east side, the lava flow rests on Paleozoic rocks. We believe that the lava flows are correlative, therefore faults earlier than Hickey age are necessary to explain this simple relationship. This faulting occurred long enough before accumulation of the Hickey for the entire Paleozoic section to be eroded from the upthrown (west) side of the fault. Even a crude approximation of the stratigraphic throw on this old fault is difficult to make, for during the recurrent movement in the Tertiary, the west side was dropped. On the east side of the trace of the older fault, all or virtually all the Redwall and underlying Paleozoic rocks crop out. Therefore, a minimum stratigraphic throw equivalent to the sum of the

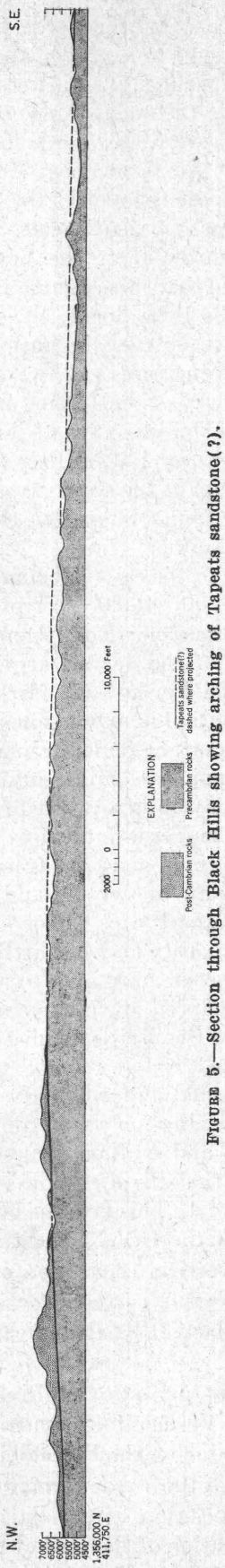

FIGURE 5.—Section through Black Hills showing arching of Tapeats sandstone(?).

Paleozoic section below the top of the Redwall (about 800 feet) is a safe assumption.

This older fault extends beyond the northern boundary of the Jerome area. To the south, it extends at least to Highway 89A, but its extent south of that point is uncertain owing to the southward overlap of the Hickey over the beveled edge of the Paleozoic rocks onto the Precambrian rocks.

The validity of this early movement is well established about 2 miles north of the Jerome area, where the Tertiary rocks are not displaced, and the older fault dropped Redwall limestone on the east against the Martin limestone on the west. Here the base of the lava west of the fault forms a gently dipping plane that projects into the base of the lava east of the fault, so that correlation of the two lava flows appears indisputable. Once this correlation across the fault is accepted, it is probable that all the lava west of the fault correlate with those to the east, because west of the fault, one lava patch is only slightly offset from the next from a point north of the Jerome area southward to the Indian Hills.

One minor strand of this older fault is exposed in the northwest corner of the Jerome area at the mouth of Martin Canyon (1,363,000 N.; 409,000 E.). The dip is steep and the maximum stratigraphic throw (east side dropped) north of Martin Canyon is about 35 feet. South of Martin Canyon, the throw increases to about 400 feet at the southernmost exposure. To the south the fault is covered by Quaternary gravel, but it may strike into the Coyote fault.

YOUNGER STRUCTURE

The greatest deformation in the Jerome area after the Precambrian took place after the accumulation of the Hickey of probable late Tertiary age and before the deposition of the older Quaternary gravel and the gravel facies of the Verde formation. Normal faults of this period elevated the Black Hills to form the present drainage system and topography. They undoubtedly are a part of the block faulting of the Basin and Range province, for the Jerome area lies astride the union of the Mountain region—which is part of Gilbert's Basin and Range province—and the Colorado Plateau, but this faulting severed it from the plateau, and attached it to the Mountain region.

The faults are distributed in two zones, one along either front of the Black Hills. In both places, the faults are multiple; along the eastern front they are known as the Verde fault zone and those along the western front, as the Coyote fault.

VERDE FAULT ZONE

The Verde fault zone is better exposed and more complex than the Coyote fault. Its stratigraphic throw is more than 2,000 feet. The Verde fault zone comprises a main strand, the Verde fault, and many parallel and subparallel subordinate faults. The Verde formation covers the fault zone south of 1,347,000 N., except for several strands along which movement was renewed after the deposition of the formation. Chief among these are the Verde and Bessie faults. North of 1,347,000 N. erosion has stripped much of the Verde formation from the underlying Hickey formation, and the fault zone is well exposed.

The Verde fault zone and its individual components strike northwestward and dip steeply to moderately either east or west. Most faults are straight for short distances, but those more than a few thousand feet long curve gently. The most arcuate of these is north of Jerome near the northern boundary of the Jerome area (1,381,500 N.; 437,500 E.). This fault curves through about 45°, and the change in strike is accompanied by a change in dip from steep east to 40° W.; this may be an intersection of two faults rather than a single one. Most of the faults dip east, but some dip west; generally, the dips are high, probably near 75°, but they range from about 35° to vertical. Faults with moderate dips—about 45°—are scattered sporadically throughout the fault zone. The Verde fault has this dip from a point about 1½ miles southeast of Jerome to the vicinity of Black Canyon, an appreciable part of its total outcrop.

The displacement on the Verde fault zone took place on many strands; as many as 13 show in a single cross section, and only those with significant throw were mapped. All faults, except a short segment of one fault, are normal. All the faults of large throw dip east, and the algebraic sum of the stratigraphic throw of the zone ranges from a few feet to more than 2,000 feet. The structural relief measured on the Tertiary volcanic rocks from the summit of Mingus Mountain to the Verde Valley resulted mostly from displacement on the Verde fault zone, though the homoclinal eastward dips are supplementary. North of the Jerome area, the Verde fault zone passes abruptly into a monocline that accounts for the entire structural relief of the northern flank of the Black Hills. The eastward-dipping normal faults increase the structural relief due to the eastward-dipping beds, but the westward-dipping normal faults decrease it. This is well illustrated near Bakers Pass, where the aggregate stratigraphic throw of the westward-dipping faults is about 200 feet greater than that of the eastward-dipping faults. The overall picture of displacement on the Verde fault zone is like that described for the Rhine graben and reproduced by Cloos (1930) in experiments with wet clay.

The stratigraphic throw on the Verde fault zone is difficult to measure, for the Verde formation masks much of the zone. North of about 1,347,000 N., however, erosion has exposed enough of the Verde fault zone to permit approximate measurements. Figure 6 shows the stratigraphic throw along the Verde fault, but it must be remembered that these figures are only approximate. There can be little doubt, however, of the two salient points shown by the illustration: (1) that the stratigraphic throw increases southward from zero near the northern boundary of the Jerome area to more than 2,000 feet south of Jerome (fig. 6), and (2) that the stratigraphic throw near Bakers Pass is in the opposite sense; that is the sum of the stratigraphic throw on the westward-dipping faults is greater than that on the eastward-dipping faults.

The Verde fault, or the eastern boundary fault of the Black Hills, merits individual description. The north end is near Bakers Pass, north of Jerome, and it extends southeastward out of the Jerome area. Although the displacement on the Verde fault farther south is unknown, the physiographic expression indicates that it continues as an important fault for many miles. From its north end to a point 1½ miles south of Jerome and from Black Canyon southeastward to where it leaves the Jerome area, the Verde fault dips 50°–70° E., and only gentle undulations mark its trace. Between the point 1½ miles south of Jerome and Black Canyon, however, the Verde fault dips 45° or less, and its trace is sinuous. The sinuosity may be partly due to changes in strike, but we ascribe most of it to moderate or low-dipping faults produced during recurrent movement after deposition of the Verde formation, because the Verde fault inclines at low to moderate angles only where the Verde formation forms one wall or both walls.

Underground openings on the Verde fault at Jerome reveal slickensides and mullion structure raking 70°–80° S., indicating that the direction of relative movement plunged steeply. This is borne out further by the relative position of the U.V.X. and the United Verde ore bodies. The vertical separation on the relatively flat-lying Paleozoic and Tertiary rocks, and the physiography of the Black Hills all indicate large dip-slip component.

In addition to the period of faulting just described, movement on the Verde fault occurred during three other periods: (1) during the Precambrian, (2) following the deposition of the Verde formation, and (3) possibly after the deposition of the Paleozoic rocks and before the accumulation of the Hickey formation (Late

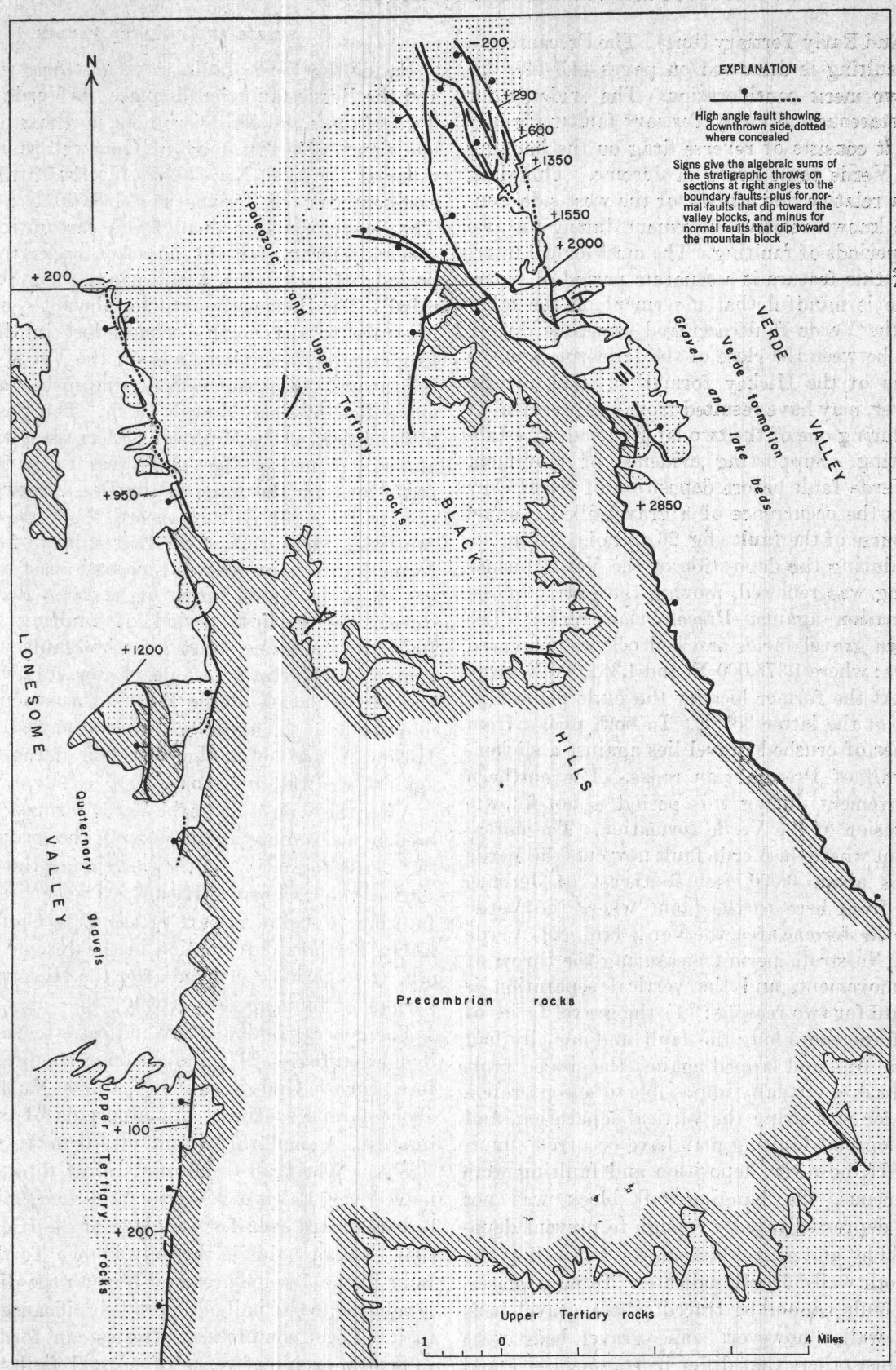

FIGURE 6.—Structural map showing Tertiary displacement in the Black Hills. Figures indicate sum of vertical separation along sections at right angles to boundary faults. Vertical separation on westward-dipping faults deducted from eastward-dipping faults and the proper sign is given to the difference.

Cretaceous and Early Tertiary time). The Precambrian period of faulting is discussed on pages 147–149, but the other two merit consideration. The evidence for the Late Cretaceous and Early Tertiary faulting is only postulated; it consists of reverse drag on the hanging wall of the Verde fault north of Jerome. This drag argues for a relative downthrow of the west side, contrary to the known relative movement during the two established periods of faulting. The most logical interpretation of this feature is a separate period of movement, if one is mindful that movement on the other strands of the Verde fault occurred sometime during the interval between the close of the Paleozoic and the accumulation of the Hickey formation. The reverse drag, however, may have resulted from local reversal of movement during one of the two later periods of established faulting. Supporting evidence of movement along the Verde fault before deposition of the Hickey formation is the occurrence of a gravel-filled channel along the course of the fault (fig. 26 and pl. 1).

After or during the deposition of the Verde formation, faulting was renewed, moving the gravel of the Verde formation against Precambrian rocks. The fault between gravel facies and bedrock was observed in two places: where 1,375,000 N. and 1,351,000 N. cross the fault. At the former locality the fault plane dips 45° E., and at the latter 35° E. In both places from 6 to 12 inches of crushed gravel lies against a slickensided footwall of Precambrian rocks. The northern limit of movement during this period is not known, owing to erosion of the Verde formation. The northernmost point where the Verde fault now cuts the Verde formation is about 6,000 feet southeast of Jerome. Southward from here to the point where the Verde fault leaves the Jerome area, the Verde fault cuts Verde formation. No strata permit measuring the throw of this later movement, and the vertical separation is indeterminate for two reasons: (1) the gravel facies of the Verde formation along the fault undoubtedly had a high initial dip, and lapped against the eroded fault scarp, so that it is virtually impossible to select a reference plane for measuring the vertical separation, and (2) deposition and faulting may have occurred simultaneously. If, however, deposition and faulting were contemporaneous, the hanging-wall block was not everywhere depressed rapidly enough to prevent deposition of cobbles and gravel beds on the footwall as the gravel on both walls clearly indicates. In most places, the Verde fault cannot be traced where gravel beds form both walls; however, some gravel beds were faulted against others that differ in fragment size and in relative abundance of rock types, and here the trace is well marked.

BESSIE AND RELATED FAULTS

Beside the Verde fault, there are three small faults and the Bessie fault that displace the Verde formation. Two of the small faults crop out northeast of Jerome; one about 1,000 feet north of Hopewell tunnel and the other at 1,376,000 N.; 439,500 E.; the third crops out east-southeast of Jerome at 1,363,000 N.; 449,500 E. The Bessie, which is about 1 mile east of Jerome, is of special interest, for the adjacent rocks record three periods of movement, two periods the reverse of the third. The Bessie crops out for about 1½ miles before its trace to the north becomes lost in the uniform lithology of the gravel facies of the Verde formation. Southward, the displacement diminishes and apparently dies out near Mescal Gulch. The Bessie dips at high angles; where Highway 89A crosses, it is vertical, and at a point 1,500 feet northwest it dips 80° E. The rocks forming the walls of the Bessie between Deception and Mescal gulches record the three periods of faulting. Here west of the Bessie, lava of the Hickey formation rests on Martin limestone; east of the fault, gravel of the Verde formation rests on Redwall limestone. The earliest period of faulting moved the Redwall limestone, east of the fault, down and against the Martin limestone. Later, following erosion and accumulation of the Hickey formation, recurrent movement in the opposite direction moved lava of the Hickey formation on the west side of the fault down against Redwall limestone.

After the deposition of the Verde formation, the fault became active once again, moving the gravel facies of the Verde formation on the east side of the fault down against the lava of the Hickey. The earliest period of faulting probably is part of the Late Cretaceous and Early Tertiary deformation in the Jerome area. The other two periods occurred after the Hickey and Verde periods of faulting respectively.

The general fault pattern indicates tension, relieved by normal faults. The indication of compression therefore near the Gadsden mine (1,361,500 N.; 446,000 E.), about 1 mile southeast of Jerome, is of considerable interest. A small thrust fault attests to the compressive stresses. The fault trends northward, dips 30° W., and moved the Paleozoic Martin and Redwall limestone beds eastward over Tertiary lava of the Hickey formation. Because this is the only known Tertiary thrust fault in the Jerome area and because the displacement is believed to be small, no special significance is attached to it. Local compressive stresses can form from the interplay or interference of normal faults where the regional stresses are tensional, and we ascribe to such an origin.

COYOTE AND RELATED FAULTS

The Coyote fault crops out along the west side of the Black Hills from the Indian Hills northward to the boundary of the Jerome area. For about 6 miles south of the Indian Hills, Quaternary gravel covers the trace of the fault, but presumably the same fault reappears at about the latitude of the road from Dewey to Cherry.

The Coyote fault is multiple: two strands occur in the Indian Hills, three, or possibly four, en echelon strands occur to the north, and three outcrops of lava faulted against gravel indicate two additional strands south of the road to Cherry. The fault zone and the individual strands comprising it strike from north to north-northwest. The dips are high, and where measured, dip westward.

After the accumulation of the Hickey formation, movement along the Coyote and Verde faults elevated the Black Hills. The base of the Hickey provides the only reference plane to measure the displacement on the Coyote fault. The approximate vertical separations at other points along the fault are shown in figure 6. They increase from about 200 feet at both the northern and southern boundaries of the Jerome area to about 1,200 feet in the Indian Hills, which is about half the maximum vertical separation on the Verde fault zone.

Several faults offset the gravel facies of the Hickey formation but not the overlying lava in Martin Canyon in the northwest corner of the Jerome area. Other similar faults may occur elsewhere, but would be difficult to recognize in the Hickey owing to poor outcrops, to original lensing of the gravel and lava, and to the irregularity of the lower surface. These faults are small; the maximum vertical separation is about 125 feet. But they do point out contemporaneous deformation and accumulation.

OTHER FAULTS

Some faults in the Jerome area displace Paleozoic rocks in areas where no Tertiary rocks are present, so that their age in relation to the Tertiary rocks is unknown. Such faults occur north of Jerome as part of the Verde fault zone, near Jerome, and in the central part of the Jerome area on the ridge northeast of Gaddes Canyon. The Hull fault appears to be the most prominent feature near Jerome, but where it cuts Paleozoic rocks it contains a basaltic dike, which suggests a pre-Hickey age. Most of these faults are probably post-Hickey, for faults of that age are by far the more abundant.

Southeast of Onion Mountain (pl. 1; 1,295,000 N.; 467,500 E.) two faults displace the Hickey; they are probably contemporaneous in part with the Verde and Coyote faults. Presumably other faults of the same age exist in the quartz diorite, but they were not recognized. One such fault is indicated by the difference in altitude between the Tapeats(?) at 1,317,000 N.; 452,500 E. and at 1,313,500 N.; 453,000 E. At the former point, the altitude of the Tapeats(?) is 6,350 feet, and at the latter point, it is 5,850 feet. Considering their proximity, the difference in altitude is too great to be explained by the regional dip of the Tapeats(?), as is shown by figure 5. A fault obscured by the uniformity of the quartz diorite is the favored explanation.

The remnants of Tapeats(?) at 1,292,500 N.; 471,750 E. and on the tops of many of the hills in the Cherry district permit reconstruction of the regional dip of the Paleozoic rocks in the Cherry district. The section in figure 5 shows a gentle arch, from north to south, of only a few degrees maximum dip. The maximum closure shown by figure 5 is about 1,200 feet, in the central region of the Black Hills. This is about the same structural relief as was produced by the late Tertiary deformation. This coincidence indicates a relationship between the gentle north-south arch in the Paleozoic rocks and the later Tertiary differential uplift, chiefly along normal faults.

PHYSIOGRAPHY

The major physiographic feature in the Jerome area is the uplifted Black Hills, bounded by eroded fault scarps on the west and east margins. The northern summit area of the Black Hills is mesalike in character because of the essential horizontal position of the Hickey formation. Beneath the lava, in the cliffs and intervening gentler slopes of the Paleozoic rocks, the scenery resembles the topography on the Colorado Plateau. In the southeastern part of the Mingus quadrangle, buttes and mesas capped by lava flows dominate the landscape.

The Ash Creek terrane, south of Mingus Mountain, is very rugged, cut by deep canyons, but to the south in the quartz diorite terrane, the country is much more open. Possibly the more rapid weathering of the quartz diorite as compared to the more resistant Ash Creek group is an important factor in the different topography.

Drainage in the Black Hills is radial; to the north in the Clarkdale quadrangle, the streams drain northward to the Verde River. In the Jerome area, the eastern drainage flows to the Verde River in the Verde Valley. On the west side of the Black Hills, the drainage flows to Granite Creek in the Prescott quadrangle and to the Agua Fria River. Southward drainage from Mingus Mountain flows to tributaries that eventually join the Agua Fria River. The streams in the area are essentially consequent, in that their courses were de-

termined by the original surface slopes. Tex Canyon, a western tributary to Ash Creek is an exception; its course parallels the structure of weaker beds in the Grapevine Gulch formation. The south wall of Tex Canyon consists of more resistant hornfels in the Grapevine Gulch formation adjacent to the quartz diorite. Grapevine Gulch for about a mile of its course (423,000 E.) follows a fault zone separating quartz diorite from Grapevine Gulch formation. The topography suggests that Grapevine Gulch captured the headwaters of Texas Gulch along the fault zone.

The rhyolitic tuffaceous unit of the Texas Gulch formation is more resistant to erosion than the associated purple slate, and 2 miles east of Humboldt, a north alinement of ridges and peaks reflects this. Spud Mountain to the west of the Iron King mine is an example of the greater resistance to erosion of the andesitic breccia unit as compared to the other units of the Spud Mountain volcanics.

The Verde Valley to the east and Lonesome Valley to the west of the Black Hills are essentially dropped blocks bounding the Black Hills horst. But both valleys are sites of active erosion and very little detritus is being deposited in them except as pediment veneer or temporary riverwash deposits.

Lonesome Valley is underlain by sedimentary rocks of the Hickey formation in which widespread pediments were formed. Locally, a pediment is cut on the Alder group, particularly in the area bounded by 1,320,000–1,330,000 N.; 410,000–417,000 E. A second pediment on Precambrian rocks is exposed west and southwest of Coyote Springs Ranch (1,349,000 N.; 407,000 E.).

The physiography of the region indicates that the earlier drainage of Lonesome Valley was northwestward to Granite Creek, which flows northward to the Verde River (Prescott and Paulden quadrangles). The pediments in Lonesome Valley, capped by reddish Pleistocene gravel, apparently formed while the drainage was controlled by Granite Creek. By headward erosion, the southward-flowing Agua Fria River captured the southern drainage in Lonesome Valley and tributaries of the Agua Fria have destroyed the pediment surface in the southwest corner of the Mingus Mountain quadrangle, exposing the Hickey formation. North of 1,290,000 N. the pediment surface is dissected, especially toward the west near the Agua Fria River. The Hickey formation is exposed in stream gulches, beneath the pediment gravel veneer, and to the north (1,320,000 N.) rocks of the Alder group are exposed in the gulches. North of the Jerome highway, the pediment is little dissected.

Along the west margin of the Mingus Mountain quadrangle from Indian Hills northward, Precambrian rocks rise in hills above the pediment surface. These hills may be "islands" or perhaps topographic highs on the Tertiary pre-Hickey surface that were buried by the Hickey and later exhumed during the period of pediment formation. Because the pediment surface carved from the Precambrian rocks is east of the west margin of these hills, it is probable that a northwestward-trending fault is present southwest of Indian Hills, dropping the Hickey against the Precambrian rocks. The pediment on the Precambrian rocks would represent an eastward extension of the pediment carved from the Hickey, and related to the probable fault southwest of the Indian Hills.

The physiographic history of the Verde Valley is comparable to that of Lonesome Valley, except that the pediments are carved from the Verde formation, which is younger than the Hickey. Pediments are cut on at least three levels, which are capped by a veneer of reddish gravel. The highest pediment is exposed north of Oak Wash near the Verde fault (1,344,000 N.; 453,000 E.). A widespread pediment, about 200 feet lower in altitude, lies south of Oak Wash; it is dissected and the Verde formation is exposed in the gulches. A lower pediment, near the Clemenceau airport, is less deeply dissected, and the Verde formation is not exposed.

It seems clear that the Verde River has been actively eroding since the Verde formation was deposited; presumably the Verde River eroded through the barrier which caused accumulation of the Verde, and the active down-cutting into the Verde resulted. The pediments at different levels between the Black Hills and Verde River are natural products of erosion in the Verde Valley, controlled by the temporary base levels established by the down-cutting Verde River. Pediments at other levels are to be expected, and pediments at higher altitude will be destroyed by those forming at lower altitudes.

Adjacent to the Verde River at the town of Cottonwood, terraces overlain by waterworn gravel represent a slightly higher course of the Verde River that has been preserved.

ORE DEPOSITS

HISTORY[4]

Centuries ago, Indians mined the oxidized copper ore at the cite of the United Verde mine at Jerome, using the colored rock for personal adornment and as a dye for their blankets. Traces of old dumps, shafts, and tunnels were found at the beginning of active ex-

[4] Much of the history of the mining in the Jerome area has been obtained from the following sources: Rickard (1918); Lindgren (1926, p. 61–63; 79–81); Young (1930); Heineman (1938); Reber (1938).

ploration in 1882. Stone hammers and other stone implements were uncovered in the old workings and parts of some of the veins were stoped (Hamilton, 1884, p. 185).

During the early Spanish exploration of New Mexico and Arizona, two explorers, Antonio de Espejo and Marcos Farfan de los Godos, visited the mines at Jerome (Bartlett, 1942). Espejo, in 1582, with 14 companions headed an expedition into New Mexico to rescue two friars who had remained behind from the Rodriguez expedition the previous year. After learning that the friars had been murdered, the Espejo expedition turned to prospecting and in due time arrived at the Hopi villages in northern Arizona, where Espejo learned of the mines to the west. Espejo with four others and Hopi guides visited the copper deposits, arriving there on May 8, 1585, but left the following day.

In 1595, Juan de Onate was given a contract for the conquest and settlement of New Mexico but it was not until 1598 that he started his expedition. When he arrived in the Hopi villages, he learned of the mines to the west and sent his Captain of the Guard and of the Horses, Marcos Farfan de los Godos, with eight companions and Hopi guides to visit the mines. Farfan arrived at the Jerome deposits on November 24, 1598, and found an old shaft, three estados (16½ feet) deep, and a large dump. Farfan and companions located many claims for themselves and companions who had remained behind in the Hopi villages, as they were impressed with the richness of the deposit.

The Spaniards never worked the deposits and no other record is available concerning the discovery or mining of ore minerals until 1863 when a party of prospectors under the leadership of Joe Walker discovered gold in Hassayampa and Lynx Creeks in the Prescott region (Hamilton, 1883, p. 47).

The oldest mine in the area described in this report is the Silver Belt, located in the Big Bug mining district, along the Silver Belt-McCabe vein, west of the Iron King mine (pl. 1). The Silver Belt mine was located about 1870, and rich silver ore was mined from 1870 to 1880. The ore was freighted to Ehrenberg on the Colorado River and shipped by boat to San Francisco. The mine was reopened in 1906, but no ore was found. The McCabe-Gladstone mine was located on the south end of the Silver Belt-McCabe vein and mining of gold-silver ore started about 1880. The mine was operated continuously from 1898 to 1913 by the Ideal Leasing Co.

The Verde Valley east of Jerome was first settled by white men in 1865 when a group of pioneers from Prescott located land near the Verde River and started farming. It is reported that in 1875, U. S. Army scouts from Fort Whipple at Prescott, discovered mineral showings near Jerome, and in February 1876, John O'Dougherty, Edward O'Dougherty, and John D. Boyd located claims on outcrops that later became the property of the United Verde Copper Co. M. A. Ruffner, one of the pioneers from Verde Valley, did considerable prospecting, and in June 1876 he and Angus McKinnon located the Eureka claim, south of the present pit of the United Verde mine, and in June 1877 they located the Wade Hampton claim which is in the west center of the present United Verde pit. Ruffner, George McKinnon, and Angus McKinnon sank a 45-foot shaft on one of these claims which showed good ore. Other prospectors were attracted to the area and much of the ground near the Eureka and Wade Hampton claims was located, later to be purchased by Charles Lenning, a wealthy Philadelphian.

Transportation was difficult in 1878 as the nearest port for water transportation was on the Colorado River, about 200 miles to the southwest, and the nearest railroad terminal was at Abilene, Kans. In spite of these difficulties, F. A. Tritle, Governor of the Territory of Arizona, who was living in Prescott, then the capitol, became keenly interested in the possibilities of exploiting the Wade Hampton and adjoining claims. He had been manager of the Yellow Jacket mine on the Comstock Lode in Virginia City, Nev., and had much experience in mining. Tritle supported F. F. Thomas in an effort to interest eastern capital in organizing a corporation to develop the claims in the Verde mining district. James A. Macdonald and Eugene Jerome (the town of Jerome was named for him) of New York provided the financial assistance to organize the United Verde Copper Co. in 1882, which purchased 12 claims including those held by Ruffner, the McKinnons, and Lenning. Macdonald was named president and Jerome the secretary-treasurer of the company.

Lenning sold his claims to the corporation under a lease and bond agreement for $75,000, and it is reported that the Ruffner and McKinnon claims were purchased for $15,000, probably through a third party who sold to the United Verde Copper Co.

The Atlantic and Pacific Railroad (later the Atchison, Topeka and Santa Fe Railway Co.) crossed northern Arizona in 1882, and a wagon road was built from Ash Fork to Jerome, 60 miles away. Supplies were freighted to Jerome by mule and ox teams at a cost of $20 per ton. Two small water-jacket blast furnaces were transported to Jerome and production of matte and bullion started in August 1883. Surface oxide ores were mined. These deposits were rich in gold and silver as well as copper, and averaged about $30 per ton.

The following statement as given by H. C. Burchard, Director of the Mint on the production of precious metals in the United States for 1884, is quoted by Lindgren (1926, p. 62):

> The mines owned by the United Verde Copper Co., at Jerome, have proved a series of surprises to the owners. The properties were purchased and worked as copper properties, but as they have been developed they are found to contain silver in large quantities—in fact, so large that the silver is sufficient to pay all the running expenses of the mine, leaving the copper as a profit to the owners.
>
> Superintendent Thomas writes that another rich strike has been made in the Wade Hampton, one of the company's mines, on a drift from the 100-foot level 25 feet north of the last body of ore struck on the same level. The extent of it had not been ascertained, but the first samples taken assayed 20 percent copper and rich in silver. In every direction that drifts or crosscuts have been run ore bodies have been encountered. The furnace has run up to October 1, 1884, 289 days, and Superintendent Thomas gives the product by assays at the mines 4,396,951 pounds of refined copper and 237,951 ounces of silver. Estimating the average price of copper at $250 per ton, the gross yield of copper amounts to $548,500, and the silver at its coining value, $1.29 per ounce, amounts to $307,655.

In this same report (p. 32) it was stated that the United Verde Copper Co. paid a dividend of $60,000 in 1884. By this date, however, costs of mining increased, the rich surface ores were exhausted, and copper prices dropped from 14½ to 7½ cents per pound. Profits were replaced by losses and the mine closed in 1884. Governor Tritle remained optimistic about the property and in 1887 secured a lease and resumed mining, but his losses were heavy.

W. A. Clark, who later became Senator from Montana, visited the New Orleans Exposition in 1885 and saw on exhibit many specimens of ore from the United Verde mine. Clark recorded a note about the specimens but on returning to Butte forgot about the Jerome deposit. Mines owned by Clark at Butte supplied ore to the Port Orford Copper Co. in New Jersey, and when this refinery went into liquidation in 1886, Clark as one of the chief creditors, assumed control of the refinery and operated it for a year or more. Records of many shipments of matte, extremely rich in gold and silver, from the United Verde mine caught Clark's attention and he sent J. L. Giroux to examine the United Verde mine. Giroux found that the controlling interest of the ownership of the mine was under option to Dr. James Douglas of Phelps Dodge & Co. Dr. Douglas decided not to exercise his option because of the inaccessibility and "spotty" character of the ore, and Clark obtained an option through the assistance of James A. Macdonald who became vice-president of the company under Clark's reorganization. Clark and Giroux visited the United Verde property in March 1888, and after inspection, Clark immediately made the first payment required under the option, acquiring 70 percent of the stock. Clark started buying the remainder of the widely scattered stock of the company and at the time of his death in March 1925, he and his family owned about 299,000 of the 300,000 issued shares.

Under Clark's guidance, active mining operations started in 1888, using the original small smelting plant. The first dividend under this regime was paid in 1892. The Bullock railroad from Seligman to Prescott had then been completed, and freighting in and out of Jerome was done by way of Prescott. A branch of the Atchison, Topeka and Santa Fe Railway Co. was built from Ash Fork to Phoenix via Prescott in 1893, and Senator Clark immediately built a narrow-gage railroad, the United Verde and Pacific Railway, connecting Jerome with the Atchison, Topeka and Santa Fe Railway Co. at Jerome Junction, north of Prescott. In 1894, Clark built a new smelter at Jerome. In this same year, the first of the many serious mine fires occurred in the United Verde mine, as the massive pyrite caught fire readily.

Active prospecting started in the Jerome district in 1900 and the Verde Queen smelter treated carbonate copper ore from near the Columbia shaft, later United Verde Extension property. In 1904–05 the Equator Mining and Smelting Co. mined and smelted copper sulfide ore from the Iron King claim, a part of what is now known as the Copper Chief deposit.

The Verde Central group of claims was located as the Verde King in 1904 and four short tunnels were driven (Fearing and Benedict, 1925, p. 609) to explore faults and contacts, as no gossan was present. Some copper-stained rock was discovered, but no ore bodies, and when funds were exhausted in 1907, exploration ceased.

Mining operations in the Yaeger mine, on the west side of the Black Hills in the Black Hills mining district, began about 1904 and produced some copper and silver until 1909, when the mine closed.

In the Cherry district, south of the Jerome mining district, gold-bearing quartz veins were explored by small-scale mining, starting in 1883; mining activity reached a maximum in 1907 when 7 mines and 6 mills were in operation. Activity slowed down and by 1916, only 2 properties produced a little bullion and a shipment of a little copper ore.

During the early history of mining, Jerome was essentially a "one mine" camp and no extensions of the United Verde ore bodies were found outside the property of the United Verde Co. Visitors were not permitted underground at the mine so that little information was available to the "outsider" concerning its geology.

The United Verde ore bodies cropped out conspicuously in the country rock west of the Verde fault. East of the fault, however, the ore-bearing host rocks are covered by a thick blanket of Paleozoic and Tertiary rocks so that the deposits were hidden, and exploration was costly.

The United Verde Extension mine, the second largest producer in the Jerome district, is east of the Verde fault and the ore body is buried by younger rocks. The history of this mine started in 1900 when J. J. Fisher, United States deputy mineral surveyor, was surveying claims for G. W. Hull, a Jerome resident of many years. Fisher was well acquainted with the district, having been employed by many companies, including the United Verde Copper Co., and he was aware of the importance of tracing the United Verde ore zone into outside ground. While surveying Hull's claims, Fisher found a fraction of vacant ground covering less than an acre, between the March claim owned by Hull and the United Verde property. Fisher located this fraction as the Little Daisy (fig. 7).

Earlier in 1899, a company organized by G. W. Hull, was called the United Verde Extension Gold, Silver, and Copper Mining Co. It became known as the U. V. X., and its successors have been known by the same abbreviation. This first company controlled a dozen claims west, south, and north of the United Verde property. L. E. Whicher of the Schofield, Whicher, and Co. of Boston became interested in the promotion of the U. V. X. Whicher acquired 190,000 shares and Hull kept 110,000 shares of the 300,000 shares issued by the company. A shaft was sunk in claim "1888" adjacent to and southwest of the United Verde mine. M. F. Johnson, mining engineer, was employed by Whicher to examine the holdings of the company, and Johnson reported his belief that if any ore was to be found outside the holdings of the United Verde Copper Co., it would be to the east. By this time Fisher had located the Little Daisy fraction, and Whicher obtained an option on the Little Daisy for $50,000 and sank a shaft 300 feet deep that passed through 125 feet of oxidized copper-bearing rock, too low grade for ore, but sufficiently promising that Whicher began negotiations with Hull for four claims adjoining the Little Daisy: the March, Conglomerate, Iron Carbonate, and Bitter Creek (fig. 7). As a result of the transaction, Hull transferred his 110,000 shares of U. V. X. stock to the U. V. X. company and deeded the four claims to U. V. X. in exchange for other claims near the United Verde mine owned by the U. V. X. company. Shortly thereafter, Whicher purchased the Little Daisy fraction from Fisher for 5,000 shares of U. V. X. stock. The U. V. X. company was reorganized in 1902 under the laws of Maine, and 300,000 shares at a par value of $10 were issued. In 1910, another reorganization of the company took place under the laws of Delaware and 400,000 shares of stock of the same par value were issued.

The Daisy shaft was sunk to the 800 level and the search for ore continued under Fisher's direction. In depth, all lateral exploratory openings to the east had to pass through the Verde fault characterized by a wide gouge and brecciated zone. From 1907 to 1911, C. C. Burger, who was confident of a favorable outcome to the exploration program, was the consulting engineer for the U. V. X. In 1911, ore assaying 2.6 to 3.1 percent copper and $2 in precious metals was found in a winze, 65 feet below the 800 level. The copper was present as a sulfide, but showed signs of leaching. In addition, a patch of chalcocite ore 5 feet wide and 15 feet long averaging 18.7 percent copper was found in a crosscut on the 700 level. Along the 800 level, a much larger body of low-grade oxidized rock was exposed, and its downward persistence was indicated by a winze. By this time, about $500,000 had been spent in exploratory work, and as the experts called in for examination of the property reported adversely, the exploration program ended.

In 1908 Major A. J. Pickrell purchased stock in the U. V. X. company from Fisher, and for a time Pickrell served as director of the company. When exploration was discontinued in December 1911, Pickrell wrote to James S. Douglas, son of Dr. James Douglas of Phelps Dodge & Co., and urged Douglas to visit Jerome and examine the exploration workings of the U. V. X. Douglas had been a frequent visitor to Jerome and at one time had examined copper carbonate showings on claims that later became a part of the Jerome Verde Copper Co. Thus Douglas had some familiarity with the local geology.

Douglas visited Jerome in December 1911 and was shown the U. V. X. workings of the 800 level, but because of water in the mine, he was unable to examine the winze or lower workings, which at that time extended to the 1,200 level. Douglas was favorably impressed with the mineralized rock on the 800 and 700 levels, and took an option on the property. This option was offered to Phelps Dodge & Co., whose legal counsel "disapproved of the transaction on account of some fancied defect in the title, in which supposition he proved to be wrong" (Rickard, 1918, p. 13). Thereupon Major Pickrell urged the younger Douglas to take over the option himself, and in the summer of 1912, Douglas and George E. Tener of Pittsburgh took the option and reorganized the company into the United Verde Extension Copper Co. and invited their friends to join in providing $225,000 to explore the property.

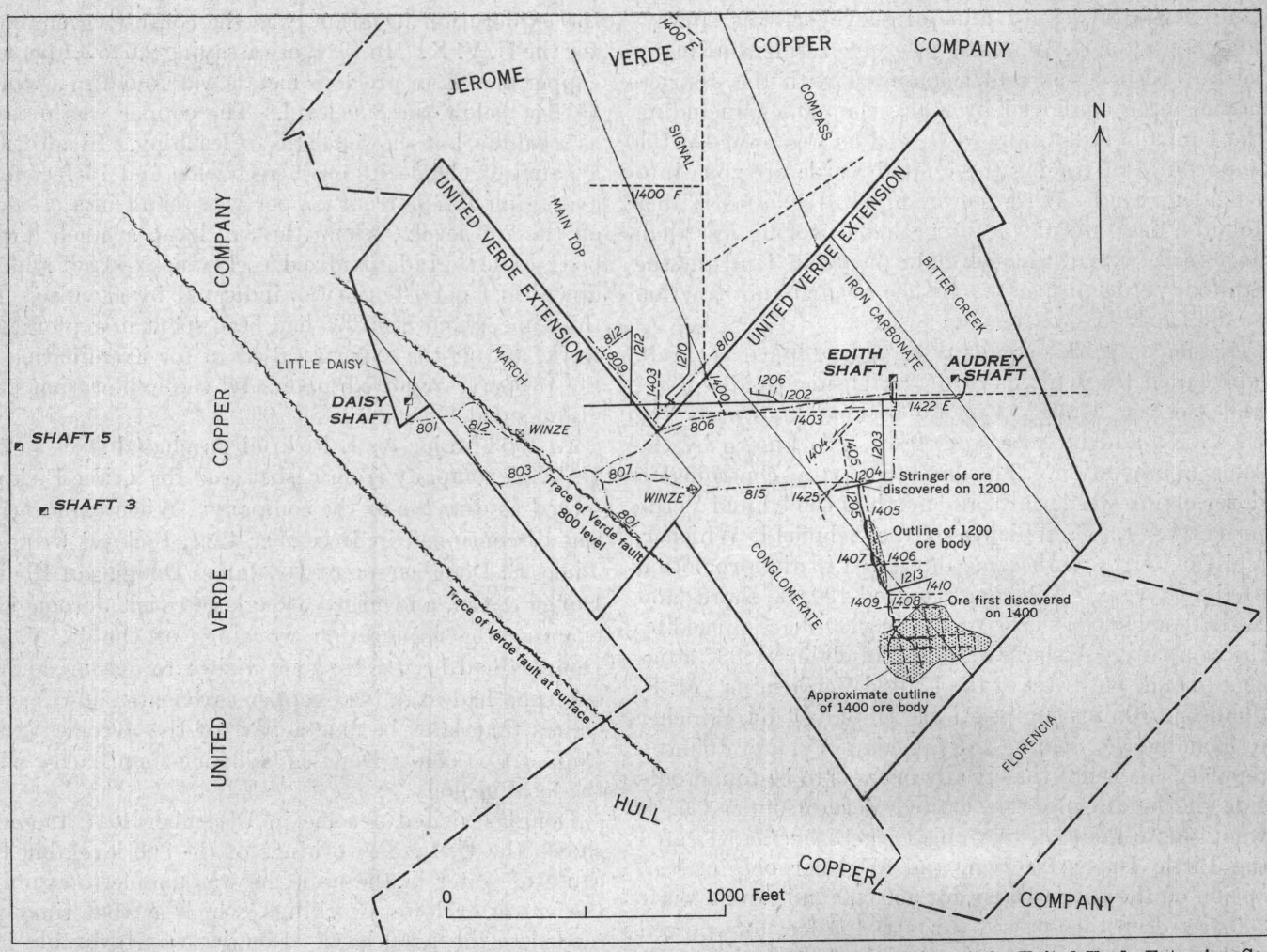

FIGURE 7.—Sketch map showing United Verde Extension property. Map of initial workings of the mine of the United Verde Extension Co. Numbers indicate the order in which the work was done. Modified from T. A. Rickard (1918, p. 15).

The necessary funds were raised by sale of stock; more than a third was bought by residents of Arizona.

Douglas decided to sink a new shaft farther from the Verde fault to avoid the trouble of driving workings through the wide crushed zone of the fault; and the Edith shaft, named after Miss Edith Tener, was started in June 1913—1,740 feet east of the Daisy shaft (fig. 7). The Edith shaft was sunk through 180 feet of lava, 400 feet of limestone, and 90 feet of sandstone, and reached the Precambrian rocks at 678 feet, which was equivalent to the 800 level of the Daisy shaft.

An agreement was made with the Jerome Verde Copper Co. to explore the Main Top, Signal, and Compass claims (fig. 7) to the north of the U. V. X. holdings, and the two properties were to be consolidated if ore was found on the Jerome Verde holdings. Under this agreement, exploratory work was done on the 800 and 1,200 levels without productive results. Likewise, search for ore on the U. V. X. ground was disappointing. The Edith shaft was deepened to 1,400 feet, and additional exploration was made in the Main Top claim of the Jerome Verde Copper Co. but no ore found; and in June 1915, the option on the Jerome Verde Copper Co. holdings was dropped.

In the fall of 1914, the U. V. X. company engaged the services of one of the foremost geologists in the country to examine the mine and show where the management had erred in their search for ore. "After a thorough examination, the distinguished geologist, whose name is purposely withheld, condemned the mine and strongly advised his clients to cease operations" (Rickard, 1918, p. 16). Lindgren (1926, p. 80) wrote, "I am not sure but that sympathy should be extended to this geologist. One may wonder how other 'experts' would have reported on the property." "However, as the Cornishman says, 'Never abandon a drift until you have driven 20 feet farther.' The management persevered a little longer" (Rickard, 1918, p. 16). Although the resources of the U. V. X. company were practically exhausted, in December 1914, a drift was driven toward the center of the U. V. X. holdings on the 1,200 level and a stringer of ore was found (fig. 7) which led to the discovery of a chalcocite body, 5 feet wide and containing 45 percent of copper. This was part of an ore body 120 feet long and reaching a little above the 1,100 level. About $600,000 worth of ore was mined from this body in 1915.

A drift was started on the 1,400 level to intersect the downward continuation of ore present on the 1,200 level, but it went beyond the point where ore was expected without finding ore. No crosscutting was done until January 1916, because of a heavy flow of water south of the ore body on the 1,200 level, which was not drained. Pumps were installed on the 1,400 level station, and the 1,400 exploratory drift continued under the "1,200 ore body," stopping in kaolinized rock 850 feet south of the Edith shaft. Crosscuts were driven both east and west in hopes of finding the high-grade ore discovered on the 1,200 level but these crosscuts failed as the ore body took a westward dip 60 feet below the 1,200 level, and the crosscuts were all too far to the east (fig. 7). The third crosscut to the east (labeled 1408, fig. 7) cut 100 feet of chalcocite ore and the main drift was extended into the ore body. Crosscuts driven 50 feet apart proved that an ore body reached within 30 feet of the Florencia side line (fig. 7). This was a new and larger ore body than the one discovered on the 1,200 level. On the 1,400 level, this mass of chalcocite ore had a maximum width of 260 feet and a length of 440, covering an area of 62,400 square feet. The upward termination of this main ore body was at 1,240 feet. "During 1916 the U. V. X. produced 36,402,972 pounds of copper from 77,461 tons of ore, an average of 23.5 percent copper, besides 2,570 ounces of gold and 128,468 ounces of silver, the total output being worth $9,949,918, of which $7,400,000 was profit. It had proved a glorious bonanza!" (Rickard, 1918, p. 17). In April 1917, the U. V. X. mine produced 6,991,480 pounds of copper, worth at the high prices then prevailing, $2,167,358.

Peak annual production of the United Verde Extension mine in terms of pounds of copper, was reached in 1917 when 63,879, 506 pounds of copper was obtained from 115,064 tons of ore. The ore averaged about 27.5 percent copper. A smelter was completed in 1918 at Clemenceau to handle the ore from the United Verde Extension mine. Peak annual tonnage production from the mine was reached in 1929 when 358,654 tons was mined, yielding 59,125,000 pounds copper, about 8.6 percent copper as the average grade mined for the year. In 1930, a 200-ton flotation mill was added to the smelter at Clemenceau to handle lower grade copper ore but in January 1937, as reserves in the mine were depleted, the smelter closed and in May 1938, the United Verde Extension mine closed; the deposit was mined out.

Naturally the discovery of the buried rich U. V. X. ore body led to many prospecting ventures in the district, but most of these were unsuccessful. An outlying ore body of the U. V. X. zone was found in 1917 in the Main Top claim of the Jerome Verde Copper Co., but this was mined out in 1920 after producing 1,500,000 pounds of copper (Elsing and Heineman, 1936, p. 101). During this period, chrysocolla-bearing gravel was explored by the Dundee-Arizona Copper Co. on two claims east of Jerome. The Copper Chief

mine operated from 1916 to 1923, when a 125-ton cyanide mill was in operation and oxidized ore from the upper levels was mined. Copper-silver ore was mined from the Shea mine, south of the Copper Chief mine, during this same period. The Yaeger mine in the Black Hills district was active in 1922, and 800 tons of copper-silver ore was shipped to the Humboldt smelter.

The Arizona National mine in the Big Bug mining district is located on the Silver Belt-McCabe vein, and high-grade silver-lead ore and concentrates were shipped intermittently between 1915 and 1931. Total production during this period was valued at $250,000 to $300,000, judging from the available records.

At the United Verde mine, mine fires and subsequent ground movement required the construction of a new smelter as the old smelter was adjacent to the mine. The new smelter was built at Clarkdale and placed in operation in 1915. A standard gauge railroad from Drake on the Prescott-Ash Fork branch of the Atchison Topeka and Santa Fe Railway Co. along the Verde River to Clarkdale was completed in 1914. Stoping within the fire area of the upper part of the United Verde mine was discontinued in 1916, and open-pit stripping operations were started in 1917 to uncover the ore above the 600 level. Open-pit mining started in 1922. A standard-gage railroad was completed from Clarkdale to Jerome in 1920, and in 1921, a direct highway was constructed from Jerome to Prescott. A flotation concentrator was added to the smelter at Clarkdale in 1927 to handle copper ore associated with silicate gangue minerals. The United Verde mine had its peak annual copper production in 1929, obtaining 142,290,000 pounds of copper from 1,737,000 tons of ore, an average grade of 4.1 percent copper.

The Verde Central mine, south of the United Verde mine at Jerome, attracted the attention of C. T. Joslin of Prescott, in April 1916, and a new exploration program was started, and in March 1918, W. F. Staunton became president and general manager of the Verde Central Mines, Inc. In December 1918, T. H. Collins purchased $50,000 worth of stock from Joslin in order to keep the exploration program active. Calumet and Arizona took an option on a large block of stock in 1921 and provided the funds for continued exploration. In 1924, Calumet and Arizona took complete control of the property, and G. A. Campbell was made president and Staunton remained in charge as vice-president and general manager. A 350-ton flotation mill was placed in operation in 1929; and in 1929 and 1930, 8,600,000 pounds of copper was mined and milled, having a value with the silver content of $1,338,000 (Elsing and Heineman, 1936, p. 101). The mine closed in 1930 and the property was purchased by the United Verde Co. in 1931. No mining has been done since that date. During 1949, R. L. d'Arcy shipped some of the low-grade siliceous copper ore from the dump, and mined some oxide ore from the top of the hill to the south of the mine where the Verde Central ore zone crops out.

The underground operations at the United Verde mine stopped in 1931 owing to the low price of copper and were not resumed until April 1937, but in the open pit, stripping operations continued and mining of ore was resumed in 1934 and continued until 1940 and bottomed on the 630 level. The mine was purchased from the Clark family in 1935 by Phelps Dodge Corp., the present owner (United Verde Branch, Phelps Dodge Corp.). Because of depleted reserves, the smelter at Clarkdale closed in June 1950 and the mine continued operations on copper-zinc ore that was concentrated at Clarkdale whence the zinc concentrates were sold to custom smelters and the copper concentrates were shipped to the Phelps Dodge smelter at Douglas, Ariz. The mine closed in February 1953.

With the depletion of ore reserves at the United Verde mine, the Iron King lead-zinc mine in the Big Bug district (shown on the southwest corner of pl. 1) is the leading productive mine in the area described in this report. Some of the older residents of Prescott believe that the original location was made about 1880 in search for gold and silver, but that a few shallow workings proved disappointing, and there are no records of production. In 1906–7 the Iron King mine was operated by Reverend Ben Blanchard, who mined oxide ore. In 1907 production amounted to 1,253 ounces of gold, 35,491 ounces of silver, and 3,933 pounds of copper (Lindgren, 1926). Some secondary copper ore was reported from near water level. The mine was inactive until World War I when it was reactivated by George Colvocoresses in order to obtain sulfide ore for his smelter at Humboldt. By 1922, it is reported that several thousand tons of ore averaging $8 per ton in gold and silver had been mined. In 1933, Fred Gibbs sampled the mine and purchased the property in 1934 under tax sale for a few hundred dollars. In 1936 Gibbs and associates formed the Iron King Mining Co. to operate the mine, and in 1937 they shipped oxide gold-silver ore. A bulk flotation mill with a daily capacity of 140 tons was built in 1938 to separate the lead-zinc sulfide minerals. In 1939 the mill was converted to differential flotation to produce separate lead and zinc sulfide concentrates, and the mill capacity was increased to 225 tons daily. Shattuck Denn Mining Corp. purchased the property in 1942, and is the present operator. Mill capacity has been increased gradually, and in 1952, was 600 tons daily; gold, silver, lead, zinc, a little copper, and fluxing pyrite are recovered.

During World War II under the incentive of premium prices for copper, oxide copper ore was mined from the Tertiary gravel beds at the Dundee-Arizona mine, a short distance east of Jerome, and from Tertiary gravel and basalt flows on United Verde Extension property, south of the Edith shaft. The demand for siliceous flux for the Clarkdale smelter resulted in some small-scale mining of the oxidized ore at the Copper Chief mine and upper levels of the United Verde Extension mine, and the richer parts of the Verde Central dump were shipped to the smelter.

Early records are incomplete but the available data indicate that to 1952, the following approximate amounts of metal have been recovered from the Jerome area: gold, 1,890,000 ounces; silver, 65,600,000 ounces; copper, 1,868,000 tons; lead, 37,725 tons; and zinc, 127,000 tons. The Verde mining district dominated by the United Verde and United Verde Extension mines accounts for 82 percent of the gold, 87 percent of the silver, 18 percent of the zinc, and more than 99 percent of the copper. The Iron King mine accounts for 11 percent of the gold, 10 percent of the silver, 82 percent of the zinc, and 96 percent of the lead. Records are incomplete concerning the value of this production, but $700,000,000 is a fair estimate.

MINERALOGY

Minerals related directly to the ore deposits are described briefly in this section. The common rock-forming minerals of no economic importance are not mentioned. The ore and associated minerals are grouped according to their origin; those formed by ascending ore-forming solutions (hypogene minerals) and those formed by the action of descending waters (supergene minerals). The supergene group includes sulfide, oxide, carbonate, and sulfate minerals. Some of the so-called ore minerals, such as arsenopyrite, have no economic importance, but because of their genetic relationship to the ore minerals are described under that heading.

HYPOGENE MINERALS

ORE MINERALS

Pyrite (FeS_2).—This is one of the most abundant hypogene minerals in the area. It is a major constituent in the massive sulfide bodies of the United Verde, United Verde Extension, and Copper Chief mines at Jerome, and the Iron King mine at Humboldt. Pyrite is common in all the fissure veins. Scattered pyrite crystals are widespread in all parts of the area where there has been any appreciable hydrothermal alteration.

Pyrite is recovered in the concentrator at the Iron King mine, largely because of the gold content, but in part because of its use as a smelter flux. The large body of massive pyritic sulfide at the United Verde mine represents an important natural resource of sulfur that undoubtedly will be mined in the future when economic conditions are favorable.

Chalcopyrite ($CuFeS_2$).—This primary sulfide of copper and iron is the chief ore mineral at the United Verde and Verde Central mines and on the lower levels of the United Verde Extension mine at Jerome. Small quantities are present in many fissure vein deposits throughout the area. Chalcopyrite is present in the "copper vein" a short distance west of the Iron King mine.

Bornite ($CuFeS_4$).—Beautiful hand specimens of bornite ore are found locally in the United Verde mine, and much of the ore in the Yaeger mine was massive bornite. Small blebs of bornite are found in fissure veins near the Yaeger mine and in the south-central part of the Mingus Mountain quadrangle.

Galena (PbS).—Very fine to microscopic grains of galena are important economic constituents of the Iron King massive sulfide ore. Small grains of galena are present also in the United Verde massive sulfide, but not in recoverable amount. Scattered medium to large crystals of galena are common in most of the fissure veins.

Sphalerite (ZnS).—Sphalerite is the most abundant ore mineral in the Iron King mine, where it occurs as minute grains forming streaks and aggregates in the pyritic massive sulfide, and as grains interstitial to the pyrite. In the United Verde mine, sphalerite, in small quantities, is present throughout much of the massive sulfide body. Locally, there is some concentration of sphalerite, particularly where the copper content is low, but not in sufficient tonnage to be mined under past economic conditions. During recent years, sphalerite has been selectively floated in the concentrator from zinc-copper ore from the north ore body on lower levels of the United Verde mine. Scattered crystals of sphalerite are common in many of the fissure veins.

Tetrahedrite ($(Cu, Fe)_{12}Sb_4S_{13}$).—Argentiferous tetrahedrite was the chief ore mineral at the Shea mine. The chemical analysis (table 15) shows the low arsenic content in this mineral, proving that it is tetrahedrite.

TABLE 15.—*Chemical analysis of tetrahedrite, Shea mine*

[Analysis furnished by the courtesy of Phelps Dodge Corp.]

Cu	percent	31.7
Fe	do	3.3
Sb	do	23.4
As	do	2.1
S	do	22.8
Ag	ounces per ton	583.57
Au	do	.21

According to Lindgren (1926, p. 100) mercurial tetrahedrite was found at the Shylock mine, and upon de-

composition, secondary cinnabar was deposited as coatings and filled cracks.

Tennantite ($(Cu, Fe)_{12}As_4S_{13}$).—Tennantite, the arsenical variety of gray copper, is present in both the United Verde and Iron King mines, where it adds to the silver content of the ore. The chemical analyses (table 16) are of tennantite from the United Verde mine.

TABLE 16.—*Chemical analyses of tennantite, United Verde mine*

[Analyses furnished by the courtesy of Phelps Dodge Corp.]

	1	2
Cu, percent	41.6	36.95
Fe, do	.3	
Zn, do	10.9	
Sb, do	Tr.	2.2
As, do	17.4	16.0
S, do	27.3	24.4
Ag, ounces per ton	38.08	31.76
Au, do	.075	.14

1. 2,700 level; 2. 1,350 level.

Tennantite high in zinc content is common although the 10.9 percent zinc here reported is higher than any reported by Palache and others (1944, p. 376). The high-zinc sample (analysis 1) occurred as massive homogeneous tennantite in a quartz-carbonate vein in the United Verde mine, in which there is no visible contamination by sphalerite.

Tennantite has been found in the Yaeger mine.

Pyrrhotite ($Fe_{1-x}S$).—Pyrrhotite is rare in the Jerome area, and has been recognized only in the Haynes massive sulfide body, west of the United Verde ore body. The pyrrhotite is associated with pyrite and chalcopyrite.

Arsenopyrite (FeAsS).—Small crystals of arsenopyrite are scattered throughout the massive sulfide ore bodies at the United Verde and Iron King mines. A cobaltiferous variety is probably present in a prospect southeast of the Shylock mine (1,315,000 N.; 422,000 E.), as cobalt bloom is present at the surface.

Molybdenite (MoS_2).—Molybdenite has been observed in only one locality in the Jerome area; a small prospect on Burnt Canyon (1,327,800 N.; 438,700 E.) reveals scattered molybdenite crystals in a quartz vein.

Gold (Au).—Gold is economically important in all the mines of the area, and is intimately associated with the sulfide minerals. At the Iron King mine, metallurgical tests and assays show that the gold is free, and occurs in galena, sphalerite, and pyrite. Pyrite carries most of the gold, chiefly because pyrite is more abundant. Visible gold has been reported from mines in the Cherry Creek district. Much of the gossan overlying the important sulfide deposits contains gold; it was this gold that attracted early attention to mining in the Jerome area.

Magnetite (Fe_3O_4).—Magnetite is rare in the United Verde massive sulfide body, but is more conspicuous in the Haynes ore body to the west where some concentration of magnetite was seen in chlorite associated with the sulfide minerals.

Magnetite forms in distinct crystals and masses in altered facies of the Spud Mountain rhyolite, 2½ miles east of the Iron King mine. Although of no present economic importance, magnetite is a major constituent in sedimentary jasper-magnetite interbeds in the Grapevine Gulch formation.

Hematite (Fe_2O_3).—The specular variety of hematite is present in late veinlets cutting the massive sulfide at the United Verde mine. Specular hematite is common as veinlets in all the Precambrian rocks south of Jerome. Hematite is an important constituent in many of the jasper-magnetite interbeds in the Grapevine Gulch formation.

GANGUE MINERALS

Quartz (SiO_2).—Quartz is one of the most important gangue minerals in the ore deposits in the Jerome area. Pure masses of quartz are closely associated with the massive sulfide deposits at Jerome and at the Iron King mine. Quartz is also common in the massive sulfide, interstitial to the sulfide grains. Many bodies of quartz and jasper south of Jerome are associated with minor sulfide minerals. Quartz is the dominant gangue mineral in the fissure veins.

Muscovite ($H_2KAl_3(SiO_4)_3$).—The fine-grained variety of this mineral, sericite, is common in the alteration zone associated with the Iron King massive sulfide ore bodies. Some of the rocks in the United Verde mine contain sericite, particularly the quartz porphyry and fine tuffaceous sedimentary rocks of the Grapevine Gulch formation, but it is debatable whether this sericite is of metamorphic or hydrothermal origin. We believe that much of it is hydrothermal.

Sericite is present in many of the alteration zones associated with the fissure veins.

Carbonate minerals.—Carbonate minerals recognized in the ore deposits of the Jerome area includes: calcite ($CaCO_3$); dolomite ($(Ca,Mg)CO_3$); ankerite $CaCO_3 \cdot (Mg,Fe,Mn)CO_3$. The distinction of these varieties is not always simple, for commonly they occur in microscopic aggregates, and are recognizable only in thin sections under the microscope. Enough indices of refraction have been measured by microscopic methods to demonstrate the occurrence of calcite, dolomite, and ankerite, in many places closely associated.

Dolomite and ankerite are common in the massive sulfide bodies, interstitial to the sulfide granules. In

the United Verde mine, dolomite and calcite are common in the late quartz-carbonate veins. Most of the fissure veins in the area contain variable amounts of these carbonate minerals.

Siderite ($FeCO_3$) has been reported by Lindgren (1926, p. 25) in the Shea and Yaeger mines, but no siderite was positively identified in our study.

Barite ($BaSO_4$.—Barite is reported (Reber, 1938, p. 59) in the Shea mine, not associated with the sulfide minerals, which may be older. Lindgren (1926, p. 25) reported barite in the Silver Belt vein, to the west of the Iron King mine.

Tourmaline (complex silicate of B, Al, Fe, Mg).—Minute needles of black tourmaline have been seen in the quartz of some fissure veins in the Cherry Creek district, and quartz-tourmaline rock occurs west of the Iron King mine.

Chlorite (complex silicate of Fe, Mg, Al_2O_3).—Chlorite is an important alteration mineral at the United Verde mine, for much of the copper is spatially related to the introduced chlorite that replaced different rocks to form nearly pure chlorite aggregates, referred to as black schist in the United Verde mine. The term "black schist" is used in this report for all chlorite-rich aggregates that have replaced any rocks in the area. Chlorite has been introduced in many of the Precambrian rocks south of Jerome, although few areas of black schist are large.

Microscopic studies revealed considerable variation in optical properties, and some chemical analyses of the chlorite were made by Phelps Dodge Corp. to check the optical data. The intermediate index of refraction, optical sign, and birefringence are the optical properties most easily determined. Much of the chlorite in the alteration zones is extremely fine grained, and the color of the aggregate is black. The rare large flakes of chlorite are pale green. In all samples tested the optical properties of the pale green coarse chlorite and of adjacent microscopic chlorite are identical.

Much of the chlorite measured has birefringence between 0.004 and 0.010, positive sign, negative elongation, and with beta between 1.585 and 1.622. The chlorite of lower index has 26.12 percent MgO and 5.48 percent FeO. Another sample of chlorite in this group, with beta index 1.610, has 18.98 percent MgO, and 20.04 percent FeO. The higher indices of refraction reflect a higher FeO:MgO ratio. Using Winchell's chart (1936, p. 649), the chlorite of lower index would be called clinochlore; the chlorite of higher index would be prochlorite. Much of the chlorite associated with the United Verde ore body belongs to the clinochlore-prochlorite group.

Chlorite with birefringence less than 0.004, and optically positive, is common in the andesitic dikes that cut the United Verde ore body and in some of the partly chloritized tuffaceous sediments of the Grapevine Gulch formation near the United Verde ore body. The intermediate index of refraction ranges from 1.622 to 1.635 in this group. The chlorite of lower index (beta, 1.622) has about 10 percent MgO and about 15 percent FeO. The higher index chlorite (beta, 1.635) has 9.9 percent MgO and 25.9 percent FeO. The chlorite of lower index would be called rumpfite and that of higher index ripidolite, following Winchell (1936, p. 648). One sample of ripidolite forms black schist associated with ore at the United Verde mine.

All chlorite studied from the chloritized Deception rhyolite near Mescal and Deception gulches, south of Jerome, has a birefringence less than 0.004, and is optically negative, with beta index between 1.618 and 1.635. Some chlorite in the black schist at the United Verde mine also belongs to this low birefringent, optically negative group, as does some chlorite in the chloritized Grapevine Gulch and gabbro in the United Verde mine. A sample from Deception Gulch, having an intermediate index of 1.624, contains MgO, 12.50 percent, and FeO, 22.83 percent. In chloritized gabbro from the Haynes ore body, on the west side of the gabbro, United Verde mine, the chlorite has an intermediate index of refraction of 1.643, and the MgO content is about 7 percent and the FeO content is about 40 percent. According to Winchell (1936, p. 648), the chlorite having an intermediate index below 1.630 would be called diabantite, and above 1.630 the chlorite would be called aphrosiderite. Following this classification, most of the low birefringent, optically negative chlorite in the Jerome area is diabantite.

Very little chlorite was found to be optically negative and with a birefringence between 0.004 and 0.010. The few specimens from this group have beta index above 1.640. Only one specimen of black schist associated with ore in the United Verde mine belongs to this group; its beta index is 1.642. Southwest of the Copper Chief mine along the Pipe Line road (1,344,000 N.; 445,000 E.), a 3-inch band of chloritized Shea basalt is composed of chlorite with beta index 1.646, 5.8 percent MgO, and 31.4 percent FeO. According to Winchell (1936, p. 648) this chlorite would be termed brunsvigite. East of the Shea mine in the Buzzard rhyolite (1,344,000 N.; 450,000 E.), small areas of black schist contain chlorite with beta index 1.660, 4.8 percent MgO, and 34.2 percent FeO. Winchell (1936, p. 648) would classify this chlorite as thuringite.

Plotting the molecular ratios of MgO; FeO, Fe_2O_3; and Al_2O_3 of 6 chlorite minerals for which data are suitable, in a triangular diagram, shows that the chlorite minerals in the Jerome area have a considerable range in the MgO-FeO, Fe_2O_3 ratio, but the Al_2O_3 ratio to the

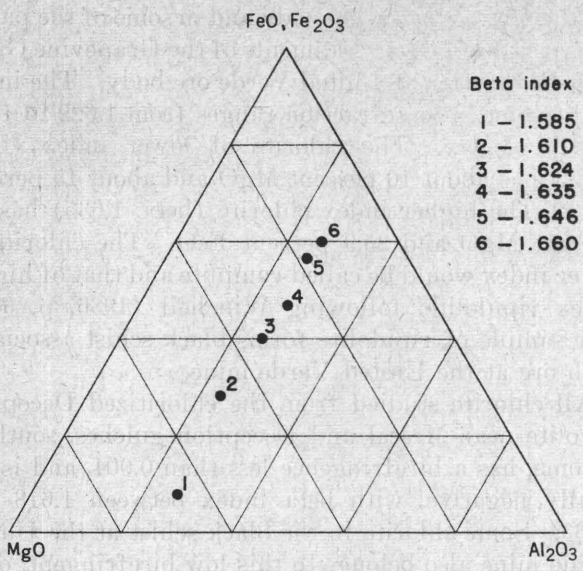

FIGURE 8.—Triangular diagram showing molecular ratio of FeO, Fe_2O_3; MgO; and Al_2O_3 in chlorite minerals.

MgO; FeO, Fe_2O_3 is nearly constant (fig. 8). The intermediate index of refraction is given, showing how the increased iron content is reflected in the higher index of refraction.

The data indicate that most chlorite forming black schist associated with ore in the United Verde mine belongs to the clinochlore-prochlorite varieties, though most other varieties of chlorite recognized here may locally form black schist also. Rumpfite-ripidolite is more common in the andesitic dikes that cut the United Verde ore body, and in the weakly chloritized Grapevine Gulch formation where the chlorite content is too low to form "black schist." Diabantite appears to be the common chlorite in the local chloritized parts of the Deception rhyolite south of Jerome. Aphrosiderite is less common among the chlorite minerals and is limited largely to chloritized gabbro in the United Verde mine and to Shea basalt, well to the south of Jerome. Brunsvigite and thuringite, the most iron-rich of the chlorite minerals, have been found only 4 to 5 miles south of Jerome except for one specimen of black schist in the United Verde mine.

Gavelin (1939, p. 70) determined the optical properties of the chlorite minerals associated with the massive sulfide deposits in the Malenas district, Sweden, and found most of the varieties to be optically positive, and to belong to the clinochlore-prochlorite-rumpfite groups of Winchell (1936).

SUPERGENE MINERALS

ORE MINERALS

Chalcocite (Cu_2S).—The rich copper ore at the United Verde Extension mine was composed of massive hard gray and soft, sooty chalcocite, formed by supergene enrichment in Precambrian time. Chalcocite was also mined from the upper levels of the United Verde mine, a product of supergene enrichment largely related to the present erosion surface. Chalcocite is present in the eastern part of the Copper Chief mine, where the grade was high enough in copper to warrant mining. Some secondary chalcocite was present in the upper part of the Yaeger mine. Small quantities of secondary chalcocite are found in all the fissure veins that contain primary chalcopyrite.

Silver (Ag).—Native silver was found in the upper levels of the United Verde mine in gossan directly overlying sulfide minerals. Enrichment in silver was noted at the United Verde Extension mine in a similar environment but Lindgren (1926, p. 86) does not state whether or not native silver was observed.

Copper (Cu).—Native copper is common in the United Verde Extension mine where it occurs with chalcocite and cuprite, and in the soft gossan overlying the chalcocite ore. Native copper was also mined from the oxidized upper part of the United Verde ore body.

Cuprite (Cu_2O).—This red oxide of copper was a common associate with native copper in the United Verde and United Verde Extension mines.

Copper carbonate.—Green malachite ($CuCO_3 \cdot Cu(OH)_2$) and blue azurite ($2CuCO_3 \cdot Cu(OH)_2$) are common in all the oxidized parts of all the copper deposits in the area. Scattered films of these two minerals are common at the surface of many of the fissure veins.

Chrysocolla ($CuSiO_3 \cdot 2H_2O$).—Chrysocolla was the chief copper mineral mined from the gravel beds in the Dundee–Arizona mine at Jerome. Surface films of chrysocolla are common throughout the Jerome area where copper is found.

Anglesite ($PbSO_4$) *and cerussite* ($PbCO_3$).—Anglesite and cerussite are common oxidation products of galena and are present in the oxidized parts of the Iron King ore body. They have also been reported from the Copper Chief mine.

Limonite (hydrated iron oxide).—This term is a desirable field term for the ferric oxide that contains varying amounts of water and occurs in yellow and brown colors. Much of the powdery aggregate includes jarosite, a hydrous sulfate of iron and potash. Limonite forms much of the gossan over all the sulfide deposits in the area. Limonite (gossan) has been mined at many of the deposits in the area because of the gold-silver enrichment. Much of the early mining at some deposits was due entirely to the precious-metal content.

MINERALS RELATED TO MINE FIRES

During the open-pit mining of the United Verde deposit, stripping operations uncovered hot rocks impreg-

nated with various hydrous sulfate minerals formed as a result of the burning sulfide. Several methods were used to extinguish the fires, such as flooding with water, adding carbon dioxide, and later steam, but none was successful. Finally air was forced into the fire area under a pressure of 2 to 5 pounds and this method forced the hot gases back and cooled the rocks (Tally, 1917).

Lausen (1928) believed that the hydrous sulfate minerals were probably formed when water was first used to put out the fires, but as crystals of one composition rest upon others, conditions must have changed from time to time. The gases must have contained a large percentage of water vapor which reacted with oxidized sulfides of iron and copper and carried these constituents to higher levels. Presumably the constituents were transported upward in a gaseous phase.

In 1924, E. D. Gardner and G. W. Jones of the U. S. Bureau of Mines collected 5 gas samples from cracks and drill holes in the pit. These samples were analyzed by G. W. Jones of the Bureau of Mines and the results are given by Lausen (1928, p. 206). Two samples contained chiefly atmospheric gases, and in only two samples did the SO_2 content equal 6 percent. One sample contained 4.67 percent CO_2 which may have been derived from burning mine timber.

The hydrous sulfate minerals were deposited chiefly in cracks in the quartz on the west and north sides of the massive sulfide, although some were found along cracks in the massive pyritic sulfide.

Altogether, nine hydrous sulfate minerals were found, five of which are new (Lausen, 1928). The new minerals are as follows: butlerite, $(Fe,Al)_2O_3 \cdot 2SO_3 \cdot 5H_2O$; guildite, $3(Cu,Fe)O \cdot 2(Fe,Al)_2O_3 \cdot 7SO_3 \cdot 17H_2O$; louderbackite, $2FeO \cdot 3(Fe,Al)_2O_3 \cdot 10SO_3 \cdot 35H_2O$; ransomite, $CuO(Fe,Al)_2O_3 \cdot 4SO_4 \cdot 7H_2O$; rogersite, hydrous sulfate of ferric iron. The other hydrous sulfate minerals that formed are alunogen, $Al_2O_3 \cdot 3SO_3 \cdot 18H_2O$; copiapite, $2Fe_2O_3 \cdot 5SO_3 \cdot 18H_2O$; coquimbite, $2(Fe,Al)_2O_3 \cdot 7SO_3 \cdot 22H_2O$; voltaite, hydrous potash iron sulfate (Anderson, 1927). In addition, some sulfur was deposited and a black globular coating on fragments of rock beneath the iron hoods placed over the vents from which the heated gases were escaping. Lausen (1928, p. 227) named this material jeromite, a sulfide of arsenic containing some selenium.

GENERAL CHARACTERS

The ore deposits of Precambrian age in the Jerome area may be classified into two types: (1) massive sulfide deposits and (2) fissure veins. Most of the metal production in the area has come from the massive sulfide deposits, represented by the United Verde and United Verde Extension copper mines at Jerome and by the Iron King lead-zinc mine at Humboldt. The fissure veins have been mined chiefly for their precious-metal content, such as the gold-quartz veins in the Cherry Creek district and silver-bearing vein of the Shea mine south of Jerome. Copper has been the chief metal from the Verde Central and Yaeger veins, although much gold and silver were recovered from the Yaeger ore. The McCabe-Gladstone vein, west of the Iron King mine, was an important producer of gold and silver, but it cannot be proven that this deposit is of Precambrian age.

At Jerome, some copper has been mined from Tertiary lava flows and gravel beds impregnated with supergene chrysocolla.

The distribution of the principal mines in the Bradshaw Mountains and Black Hills is shown in figure 9;

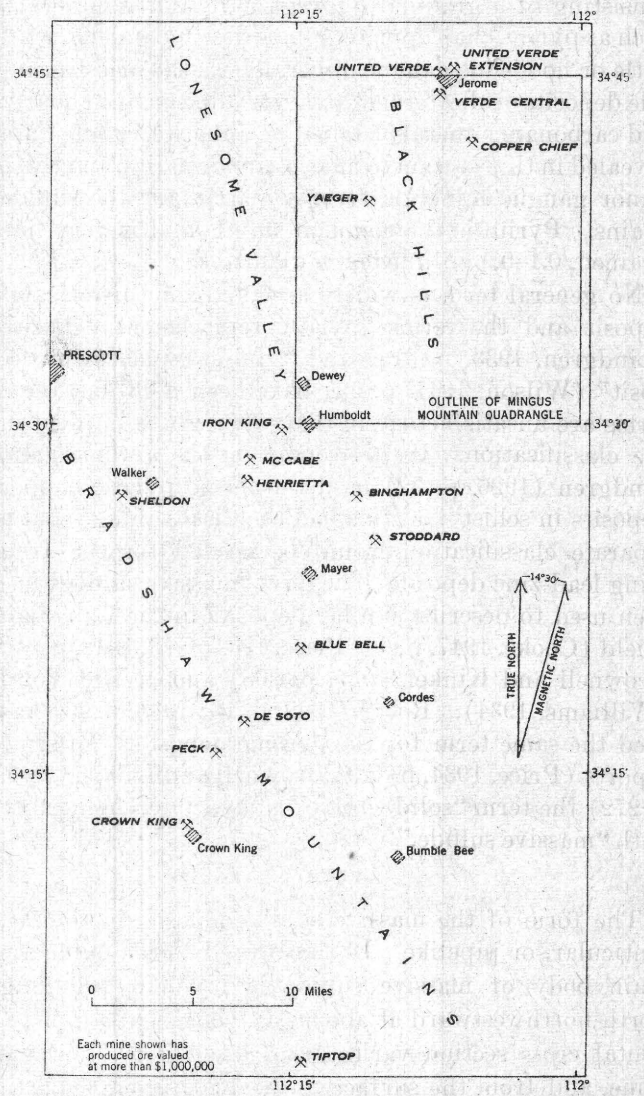

FIGURE 9.—Sketch map of Bradshaw Mountains and Black Hills showing location of principal mines.

only the mines north of the McCabe mine, southwest of Humboldt, are described in this report. Figure 9 shows that except for the Sheldon, southwest of Walker, the mines of significant production lie within two northeastward-trending zones, one extending from the Henrietta to the United Verde and United Verde Extension mines at Jerome, and the other extending from the Crown King to the Binghampton. Part of these deposits are massive sulfide and part are fissure veins. This northeast trend in part is reflected by the northeastward-trending structures in the Alder group, but not by any known structure in the Black Hills. Possibly this distribution of the principal mines is fortuitous; no satisfactory explanation can be advanced with our present information.

MASSIVE SULFIDE DEPOSITS

The term "massive sulfide" is used for the deposits consisting of a granular aggregate of sulfide minerals, such as pyrite, chalcopyrite, sphalerite, and galena, with little or no visible gangue minerals. In the massive sulfide deposits of the Jerome area, very fine grained quartz and carbonate minerals (dolomite, ankerite, calcite) are revealed in thin section by microscopic studies, and these minor gangue minerals form a matrix for the sulfide grains. Pyrite is the common sulfide and is very fine grained, 0.1–0.2 millimeter in diameter.

No general term is widely accepted for this class of deposit and the terms "pyritic replacement deposit" (Lindgren, 1933, p. 618) and "sulfide replacement deposit" (Wilson, 1941, p. 59) have been used, but these terms are unsatisfactory in that an origin is implied in the classification. In his report on the Jerome area, Lindgren (1926, p. 32) used the term "pyritic copper deposits in schist" but this had the disadvantage that a separate classification would be needed for the Iron King lead-zinc deposit. The term "massive sulfide" has been used to describe similar deposits in the Canadian shield (Cooke, 1947, p. 66; Price, 1934; Gill, 1948, p. 36; Brownell and Kinkel, 1935, p. 268) and at Rio Tinto (Williams, 1934). Reber (1922, p. 16; 1938, p. 43) has used the same term for the Jerome deposits. In some reports (Price, 1934, p. 129; Brownell and Kinkel, 1935, p. 272) the term "solid sulfide" is used interchangeably with "massive sulfide."

FORM

The form of the massive sulfide deposits is tabular, lenticular, or pipelike. In the United Verde mine, the main body of massive sulfide is pipelike, plunging north-northwestward at about 65°. The shape in horizontal cross section varies from level to level in the mine, and from the surface to the 1,200 level, the area of the massive sulfide deposit averages about 250,000 square feet, but on the 1,500 level, it is almost 500,000 square feet. On lower levels, the horizontal area of massive sulfide decreases so that on the 3,300 level it is only 50,000 square feet. The North ore body in the United Verde mine extends from above the 3,750 to below the 4,500 levels and is a northward-trending, steeply dipping, elongate lens, arcuate at the south end. The average width of this North ore body ranges from 10 to 30 feet, swelling locally to 60 feet.

The massive sulfide body in the United Verde Extension mine is a lens trending eastward; on the 1,300 level, the horizontal cross section area is about 100,000 square feet. The body tapers in depth and is in fault contact with the Verde fault.

The Copper Chief deposit is also lenticular, elongated in an easterly direction, but it is only about 300 feet deep. The ore body is about 800 feet long and 60 feet wide.

The Iron King massive sulfide deposit forms a set of en echelon veins, striking N. 22° E. and dipping 71° W. The north ends of the veins plunge northward at nearly 60°. Each vein on the west extends farther to the north than adjacent veins to the east. The width of the vein ranges from 1 to 14 feet, and the length measures in hundreds of feet.

The regional structure apparently has considerable control on the form of the massive sulfide bodies, as pointed out by Gavelin (1939). In areas characterized by pronounced foliation (S-tectonites) the massive sulfide minerals form veins or disseminated deposits parallel to foliation, as exemplified by the Iron King deposit. However, in a metamorphic terrane dominated by folds (B-tectonites) a close relationship may exist between fold axes, linear structures and form of the deposit, which results in lenticular and pipelike bodies. The United Verde pipelike deposits are apparently localized in the nose of a north-northwestward-plunging anticline, and the plunge of the pipe is parallel to the plunge of the minor folds and lineation marked by cleavage-bedding intersections.

MINERALOGY AND STRUCTURE

The massive sulfide deposits are dominantly pyritic and contain variable amounts of chalcopyrite, tennantite, sphalerite, and galena, as well as minor amounts of arsenopyrite and locally, pyrrhotite and bornite. The pyrite-rich facies is generally nonbanded, but where appreciable amounts of sphalerite, chalcopyrite, or galena are associated with the pyrite, banded facies are present.

The margins of the massive sulfide bodies are generally in sharp contact to the adjacent rocks, but in

places the massive sulfide bodies grade into disseminated deposits where the sulfide minerals are in scattered crystals or aggregates or in veins in the host rocks. In the United Verde mine, some of the country rock containing veins and disseminated grains of chalcopyrite constitutes ore, but at the Iron King mine, the disseminated sulfide is pyrite that is not ore.

Veins of quartz or quartz-carbonate cut the massive sulfide bodies and related disseminated mineralized rocks, and are late in the sequence of mineralization. These veins locally contain the sulfide minerals found in the massive sulfide bodies except bornite and pyrrhotite. Argentiferous tennantite is common in these veins.

Table 17 shows the mineralogic similarity of several massive sulfide deposits.

TABLE 17.—*Comparison of the mineralogy of the Iron King and United Verde deposits with other massive sulfide deposits*

Iron King deposit, Arizona	United Verde deposit, Arizona	Huelva, Rio Tinto district, Spain (Bateman, 1927)	Shasta County deposits, California (Graton, 1909)	Sullivan deposit, Canada (Swanson and Gunning, 1945)	Rammelsberg deposit, Germany [1] (Lindgren and Irving, 1911)	Mandy deposit, Manitoba (Hanson, 1920)	Deposits at Kyshtim, Russia (Stickney, 1915)
Sulfide minerals							
Pyrite Arsenopyrite Sphalerite Galena Chalcopyrite Tennantite	Pyrite Arsenopyrite Pyrrhotite Sphalerite Tennantite Bornite Galena Chalcopyrite	Pyrite Arsenopyrite Sphalerite Galena Tetrahedrite Enargite Luzonite Famatinite Chalcostibite Whitneyite Umangite Hauchecornite Ullmanite Berthierite(?)	Pyrite Chalcopyrite Sphalerite Galena Bornite	Pyrite Arsenopyrite Sphalerite Chalcopyrite Galena Bornite Chalcocite Tennantite	Sphalerite Chalcopyrite Galena Pyrite Arsenopyrite	Pyrite Arsenopyrite Sphalerite Chalcopyrite Galena	Pyrite Chalcopyrite Sphalerite Tennantite Galena Bornite (rare)
Gangue minerals and hydrothermal alteration products							
Quartz Ankerite Sericite Apatite(?)	Quartz Ankerite-dolomite Chlorite Magnetite Specular hematite	Sericite	Quartz Calcite Barite	Chlorite Quartz Specularite Dolomite	Barite	Sericite Carbonates Quartz Chlorite Rutile	Quartz Barite Sericite Rutile

[1] Listed in order of abundance.

ALTERATION

The host rock is commonly silicified at some margins of the massive sulfide deposits. At the United Verde mine, silicified rock is largely restricted to the north contact where flinty quartz forms a narrow screen between the hanging-wall gabbro or Grapevine Gulch formation and the massive sulfide bodies. At the Iron King mine, fine-grained quartz is present at the north ends of the veins.

Sericitization of the host rock is common adjacent to the massive sulfide deposits. At the United Verde mine the host rocks are largely composed of sericite, quartz, and a minor amount of chlorite. At the Iron King mine, sericitization and minor silicification are in the zone of veins and in the zone of hydrothermal alteration lying to the west.

Widespread chloritization of the host rock is common on the south (footwall) side of the United Verde pipe of massive sulfide, and tuffaceous sedimentary rocks of the Grapevine Gulch formation and the quartz porphyry are converted in part or entirely to chlorite rock that forms appreciable masses. The chlorite rock is weakly schistose, mimetic after the foliation in the quartz porphyry and tuffaceous sediments. This rock has been called "black schist" by the geologists of the United Verde mine, and this term is used in this report for the chlorite-rich alteration product.

Chloritization probably followed the formation of the massive sulfide pipe because the black schist is localized on the footwall side of the pipe, and nonchloritized inclusions of Grapevine Gulch formation and quartz porphyry lie within the pipe on upper levels of the mine.

Quartz-carbonate veins and carbonate nodules and augen are common, and indicate formation of carbonate minerals throughout the period of mineralization.

ORE SHOOTS

It should be emphasized that the massive sulfide bodies in the United Verde and Iron King mines have not constituted ore unless enriched in base or precious metals. Much of the enrichment was primary and caused by the addition of metals in the fracture zones in the pyrite-rich facies resulting in replacement of some or much of the pyrite and associated quartz and carbonate gangue.

At the United Verde mine, the bulk of the chalcopyrite was not deposited until after the large bodies of black schist (chlorite rock) on the footwall side of the massive sulfide pipe were formed, and on the lower levels in particular, fractured black schist contains much chalcopyrite, resulting in large and important shoots of "schist ore." Locally, fracturing and introduction of chalcopyrite extended into the quartz porphyry on the footwall (south) of the black schist and some of the smaller ore shoots consist of chalcopyrite-bearing quartz porphyry. The largest and richest ore shoots in the United Verde mine occur in the massive sulfide in the intermediate and upper levels.

Sphalerite is locally abundant in the United Verde mine, appearing in the pyrite facies of the massive sulfide body, but no zinc ore has been mined except in the high-zinc ore shoot in the north ore body on the lower levels. Small amounts of sphalerite and galena are present in all the copper ore shoots, whether in the massive sulfide pipe, black schist, or quartz porphyry.

The ore shoots at the Iron King mine are near the north ends of the massive sulfide veins, and locally include parts of the quartz noses where the precious-metal content may be higher than average. The ore shoots formed by the introduction of galena and sphalerite along fracture zones in the earlier pyrite, replacing pyrite and carbonate.

Although primary enrichment in chalcopyrite was important in forming the ore shoots of the United Verde Extension mine, secondary enrichment in Precambrian time formed the rich chalcocite ore bodies. On the lower levels, the primary enrichment was similar to that of the United Verde mine.

Secondary enrichment was important in the upper levels of the United Verde mine, possibly enriching the chalcopyrite ore as deep as the 500 level, but detailed records of early mining activity that might have information on this matter are lacking.

The precious metals were residually enriched during oxidation of the massive sulfide deposits of the United Verde mine and high concentrations of native silver were found overlying the sulfide minerals in the open pit. Practically all the soft gossan at the United Verde mine was minable; the parts relatively low in gold and silver had sufficiently high silica for the ore to be used for converted flux. Much of the oxidized top of the United Verde Extension ore body was mined for copper and precious metals, yielding about one-eighth of the production of the mine. At the Copper Chief mine the soft limonitic gossan was like the gossan at the United Verde mine, except that the lead content was higher. The Copper Chief gossan was mined chiefly for the precious-metal content, like the gossan at the Iron King mine.

ORIGIN

All previous investigators (Reber, 1922, 1938; Fearing and Benedict, 1925; Lindgren, 1926) of the massive sulfide deposits in the Jerome area are unanimous in concluding that these deposits are of the replacement or metasomatic type and that schistose rock is the replaced material. We agree with this conclusion, although we place other interpretations on some features they considered as evidence of replacement.

In the United Verde mine, tuffaceous and cherty sedimentary rocks of the Grapevine Gulch formation and quartz porphyry, both foliated, are the host rocks that have been replaced by massive sulfide deposits. The distribution of the sedimentary rocks in relation to the massive sulfide deposits indicates replacement rather than displacement of the sedimentary rocks. The Grapevine Gulch formation on most of the levels, if not all, intertongues with, and abuts against the massive sulfide. On certain levels, such as the southwestern part of the 1,650 level (pl. 7), the bedding in the Grapevine Gulch formation strikes into the massive sulfide, indicating replacement. The extent of replacement that we ascribe to is shown on figure 10 by the assumed trace of the former Deception rhyolite-Grapevine Gulch formation contact through the massive sulfide.

Evidence of replacement origin is also given in the upper levels of the United Verde mine by the concordant structural orientation of unsupported rock masses within the massive sulfide pipe with the host rocks adjacent to the sulfide pipe. Near the margins of the massive sulfide, many small masses and streaks of unreplaced rock retain their original orientation, and locally layering of the sulfide and gangue minerals in the massive sulfide reflects the orientation of the foliation in the contiguous wall rocks. Relict bedding and drag folds in the Grapevine Gulch formation were recognized on the hanging wall (north side) of the massive sulfide pipe on the upper levels, proving replacement of the sedimentary rocks by the massive sulfide. Much of the banding in the massive sulfide is not relict schistosity, as has been assumed by earlier investigators, but represents closely spaced parallel fractures along which sphalerite, galena, or chalcopyrite have replaced earlier pyrite. In places the banding in the massive sulfide parallels schistosity in adjacent rocks, but in many others it is discordant.

The contacts of the massive sulfide bodies with the host rock are usually sharp but in the United Verde mine, gradational contacts between black schist and massive sulfide are common. The pyrite crystals are coarse in the gradational facies, and seem to represent pyrite introduced into the black schist after the massive sulfide was formed. The contact of the massive sulfide

pipe shows all the embayments and irregularities characteristic of the walls of clear-cut sulfide replacement bodies in limestone.

Evidence of the replacement origin of the Iron King veins is less satisfactory than for the United Verde pipe. The absence of crustification and notable quantities of gangue minerals indicate that the veins are not fissure fillings. In some narrow veins, banding in the massive sulfide parallels the foliation in the chloritic schistose host rock (Spud Mountain volcanics) suggesting that the banding is perhaps relict foliation. Much banded ore, however, resulted from fracturing of early pyrite, quartz, and ankerite, and later introduction of sulfide that replaced the pyrite and ankerite. The gradation southward from massive sulfide to disseminated pyritic schist suggests that the massive sulfide veins formed by replacement. But in the final analysis, the theory of replacement origin is based largely by analogy to the United Verde deposit where the evidence is compelling.

The mineralizing solutions, presumably, were hydrothermal, as shown by the associated sericite at the Iron King mine and widespread chlorite at Jerome. Several chemical analyses of the Jerome chlorite and black schist contain 9 to 10 percent of water of crystallization (+105°C), indicating that the temperatures during mineralization were not high, possibly belonging to Lindgren's mesothermal zone.

The source of the mineralizing solutions and metals is debatable. Lindgren (1926, p. 35) concluded that no evidence is available to connect the solutions definitely with the Bradshaw granite, but they probably came from great depths, perhaps from granitic magma, and from levels where mineralizers such as fluorine, boron, and phosphorus had not separated to any great degree. Reber (1938, p. 59) stated that our general knowledge of metalliferous ore deposits makes it almost certain that the solutions were related to igneous rocks or magma reservoirs, but that the particular igneous rock or inferred magma reservoir to which the ore solutions are related is speculative. Reber suggested that the proximity of the massive sulfide deposits to the Bradshaw granite makes this granite a plausible deep-seated source for solutions, and that the ore solutions probably had a common deep-seated origin in the granitic magma of the Bradshaw.

The quartz diorite in the southern part of the Mingus Mountain quadrangle is part of the Bradshaw granite terrane of the Prescott-Jerome area. At the Copper Chief ore deposit, granodiorite porphyry dikes cut the massive sulfide lens and are younger (Reber, oral communication). These dikes appear to be satellites of the quartz diorite to the south, suggesting that the massive sulfide lens at the Copper Chief deposit is older than the quartz diorites; perhaps it can be presumed that all the massive sulfide deposits in the Jerome area are older than the quartz diorite.

At Jerome, the massive sulfide deposits are clearly younger than the quartz porphyry and gabbro, but older than eastward-trending andesitic dikes that cut both the gabbro and massive sulfide. These dikes only occur near the United Verde and United Verde Extension mines, and they are only distinguished on the mine maps (pl. 7). The general similarity in chemical composition of the andesitic dikes and gabbro as well as their close spatial relationship may be used as evidence that the dikes are genetically related to the gabbro and were intruded shortly after the emplacement of the gabbro. The massive sulfide deposits therefore would be closely related in time to the intrusion of the gabbro, and possibly, genetically related. Gabbro is widespread in the Jerome area, and any one massive sulfide deposit could be related to the nearest gabbro body.

The gabbro may be part of the early igneous activity culminating in the emplacement of the composite batholith of central Arizona (Bradshaw granite), and it may be academic to debate as to the relative importance of the gabbro and quartz diorite in relation to the ore-forming solutions. Granting that these solutions are of magmatic source does not answer the question as to the source of the copper, iron, magnesia, and sulfur introduced into the host rock to form the massive sulfide and black chloritic schist. These elements could either have been derived from deep magma chambers or picked up from country rocks and transported by the solutions for deposition above. If our structural interpretation is correct, an appreciable thickness of Shea and Gaddes basaltic flows should be present in depth below the massive sulfide deposits at Jerome, which are in the youngest rocks of the Ash Creek group. These basaltic flows could provide an ample source of iron and magnesia. It does not appear likely, however, that the copper in the Jerome deposit was derived from them.

The copper content was determined for the Gaddes basalt from Black Canyon, about 7 miles south of Jerome, away from the zones of pronounced alteration and metallization (table 18).

TABLE 18.—*Copper content of Gaddes basalt, in percent*

[Gaddes basalt from Black Canyon. Spectographic determinations by K. J. Murata]

Sample no.	*Percent Cu*
1	0.0023
2	.0012
3	.0012
4	.0030
Average	.0019

The average copper content of the earth's crust was estimated to be 0.01 percent by Clarke and Washington

(1924, p. 10), an estimate that has been widely accepted. Later, Sandell and Goldich (1943, p. 181) found 0.007 percent copper as the average, and that mafic igneous rocks contain about 10 times more copper than silicic rocks.

The copper content of 24 samples of Keweenawan lava averaged 0.012 percent, and 15 other subsilicic rocks averaged 0.0149 percent (Sandell and Goldich, 1943).

The average content of copper in the Gaddes basalt, 0.0019 (table 18), is below that of the average of the earth's crust and far below that normally present in basalts. This low-copper content does not substantiate the theory that the Gaddes basalt was the source.

FISSURE-VEIN DEPOSITS

Production from the fissure-vein deposits in the Jerome area has been sporadic, and none of the mines located in veins were active during the period of this study. Most of the available information concerning these veins is obtained from Lindgren (1926) and Reber (1938).

The gangue minerals of the veins are quartz, calcite, ankerite, and barite. The sulfide minerals are pyrite, arsenopyrite, chalcopyrite, galena, sphalerite, tetrahedrite (or tennantite), and bornite. Some veins formed through open-space filling and show local banding and crustification, and the sulfide minerals fill their centers or collect in local aggregates scattered through the gangue. Although replacement may have been locally important, in forming these veins, their general characters are commonly those of fissure veins.

The gold-quartz veins in the Cherry Creek district are chiefly north-striking shear zones locally mineralized with lenses composed of quartz, subordinate sulfide and ankerite, and minor tourmaline. Free gold was common in the massive quartz, but some gold was derived from oxidized sulfide deposits. The ore formed pockets or irregular small ore shoots and mining was largely limited to the oxidized zone which was locally as deep as 300 feet. Small quantities of chalcopyrite, bornite, sphalerite, and galena have been observed in the primary parts of the veins, which are discussed in more detail in the section on the Cherry Creek mining district.

Copper-silver veins with general easterly strikes have produced some ore from the Shea and Yaeger mines. The Shea vein, on the east side of Mingus Mountain, consists of coarsely crystalline quartz containing bunches of ankerite and pyrite with some chalcopyrite. Locally, a little arsenopyrite and argentiferous tetrahedrite are present, and where sulfide deposits are abundant, the vein has a banded structure. A little galena is present near the ore shoot that produced high-grade silver-copper ore from the abundant argentiferous tetrahedrite.

The Yaeger mine on the west side of Mingus Mountain has been the most productive of any of the fissure-type veins in the district. It contains some high-grade bornite ore in an east-striking calcite-quartz vein. Tennantite and a little pyrite are present, as well as some secondary chalcocite in the upper part of the vein. Some gold and silver were recovered. The Shylock vein, south of the Yaeger mine, contains tetrahedrite, galena, and sphalerite in massive quartz. No production has been recorded.

The quartz-copper vein in the Verde Central mine, south of Jerome, may not belong to the fissure-type vein, for although it contains considerable white quartz, most of it is greenish-gray, fine-grained quartz that may represent silicified Deception rhyolite (Fearing and Benedict, 1925). Pyrite and chalcopyrite are irregularly distributed in the quartz. In addition to the quartz-chalcopyrite vein, chalcopyrite and pyrite were found in black chloritic schist formed along the contact of the Deception rhyolite and quartz porphyry, resembling the "black schist ore" on the footwall side of the United Verde massive sulfide pipe.

West of the Iron King mine, the Silver Belt-McCabe vein strikes northeast; it was mined chiefly for the gold-silver content. Ankerite or siderite are present next to the vein walls, and the sulfide minerals fill the centers of the veins. Argentiferous galena and argentiferous tetrahedrite are the important sulfide minerals, but some sphalerite and pyrite are present.

Most of the veins in the area studied are clearly of Precambrian age; in the Cherry district and east and west of Mingus Mountain, nonmineralized Paleozoic sedimentary rocks overlie or crop out near the veins. West of Humboldt where the Silver Belt-McCabe crops out, however, there are no Paleozoic sedimentary rocks to date the mineralization. Lindgren (1926, p. 127) suggested that these veins may be related to unmetamorphosed rhyolitic dikes presumably younger than Precambrian and possibly of Late Cretaceous or Early Tertiary age.

VERDE MINING DISTRICT

The Verde mining district is located on the eastern slope of the Black Hills, parallel to the northwest front of the range. The district is about 7 miles long and about 3 miles wide. Two mines, the United Verde and the United Verde Extension, located at Jerome, have contributed more than 99 percent of the total production of the district, and commonly the district is called the "Jerome district."

Annual production data on the Verde mining district for 1908–51 are given in table 19. Earlier annual production data are not available. From production data assembled by Elsing and Heineman (1936, p. 101) and combined with the data in table 19, it appears that from 1883 to 1951 the dollar value of recoverable metals from the Verde district has been about $650,000,000. For the same period, the following amounts of metals have been produced: gold, about 1,565,000 ounces; silver, more than 57,000,000 ounces; copper, about 1,860,000 tons.

UNITED VERDE MINE

PRODUCTION

Underground production at the United Verde mine totaled 22,519,846 tons to January 1, 1952.[5] From 1900 to 1910, the annual production averaged 245,000 tons, and except for a shutdown in 1921–22, when the mine

[5] The following information on production and exploration at the United Verde mine was obtained from a mimeographed pamphlet prepared by C. E. Mills, general superintendent, United Verde Branch, Phelps Dodge Corp., and distributed to members of the Underground Mining Branch, Arizona section, Am. Inst. Min. Metall. Eng., March 1947, and from subsequent annual reports of Phelps Dodge Corp.

TABLE 19.—*Production of gold, silver, copper, lead, and zinc from Verde (Jerome) district in terms of recoverable metals* [1]

Year	Number of mines producing	Ore treated or sold (short tons)	Gold (ounces)	Silver (ounces)	Copper (pounds)	Lead (pounds)	Zinc (pounds)	Total value
1908	1		20, 335	494, 574	36, 183, 089			$5, 460, 818
1909	1		17, 021	495, 479	36, 695, 259			5, 370, 766
1910	2		19, 316	398, 247	38, 663, 880			5, 624, 449
1911	4		15, 240	461, 145	33, 167, 987			4, 705, 395
1912	2		15, 083	484, 222	31, 565, 539			5, 817, 855
1913	2	393, 866	20, 667	641, 074	35, 334, 694			6, 291, 252
1914	1		21, 401	646, 572	32, 448, 170			5, 115, 499
1915	5		28, 405	922, 273	50, 266, 821			9, 851, 397
1916	3		37, 725	[2] 1, 250, 000	[2] 103, 000, 000			[2] 26, 992, 986
1917	6	949, 094	36, 443	1, 884, 673	145, 933, 703	205, 817		42, 177, 535
1918	7	1, 011, 966	37, 327	1, 986, 845	133, 780, 865	7, 962		35, 802, 862
1919	4	566, 445	21, 148	949, 436	72, 332, 704			14, 954, 395
1920	4	870, 129	24, 363	1, 275, 323	107, 084, 032			21, 597, 163
1921	3	189, 474	4, 671	319, 646	26, 567, 602			3, 845, 288
1922	3	572, 246	20, 345	843, 151	74, 793, 982			11, 360, 899
1923	4	1, 465, 006	84, 418	2, 480, 692	140, 425, 401			24, 421, 705
1924	3		52, 030	1, 839, 844	142, 238, 947			20, 941, 417
1925	4		56, 740	2, 097, 951	151, 920, 324			24, 201, 438
1926	2		55, 538	1, 971, 914	150, 269, 756			23, 416, 169
1927	3		53, 513	1, 920, 901	141, 261, 930			20, 700, 536
1928	3		61, 677	2, 195, 595	163, 312, 041			26, 076, 159
1929	4		75, 404	2, 749, 290	211, 542, 611			40, 255, 410
1930	3		[2] 50, 796	[2] 1, 530, 000	[2] 117, 600, 000			[2] 16, 925, 000
1931	2		[2] 18, 141	[2] 1, 050, 000	[2] 44, 100, 000			[2] 4, 525, 000
1932	3	241, 113	10, 086	361, 273	35, 790, 460			2, 565, 165
1933	3	243, 245	12, 288	338, 860	33, 201, 565			2, 497, 543
1934	1	200, 754	8, 065	242, 632	26, 147, 463			2, 530, 505
1935	4	851, 993	30, 850	1, 049, 934	75, 535, 518			8, 103, 822
1936	5	1, 312, 012	64, 405	1, 911, 774	99, 157, 152			12, 857, 309
1937	2	1, 326, 940	62, 748	1, 840, 150	84, 742, 620			13, 873, 393
1938	3	799, 564	45, 541	1, 144, 652	57, 863, 745			8, 004, 559
1939	4	999, 023	40, 312	1, 327, 472	75, 430, 241			10, 156, 737
1940	6	896, 727	24, 652	1, 144, 028	74, 459, 646			10, 090, 291
1941	6	1, 121, 004	32, 047	1, 544, 317	84, 484, 800			12, 189, 032
1942	6	1, 242, 133	28, 429	1, 532, 108	87, 874, 000			12, 717, 268
1943	5	898, 699	18, 117	1, 036, 194	68, 467, 100			10, 271, 667
1944	5	525, 263	8, 620	589, 538	52, 429, 000			7, 798, 842
1945	5	392, 295	8, 602	475, 290	40, 224, 000			6, 069, 294
1946	4	350, 465	8, 132	418, 578	32, 351, 000			5, 863, 693
1947	3	350, 645	6, 931	367, 778	29, 205, 600			6, 708, 600
1948	4	358, 491	11, 374	408, 669	29, 087, 800		917, 000	7, 201, 970
1949	2	410, 607	10, 790	509, 828	34, 429, 800	71, 600	8, 700, 000	8, 711, 854
1950	2	361, 320	9, 412	456, 254	26, 581, 400	100, 000	15, 600, 000	8, 500, 299
1951	1	299, 729	7, 325	408, 891	19, 484, 000	205, 000	20, 310, 000	9, 073, 455
Total			1, 296, 473	47, 997, 067	3, 287, 436, 338	590, 379	45, 527, 000	$572, 216, 691

[1] Figures for 1908–31 are from Elsing and Heineman (1936, p. 86). Figures for 1932–51 are from Minerals Yearbooks, U. S. Bureau of Mines.
[2] Estimated.

was closed, underground production increased to a maximum of 1,062,500 tons in 1924. Underground mining stopped in 1931 owing to the low price of copper and did not start again until April 1937. Production then increased to a maximum tonnage of 1,015,200 tons in 1942. Since then, the rate of production has decreased steadily. In 1949, copper-zinc ore was mined for selective flotation and by January 1, 1952, 489,988 tons of copper-zinc ore had been mined, compared with 570,753 tons of copper ore for the same period. In 1951, copper-zinc ore comprised much of the production, 211,578 tons as compared with 87,858 tons of copper ore. The mine and concentrating mill closed in February 1953.

Approximately 15,542,000 cubic yards or 32,000,000 tons of rock was stripped from the open pit to produce 8,153,000 tons of direct shipping ore. The stripping operation included oxidized ore placed on surface dumps and later used for siliceous flux; by January 1, 1947, the smelter had used 1,688,200 tons of this dump ore. The open-pit mining was completed by April 1, 1940, and bottomed at the 630 level.

DEVELOPMENT AND MINING

According to C. E. Mills, the development throughout the period that the property was in operation totaled 81 miles of underground workings: 304,470 feet of drifts, 123,800 feet of raises, and 14,100 feet of shafts. This represents about 1 ton of development for each 46 tons of ore mined underground.

The upper levels are spaced 100 feet apart, except for the 1,200 level which is 200 feet below the 1,000 level. The remaining levels to the 3,450 level are 150 feet apart. The lower levels are the 3,750, 4,050, 4,200, and 4,500 levels. The 4,050 and 4,200 levels are not complete and extend only from shaft 8 to the North ore body.

Eight shafts were sunk since mining operations started. Shaft 1, which was started in 1883, was sunk to the 500 level and used until 1894. In 1900 shaft 2 was sunk to the 700 level, deepened to the 1,000 level in 1909 and abandoned in 1912. Shaft 3 is a three-compartment timbered shaft, started in 1901 and sunk to the 1,950 level by 1917. This shaft was abandoned after completion of shaft 5 and was used later as a mine ventilation raise. In 1910 shaft 4, a timbered shaft, was started; in 1914 it was extended to the 1,000 level and was abandoned in 1921.

Shaft 5, which has three compartments and is of concrete construction, served for hoisting. It was begun in 1915 and completed to the 3,150 level in 1927. A timbered section below the 3,150 level was sunk to the 3,515 in 1931. A double-drum hoist (5 by 10 feet) with an 800-horsepower motor handled 7.5-ton ore skips. The hoist room is on the 1,000 level and the ore pocket at the 800 level.

Service shaft 6, of concrete construction, was started in 1919 and completed to the 3,000 level in 1929. A single-drum hoist (6 by 12 feet) with a 500-horsepower motor handled a double-deck cage in an 8 by 13-foot service compartment. The hoist room is on the 500 level, and the sheave room is on the 400 level.

Shaft 7, which has five compartments of concrete construction, was started in 1929 and completed to the 3,000 level in 1933. A double-drum hoist (7 by 12 feet) equipped with a 2,700-horsepower motor to handle 10-ton ore skips was installed in the hoist room at the surface (300 level). An ore pocket was built at the 800 level. This shaft was sunk to replace shafts 5 and 6 when it was feared that mine subsidence might make these shafts inaccessible. However, both shafts have been kept in operation and it has not been necessary to use shaft 7.

Shaft 8 is a two-compartment timbered shaft from the 2,850 to the 3,300 level and has three compartments from the 3,300 to the 4,630 level. This shaft was started in 1940 and completed in 1942. The hoist room is on the 3,000 level and is equipped with a double drum (3⅓ by 7 feet) driven by a 450-horsepower motor. Skips in tandem with cages handle a 3½-ton load. The dump pocket is below the 2,850 level, and small loading pockets are below the 3,450, 3,750, 4,050, 4,200, and 4,500 levels. All ore below the 3,000 level was hoisted in shaft 8 to the 3,000 level and hauled to shaft 5.

Ore was transported from shaft 5 ore bins on the 1,000 level through the Hopewell tunnel, 6,600 feet long, to ore bins on the surface. The Hopewell tunnel is 9 by 11 feet in cross section. Ore was transported by 25-ton trolley locomotives hauling 40-ton cars on standard-gage track.

Men and supplies were transported through the 500-foot level adit connecting the 500-level surface plant with shaft 6. This adit is 1,600 feet long and 8 by 9 feet in cross section.

Early underground mining was done chiefly by horizontal cut and fill stoping, leaving vertical pillars where ore bodies are large (Cowley and Quayle, 1930). The vertical and floor pillars on the middle levels were later mined by square-set method. Starting in 1937 when the mine was reopened, scrapers were introduced and the mining method was changed to inclined cut and fill stoping. For the pillars and remnants of ore in the upper levels, a modified Mitchell slice system was adopted (Pullen, 1941).

The zinc-copper stope in the ore body 800 feet north of the main sulfide pipe was mined from the 4,500 to the 3,750 levels, using diamond-drill long-blast holes

and drawing ore without filling the stope with waste (Little, 1950).

SMELTING OPERATIONS

The smelter at Clarkdale had a daily ore capacity of 5,000 tons and was operated until June 1950 when it was partly dismantled. A concentrating mill for handling black schist ore started operations in 1927 (Barker, 1930), and late in 1948 one section was converted to differential flotation for the separation of sphalerite and chalcopyrite from the North ore body. The history of smelting (Lanning, 1930) and descriptions of the crushing plane (Keefe, 1930), roasting plant (Parsons, 1930), Cottrell plant (Denny, 1930), blast-furnace smelting (Kuzell, 1930), reverbatory smelting (Mooney, 1930), and converters (Williams, 1930) are available.

GEOLOGY

OLDER PRECAMBRIAN ROCKS

Two formations of the Ash Creek group, Deception rhyolite and Grapevine Gulch formations are exposed in the underground workings of the United Verde mine and on the surface adjacent to the mine. Quartz porphyry, gabbro (United Verde diorite of Reber, 1922), and andesitic dikes are associated intrusive rocks.

Deception rhyolite

The Deception rhyolite crops out in discontinuous exposures from the Hull fault to the south to the 500 level surface workings to the north (pl. 5). The intrusive quartz porphyry has cut out much of the Deception rhyolite south of the mine. A small patch of Deception rhyolite, partly covered by Paleozoic sedimentary rocks, crops out in the Verde fault zone at the east margin of the gabbro. This patch of rhyolite must have been down-faulted as shown by the capping of Paleozoic rocks, and before faulting was not closely associated with the Deception rhyolite now exposed to the south.

The Deception rhyolite occurs west of the Verde fault near the portal of the 500 level adit in the northern part of the United Verde mine area (pl. 5). Some andesitic (or basaltic) rock is intercalated with the rhyolite, like other mafic flows in the Deception rhyolite in Mescal Gulch.

In the underground workings of the United Verde mine, rhyolite is exposed on the 1,650, 1,000, and the 500 levels (pl. 7) and in the open pit. These workings expose three separate masses of rhyolite. One is in the haulage adit on the 500 level and on the surface near the portal. The second mass extends from the surface in the pit downward through the 500 level, where it is exposed 500 feet east and 600 feet northeast of the massive sulfide, and through the 1,000 level where it is cut 800 feet east of the massive sulfide. It is a thin wedge along the east side of a larger mass of Grapevine Gulch formation; both trend from northward to north-northeastward, and dip very steeply eastward. The third mass presumably extends from the surface in the pit downward through the 1,000 and 1,650 levels, southeast of the massive sulfide (not shown on pl. 7). On these levels, the north margin of the rhyolite is near 800 S., whereas on the surface, it is at 500 N. (pl. 5), indicating a southward plunge. This mass is bounded on the east by the Verde fault, and on the other sides by quartz porphyry.

Breccia structure locally occurs in the rhyolite on the 1,000 and 1,650 levels. Elsewhere the rhyolite is massive, similar to most of the Deception rhyolite near Jerome that has been hydrothermally silicified and sericitized.

Grapevine Gulch formation

The surface exposures of the Grapevine Gulch formation are west of the Deception rhyolite. The Grapevine Gulch formation is (1) in lenticular masses associated with Deception rhyolite surrounded mostly by quartz porphyry, (2) in a screen separating quartz porphyry from the gabbro (United Verde diorite of Reber, 1922), and (3) in an irregular mass that intertongues with the quartz porphyry east of the massive sulfide (pl. 5). Although now largely covered by dump, the Grapevine Gulch formation west and north of the gabbro occurs on both sides of the Haynes fault.

Most of the Grapevine Gulch formation in and near the United Verde mine consists of the fine-grained tuffaceous sedimentary facies with conspicuous bedding. Crystal tuff, chert, cherty shale, and some slate are the dominant rock types. Locally, a little limestone is present. The lithic tuffaceous facies is exposed in a few places underground, but absent from most surface exposures. Reber (oral communication) noted conglomeratic bands in the area now covered by dump.

Rhyolitic breccia intercalated in the Grapevine Gulch formation is exposed west of the Warrior fault and south of the Haynes fault (pl. 5). Flow-banded fragments 3–4 inches across are conspicuous, and no bedding is recognizable. In the western exposure, some of the rhyolite is strongly foliated and the clastic structure is not recognizable. Possibly some of these rhyolitic rocks represent lava flows. Diamond drilling shows that the rhyolitic rocks disappear at depth, only to reappear at greater depths. Two interpretations are possible: (1) the rhyolite is the Deception rhyolite exposed in the crest of an anticline, or (2) the rhyolite is intercalated between tuffaceous sedimentary rocks of the Grapevine Gulch formation, and represents a return of volcanism after the start of deposition of the rocks of the Grape-

vine Gulch formation. The second interpretation is favored because of the probable lenticular nature of the rhyolitic rocks revealed by drilling.

Quartz porphyry

The quartz porphyry in the United Verde mine area occurs in three masses, southeast, north, and west of the main body of gabbro. The southeast mass is the northern part of the large quartz porphyry body that extends southward beyond Mescal Gulch. This large mass forms the footwall of the main massive sulfide pipe, and is present from the surface to the lowest level in the mine.

About 400 square feet of quartz porphyry is exposed northwest of the portal of the 500 level adit. This quartz porphyry and associated Grapevine Gulch formation and Deception rhyolite occupy a trough in the gabbro, which is concealed by the Paleozoic rocks to the north. On the lower levels north of the gabbro near the north ore body, a much larger mass of quartz porphyry is present. It was probably connected to the quartz porphyry northwest of the 500 level adit before the intrusion of the gabbro. On the 3,300 level, diamond drills cut through 350 feet of quartz porphyry north of the gabbro (pl. 7, 2,500 N.; 500–1,000 W.). On the 3,750 level, quartz porphyry is exposed for 800 feet north of the gabbro, and on the 4,500 level (pl. 7) a similar amount is cut by drill holes and drifts. On the 4,500 level, part of the Grapevine Gulch formation separates the gabbro from the quartz porphyry. The exact size and shape of this northern mass of quartz porphyry is not known, but the horizontal area at depth is certainly greater than exposed at the surface.

The small mass of quartz porphyry west of the gabbro (pl. 5, 320 N.; 1,150 W.) is only 30 feet wide and 100 feet long, appearing as a tongue in the Grapevine Gulch formation. This small tongue probably was formerly connected with the larger mass of quartz porphyry exposed southeast of the gabbro, and that the intrusion of the gabbro separated them.

The quartz porphyry is well foliated south of the massive sulfide pipe, but is more massive to the east (pl. 5). Locally, feldspathic facies are present; the feldspar phenocrysts are now composed of a felt of sericite. The quartz porphyry from near the mine has the same mineralogy as the Deception rhyolite, that is quartz, sericite, and some chlorite, but the large quartz phenocrysts in the quartz porphyry aid in the distinction of the two rocks.

On many levels, a highly foliated purplish rock contains large quartz crystals and elongated sericitic aggregates probably pseudomorphs after feldspar. This purplish rock is present only as narrow locally persistent bands. In most places, the purple rock is at the margin of quartz porphyry and Grapevine Gulch formation. Its origin is somewhat in doubt; it may be a marginal facies of the quartz porphyry impregnated with hematite derived from the adjacent tuffaceous rocks, or perhaps a crystal tuff in the Grapevine Gulch formation, whose hematite pigment may be a contact metamorphic effect of the quartz porphyry.

At the surface in Hull Canyon, tuffaceous sediments of the Grapevine Gulch formation are locally purplish, where they are in contact with or form inclusions in the quartz porphyry, and locally, the quartz porphyry is purplish at the margin. If these rocks were highly foliated, the original character would be uncertain. Because of this doubt concerning their origin, the purplish rocks found in the United Verde mine are of dubious value for stratigraphic purposes.

Gabbro (United Verde diorite of Reber, 1922)

The term "United Verde diorite" was given by Reber (1922, p. 6) to designate the mafic plutonic rock exposed in the United Verde mine, but as was pointed out earlier in this report, the chemical composition compels designating the rock as gabbro. Because the gabbro is widespread in the Prescott and Mingus Mountain quadrangles, this variance with local nomenclature seems justified for this report.

At the United Verde mine, the gabbro forms a large northward-trending semiconcordant mass. To the north, the gabbro at the surface is cut off by the Haynes fault (pl. 5). Southward, the gabbro extends to Deception Gulch. At the surface, the eastern contact of the gabbro is essentially concordant with the bedding of the Grapevine Gulch formation; with depth, it dips west to the 3,000 level, and east below that level.

At the surface (pl. 5) 1,200 N.; 2,000 W. a small irregular-shaped mass of gabbro is separated from the main mass of gabbro by a belt of Grapevine Gulch formation and the Warrior fault. At depth, the two gabbro bodies probably are connected, for on the 1,500-foot level, no Grapevine Gulch formation is in the crosscut that cuts the gabbro to 2,050 W.; 1,700 N. (pl. 5).

The gabbro is, in general, very massive and well jointed, but along the margin of the gabbro, foliated facies are present; the foliation plane essentially parallels the contact.

Andesitic dikes

Many dikes, ranging from 1 inch to 50 feet in width, cut the Grapevine Gulch formation, quartz porphyry, and gabbro, as well as the massive sulfide pipe. The dikes do not intrude the overlying Paleozoic sedimentary rocks indicating that they are also of Precambrian age.

These dikes are called andesite following Reber (1922, p. 19). Lindgren (1926, p. 58) suggested that a more suitable term would be diorite porphyry, but the original texture and mineral composition cannot be determined owing to the altered nature of the dikes, and there is little merit in debating between andesite and diorite porphyry. The original dikes apparently had a fine-grained texture and were rich in mafic minerals and plagioclase.

The dikes trend eastward and dip almost vertically, except for a few larger ones on the lower levels that trend northeastward and are only on the northwest side of the massive sulfide.

The dike rocks are intensely altered so that the original mineralogic character is largely masked. In most of the thin sections examined, the rock consists of abundant sericite, dolomite, and chlorite and a few tiny quartz grains. Apatite prisms and leucoxene are accessory minerals. In two thin sections, polysynthetic twinned albite was recognized in randomly oriented euhedral crystals.

The chlorite in the dikes is pale green in thin section, with low birefringence. The optic angle is small (15°–20°), positive. The intermediate index of refraction ranges from 1.620 to 1.628; following Winchell (1936), this chlorite mineral is either rumpfite or ripidolite.

Weak foliation in the dikes is best shown in the older mine workings, where it is parallel to the stronger foliation in the adjacent quartz porphyry; it is recognizable in thin sections by a subparallel arrangement of the chlorite and sericite.

TABLE 20.—*Chemical analyses of andesitic dikes and related materials in United Verde mine*

[In columns 2, 3, and 4, the location in the mine is not known. Analyses furnished by courtesy of Phelps Dodge Corp.]

	1	2	3	4	5
SiO_2	41. 03	52. 00	45. 50	48. 40	2. 88
Al_2O_3	17. 20	19. 75	18. 41	27. 12	39. 33
Fe_2O_3	1. 52	3. 43	1. 28	1. 83	1. 12
FeO	8. 09	7. 86	19. 99	8. 64	------
MnO	. 22	------	------	------	------
MgO	5. 75	3. 62	6. 30	3. 55	. 72
CaO	10. 00	9. 40	Tr.	Tr.	. 60
K_2O	------	1. 97	1. 16	. 89	8. 42
Na_2O	------	1. 46	2. 04	1. 04	------
TiO	. 80	------	------	------	------
H_2O-	. 07	. 10	. 20	. 40	. 50
H_2O+	------	. 10	5. 10	6. 75	11. 20
Loss on ignition	11. 16	------	------	------	------
S	. 02	. 66	. 40	------	------
ZnO	. 81	------	------	------	------
Cu	. 02	------	------	------	------
CO_2	------	Tr.	. 25	1. 95	------
SO_3	------	------	------	------	35. 41

1. Andesite dike, 3.000 level.
2. Andesite dike, fresh material.
3. Partly kaolinized dike.
4. Kaolinized dike.
5. Alunitelike material from altered dike, 1,650 level.

In the upper levels of the United Verde mine, the andesitic dikes have been extensively altered to clay containing gypsum, and form "water courses." Chemical analyses of partly altered and altered dike rock from the upper levels are given in table 20. Presumably sulfuric acid solutions derived from the weathering of pyrite caused the clay alteration, as this type of alteration is not found in the lower levels. The acid solutions must have had a low-copper content, as no chalcocite was found in the adjacent sulfide-bearing rocks.

PALEOZOIC AND TERTIARY ROCKS

The Tapeats sandstone(?) crops out in a nearly continuous band west and north of the United Verde mine (pl. 5). It is an important marker for determining displacements along the many minor faults. One small patch is exposed in the Verde fault zone, northeast of the pit area.

The Martin and Redwall limestone beds overlie the Tapeats sandstone(?) west and north of the United Verde mine and also appear in the Verde fault zone to the northeast.

The Tertiary rocks include gravel overlain by basalt (Hickey formation) and are exposed only on the east (hanging-wall) side of the Verde fault (pl. 5).

MAJOR STRUCTURES

Folds

Evidence indicates that a major control for the location of the United Verde deposit is a small north-northwestward-plunging anticline (fig. 10) on the western flank of the north-northwestward-plunging Mingus anticline (fig. 3). The evidence, however, is not clear cut; a major difficulty is the obscuring effect of the intrusive bodies of quartz porphyry and gabbro.

Regional studies indicate the existence of the Mingus anticline, as discussed on page 64. The existence of the smaller fold localizing the mineralizing activity is indicated by the trace of the Deception rhyolite-Grapevine Gulch formation-contact, the configuration of the southeast margin of the gabbro, and the minor structural elements such as the minor folds, lineation, and foliation.

The trace of the Grapevine Gulch formation-Deception rhyolite contact provides an important link in the chain of evidence, if the interpretation of the course of the contact through the quartz porphyry (fig. 10) is correct. Southwest of the mine, the contact is well exposed in Deception Gulch (pl. 1); and here, excellent evidence shows that beds face west, indicating an anticlinal structure to the east. The contact strikes slightly east of north and is buried farther north by the Paleozoic sedimentary rocks. North of the Hull fault, tongues of quartz porphyry intrude the Grapevine

Gulch formation and Deception rhyolite, but the contact can be traced with assurance to 700 S. (pl. 5). Northward, quartz porphyry has removed all of the Deception rhyolite, but patches of Grapevine Gulch formation remain between the gabbro and quartz porphyry or massive sulfide.

East of the mine on the quarry bench (300 level of the mine), the contact between Grapevine Gulch formation

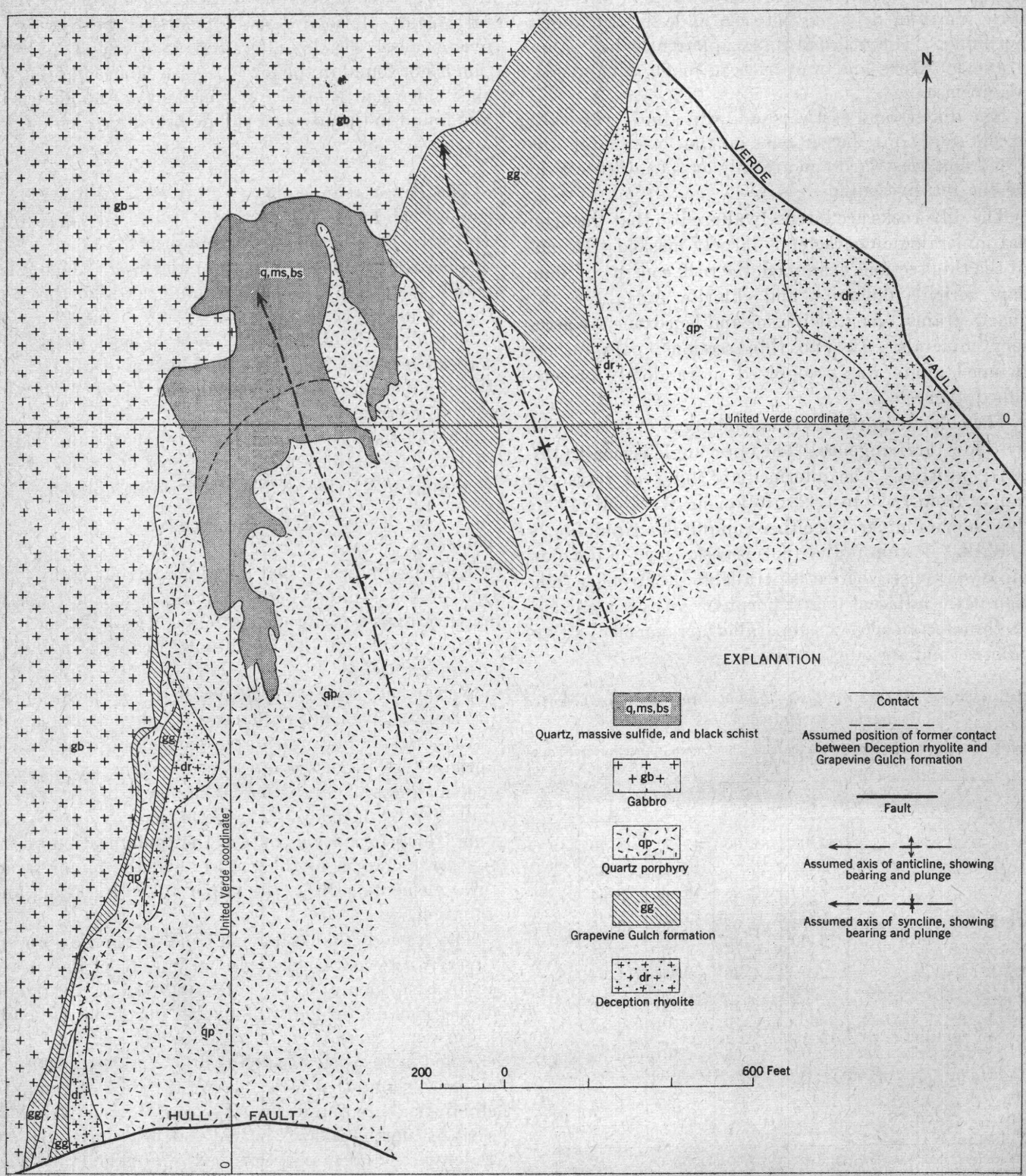

FIGURE 10.—Suggested structural interpretation at United Verde mine.

and Deception rhyolite is exposed for 950 feet (pl. 5, 200 S.–750 N.; 900 E.); the trend of the contact swings from north-northwest to north-northeast to within 200 feet of the Verde fault where it is cut out by quartz porphyry. The south end of the contact, which is at 200 S. (pl. 5), has shifted eastward from 100 W. to 1,000 E. The offset of this contact can be explained by faulting or by a north-northwestward-plunging anticline and parallel syncline lying to the east (fig. 10). No evidence of such a fault has been found, and it can be considered only as a remote possibility.

Minor folds are common in the Grapevine Gulch formation, and the abundant tongues of quartz porphyry cutting through the formation may in part be guided by them. They are best exposed in the southern walls of the open pit (pl. 5, 100 N.; 800–900 E.). The fold axes trend N. 20° W. and the folds plunge north-northwestward 40°–55° (pl. 4*B*). Foliation in the Grapevine Gulch formation and the quartz porphyry parallels the axial planes of the folds (pl. 4*A*), and the plunge of the intersection of bedding and foliation (lineation) (pl. 3*F*) is essentially the same as the plunge of the minor folds.

Northeast of the massive sulfide pipe (pl. 5, 650 N.; 500–700 E.) many minor folds in the Grapevine Gulch formation crop out. The foliation, lineation, and axes of the folds strike more nearly north in contrast to the N. 20° W. strike on the south wall of the pit. Most of the foliation in the quartz porphyry south of the United Verde mine strikes N. 20° W., indicating that this is the prevailing trend of foliation and bearing of lineation and fold axes. Northeast of the massive sulfide the northerly strikes are adjacent to the gabbro and their deviation in bearing is interpreted as the result of eastward "push" by the gabbro as it was emplaced. Local foliation along the gabbro parallel to its margin, also indicates movement of the gabbro after emplacement.

The axis of the massive sulfide pipe in the United Verde mine trends N. 20° W. and plunges 65° northward, coinciding very closely with the general bearing of the foliation, lineation, and minor folds and plunge of lineation and fold axes.

The gabbro is largely concordant wherever its contact with the Grapevine Gulch formation can be observed. The north-northeast trend of the gabbro mass is essentially that of the Grapevine Gulch formation. In detail, the gabbro contact parallels adjacent bedding in the Grapevine Gulch. The configuration of the gabbro contact in the United Verde mine has been determined from many diamond-drill holes and some underground workings. Northwest of the massive sulfide pipe the gabbro contact forms an inverted trough striking N. 20° W. and plunging 60°–65° to the north to the 3,000 level where it becomes straighter and strikes N. 45° E. (fig. 11).

The conformity of the bearing and plunge of the "inverted trough" in the margin of the gabbro with the bearing and plunge of the foliation, lineation, and minor fold axes appears to be more than fortuitous, particularly as the gabbro is essentially concordant wherever the contact can be observed.

The possibility that the gabbro has been folded to form the inverted trough is disproved by the lack of all but the local marginal foliation, which is discordant to the north-northwest regional foliation. In the Grapevine Gulch formation, the foliation which has an axial plane relationship to the folds, must have formed during the period of folding. The appearance of strongly foliated quartz porphyry at 400 N.–1,150 W. (pl. 5), west of the gabbro, and strongly foliated quartz porphyry southeast of the gabbro denies the possibility that the gabbro was a buttress and not susceptible to foliation. Instead, it shows conclusively that the nonfoliated gabbro separated the foliated quartz porphyry into two segments. Furthermore the gabbro is susceptible to foliation in the proper environment as shown by the foliated gabbro northwest of the Iron King mine (pl. 1).

Relict bedding in the Grapevine Gulch formation in the zone of mineral deposition is not abundant, but attitudes that can be measured are in harmony with the concept of a plunging north-northwest anticlinal fold where the massive sulfide pipe formed. On the 600 level, well within but near the west side of the massive sulfide pipe, bedding in unreplaced rocks of the Grapevine Gulch formation strikes N. 10° E. and dips 70° W. On the 900-foot level at the northeast corner of the massive sulfide, bedding within the sulfide pipe strikes N. 70° E. and dips 60° N. On the 1,000 level in the northern part of the massive sulfide pipe, some of the bedding in the tuffaceous rocks strikes nearly east and northeast (pl. 7). These more easterly strikes conform with the concept of a plunging anticline.

These data, though not conclusive, suggest that a drag fold on the flank of the major Mingus anticline controlled the form of the gabbro contact and, in turn, controlled the path of the later mineralizing solutions. The massive nonfoliated gabbro was undoubtedly less permeable to the mineralizing solutions than the underlying foliated quartz porphyry and Grapevine Gulch formation, and the "inverted trough" of the overlying gabbro channeled the solutions along the nose of the northward-plunging anticline.

An anticlinal structure localizing the massive sulfide pipe requires a complementary syncline to the east as

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the Grapevine Gulch formation and Deception rhyolite are in normal contact east of the mine. Furthermore, if the two formations form an anticline coinciding with the massive sulfide pipe, a bend of this contact to the southeast is needed to join the contact exposed on the 300-foot level bench (pl. 5). Confirmation of a small syncline to the east of the anticline is obtained on the 1,000 level (pl. 7, east of 500 E.) where sufficient under-

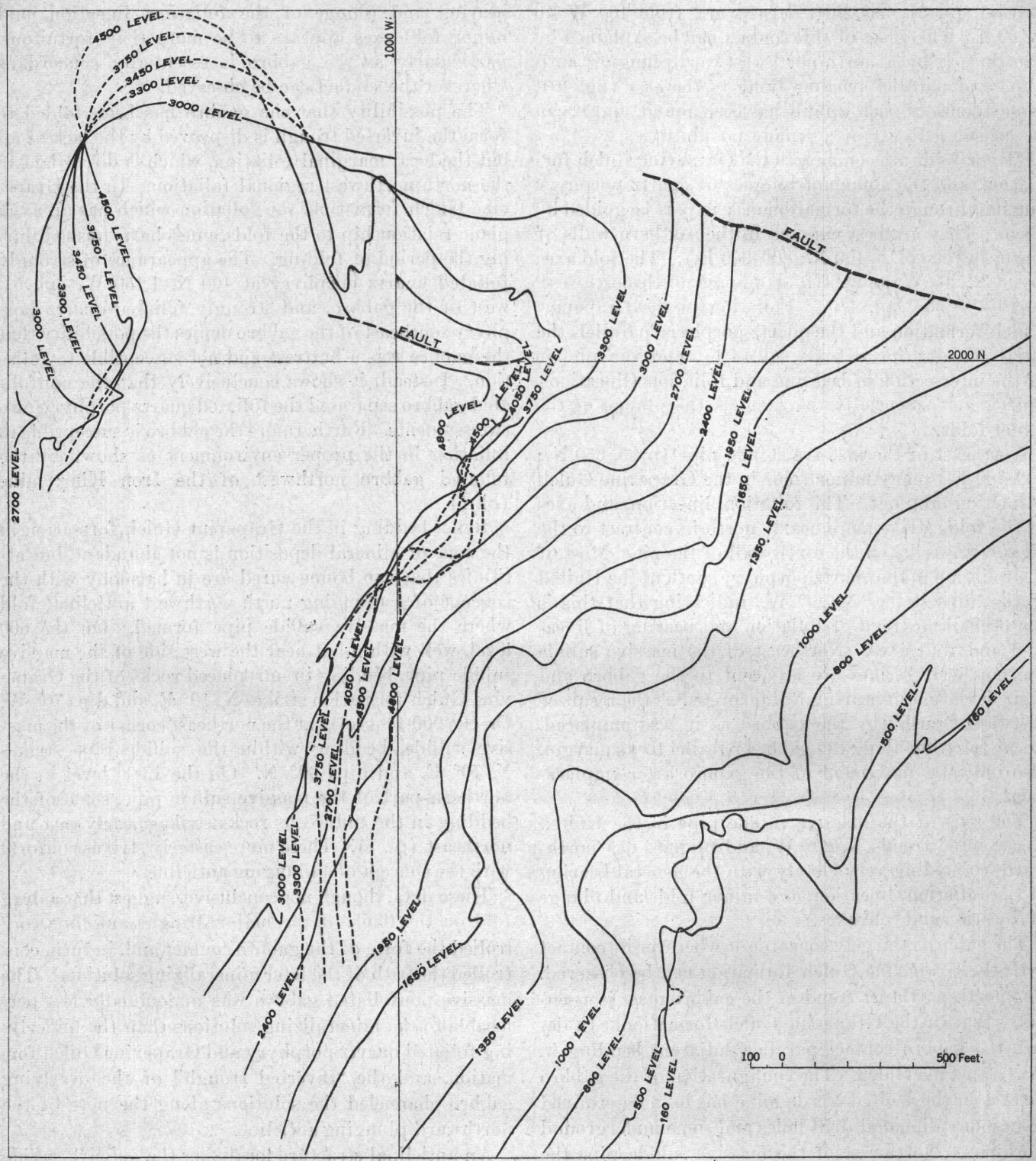
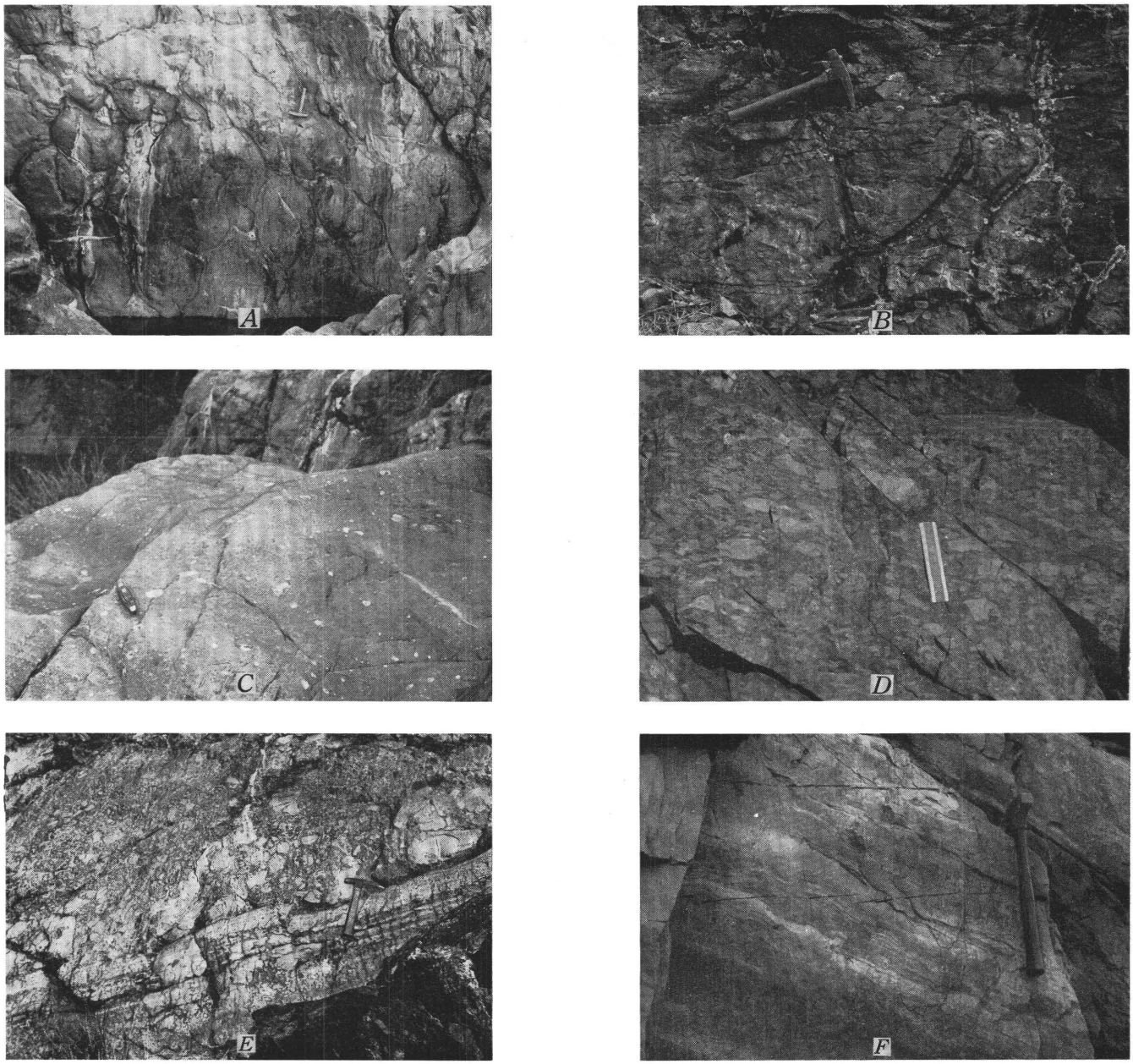
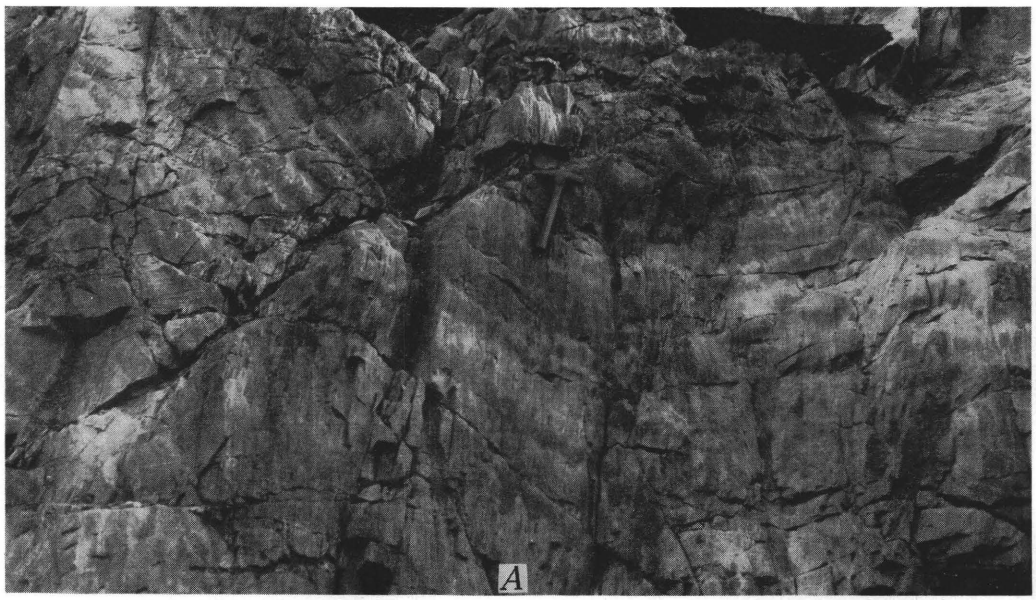


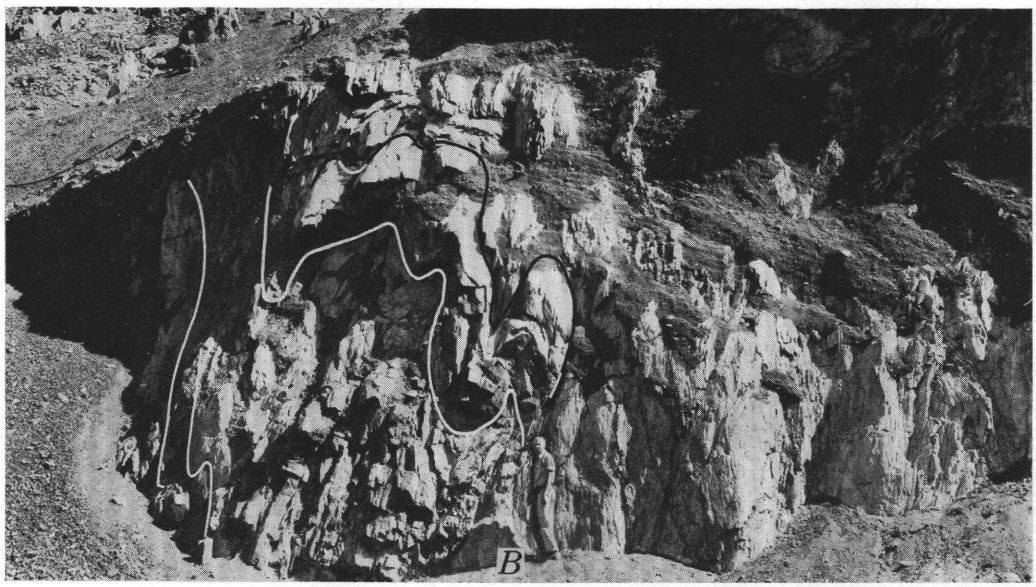
FIGURE 11.—Contour map of the contact between gabbro (above) and massive sulfide, quartz porphyry, and Grapevine Gulch formation (below), United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.



PHOTOGRAPHS OF ROCKS OF ASH CREEK GROUP

A. Pillow structure in Gaddes basalt. *B*. Pillow structures in Gaddes basalt showing fine-grained selvage on pillows (black). *C*. Quartz amygdules in Gaddes basalt. *D*. Rhyolitic breccia, Buzzard rhyolite. *E*. Coarse breccia resting on finely bedded tuffaceous rock, Deception rhyolite. *F*. Bedding exposed on foliation plane, Grapevine Gulch formation. Lineation dipping at low angle to right produced by intersection of bedding and foliation.





FOLDS IN GRAPEVINE GULCH FORMATION

A. Syncline in Grapevine Gulch formation showing foliation dipping steeply to the left (east) essentially parallel to the axial plane of the fold. Exposed in south wall of pit, United Verde mine. *B*. Minor folds in Grapevine Gulch formation plunging 40°–55° NNW. Exposed in south wall of pit, United Verde mine. Man at base of exposure gives scale.

ground minor structures reveal the major structure. In the absence of good marker beds, the syncline cannot be recognized at the surface owing to the presence of many minor folds.

The structure of the Grapevine Gulch formation west and north of the gabbro can only be conjectured because the formation is largely covered by dump derived from stripping for the open pit. The information for this area, shown on plate 5, is chiefly from maps prepared by L. E. Reber, Jr., formerly chief geologist for the United Verde mine. The rocks south of Haynes fault and west of the gabbro trend northward. The regional pattern indicates that the beds face west, and that probably the overturned beds or minor folds account for the local east dips. North of the Haynes fault, the beds trend northwestward, and probably face northeast, to form an anticline that may be the northwesterly extension of the one at the United Verde mine, or possibly another.

Foliation

Reference has been made earlier to the regional foliation, striking N. 20° W. and dipping 75°–80° E. Near the United Verde mine, foliation is well developed in the quartz porphyry and Grapevine Gulch formation, and is less pronounced but recognizable in the Deception rhyolite. Foliation appears to decrease in intensity to the east and is weak in the Grapevine Gulch formation north of the Haynes fault in contrast to the strong foliation in the same rocks south of the fault.

The intensity of this foliation is strongest in a northwestward-trending zone south of the open pit; the massive sulfide body lies at the north end of this zone, and the partly chloritized quartz porphyry coincides with it (pl. 5). The axis of the massive sulfide pipe plunges 65° N. along a bearing of N. 20° W. so that in depth, this eastward-dipping zone of intense foliation passes to the east of the center of the massive sulfide; and on the 2,250 level (pl. 7), the massive sulfide forms a south-southeastward-trending prong in the strongly foliated zone, indicating that the foliation was an important local control during mineralization. A minor control exerted by the foliation in zones of disseminated sulfide is revealed by the concentration of pyrite cubes in certain foliation planes. Where the disseminated pyrite can be observed on foliation planes, it forms a lineation whose northward plunge is about equal to that of the massive sulfide pipe.

A second and younger foliation, striking N. 15°–20° E. and dipping chiefly 75°–80° E., occurs along the footwall (south) side of the massive sulfide, and near the gabbro (pl. 7). Where both structures occur together, the northeastward-trending one cuts the northwest. Despite the southeast dip of the younger foliation, it does not extend downdip to lower levels; on each successive lower level, the zone of northeast foliation is close to the northwestward-plunging footwall of the massive sulfide pipe.

The origin of the northeast foliation is uncertain, but it may have some connection with the gabbro which is foliated weakly along and parallel to its east margin. A northeastward push exerted by the gabbro magma in the late stages of emplacement after partial crystallization may have produced the northeast foliation in the adjacent rocks, as well as bending the fold axes and foliation as observed on the surface.

Weak northwestward-trending foliation in some of the andesitic dikes parallels the stronger northwest foliation in the quartz porphyry host rock, which is probably caused by recurrent movement along these planes.

Faults

The United Verde mine is in a horst bounded by four normal faults: the Verde to the east, the Warrior to the west, the Hull to the south, and the Haynes to the north (pl. 5). The Hull and Verde faults are discussed in greater detail on page 70. All four faults displaced Paleozoic sedimentary rocks, and near the United Verde mine the Verde fault displaced the Tertiary Hickey formation.

The Haynes fault strikes east and dips steeply north, and extends across the northern part of the area to the Verde fault. The Warrior fault strikes north-northeast; and extends from the Haynes fault southwest for several miles. Both faults offset Paleozoic rocks; whether they were active during the Precambrian is uncertain, but the distribution of the Precambrian rocks on each side of the Hull fault indicates an earlier period of movement.

North of the Haynes fault, the divergent structural trend in the Grapevine Gulch formation from that to the south indicates greater displacement than the 200 feet of vertical separation shown by the Tapeats sandstone(?).

Many small faults trending westward or northwestward cut Paleozoic rocks in the northwestern part of the United Verde area (pl. 5), but none of these are known to have moved during the Precambrian.

SEQUENCE OF MINERALIZATION

Ore deposition at the United Verde mine was connected closely with special types of alteration, so that it is difficult to draw a clear-cut distinction between alteration and deposition of sulfide minerals. The evidence is convincing that in and near the United Verde mine, tuffaceous rocks of the Grapevine Gulch formation were silicified to masses of fine-grained quartz (fig. 12, *2*). This early silicification was followed by

the replacement of Grapevine Gulch formation and quartz porphyry by massive pyrite forming a pipelike body (fig. 12, *3*). Carbonate minerals and quartz are part of the pyritic pipe; the carbonate minerals, as well as the pyrite, clearly represent partial alteration of the host rocks. The quartz may be residual in part from

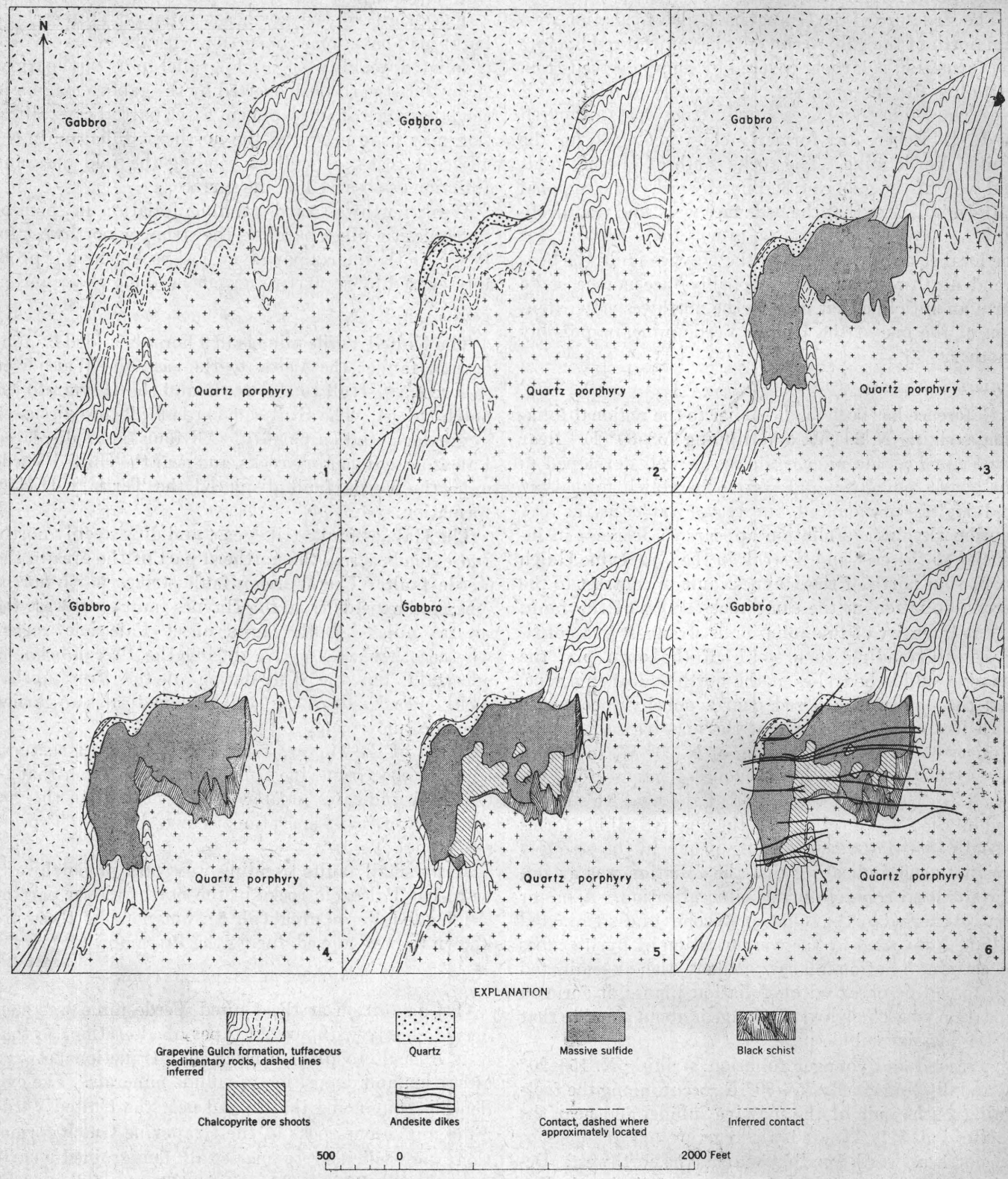


FIGURE 12.—Series of sketch maps to show sequence of mineralization, United Verde mine.

the host rock and redistributed during replacement. Extensive chloritization of the rocks on the footwall (south) side of the massive sulfide pipe was next in the sequence, overlapping somewhat with the main period of pyritic deposition (fig. 12, *4*). Quartz porphyry was the dominant rock chloritized to form black schist, but part of Grapevine Gulch formation was likewise chloritized. The main period of copper deposition followed the extensive chloritization (fig. 12, *5*). Quartz-carbonate veins and nodules were the final phase of mineralizing activity, containing some sulfide minerals. The andesitic dikes were intruded before the end of the quartz-carbonate veining and after the main period of copper deposition (fig. 12, *6*).

The separation of the products of dynamic metamorphism and widespread hydrothermal alteration is not easy, for each process resulted in the formation of sericite from the feldspar in the silicic rocks. Near Jerome the alkali content drops appreciably in the Deception rhyolite and the K_2O and MgO content increases in the quartz porphyry. To the south, the Na_2O content in the quartz porphyry is higher. These chemical data seem to indicate widespread hydrothermal activity, which near Jerome is revealed in the quartz porphyry and Deception rhyolite by a simple mineralogy of quartz and sericite, locally modified by clots and streaks of chlorite.

The writers believe that from south of the United Verde mine to Mescal Gulch, the quartz porphyry and Deception rhyolite have been hydrothermally altered after the regional metamorphism. In places the foliation is destroyed or so weakened, that it can be recognized only in favorable exposures. Near the United Verde mine, the Grapevine Gulch formation locally shows the same type of alteration.

Eastward, beyond the Verde fault, alteration is recognizable in the underground workings of the United Verde Extension mine near the ore body, but near the Audrey shaft (pl. 10) bright feldspar crystals indicate an absence of sericitic alteration.

The chloritization and copper mineralization in the United Verde ore body are clearly younger than the sericite in the sericitized and silicified quartz porphyry, but we are uncertain as to the relation between (1) the local early silicification and main period of pyritic mineralization and (2) the widespread sericitization.

SILICIFICATION

The process of silicification resulted in the formation of large and small masses of nearly pure quartz, limited largely to the hanging-wall side (northwest) of the massive sulfide pipe (pl. 7). The quartz masses in places form a screen between the massive sulfide and overhanging gabbro. Information on the distribution of the quartz has been obtained chiefly from diamond-drill holes, for it is only in the upper levels of the mine that drifts and crosscuts regularly reach into the hanging-wall side of the massive sulfide pipe. Some of these workings in the upper levels were accessible at the time of our study.

The quartz is fine grained, almost flinty, and is red, black, or white. Color variations are not sharp. The quartz bodies formed partly by replacement of tuffaceous sedimentary rocks of the Grapevine Gulch formation, and in favorable exposures, relict bedding and minor folds are sufficiently preserved to prove the replacement origin of the quartz. In other exposures, the quartz bodies grade into quartz porphyry in such a manner to suggest that the igneous rock has been replaced. Although the data are incomplete it appears that the tuffaceous rocks were more commonly replaced by quartz than the quartz porphyry.

The form of the quartz bodies is lenticular to irregular. In horizontal outline, some of the quartz bodies are 500 feet long and as much as 150 feet wide; others are only 10 feet wide but 200 feet long. Contacts between quartz and massive sulfide are irregular, but between quartz and gabbro, they are even. The large quartz masses formed adjacent to the gabbro in local structural terraces in the gabbro contact (fig. 13).

Quartz bodies occur well within the main body of massive sulfide; and local veins, nodules, and streaks of quartz are common. Some of the quartz veins appear to cut the massive sulfide, but many of the nodules and streaks and larger bodies of quartz are cut by veins of pyrite that project into the quartz from the massive sulfide. The larger masses of quartz on the hanging wall of the massive sulfide are also veined by fine-grained pyrite; and locally, brecciated quartz was cemented by pyrite. The quartz therefore was deposited before the pyrite that forms the massive sulfide pipe.

In the footwall (southeast) side of the massive sulfide, particularly on the lower levels, jasper nodules are common in the quartz porphyry. These nodules range from 1 inch to 2 feet in length, and range in shape from rounded to elliptical or irregular. Some truncate foliation in the quartz porphyry whereas others parallel it, or the foliation wraps around the nodules. The interior of some contains pyrite in disseminated grains and in aggregates. The jasper nodules may be either contemporaneous with the quartz masses in the hanging-wall side of the massive sulfide, or they may be older, or younger. The nodules definitely formed before the end of mineralization, as a few are cut by quartz-carbonate veins.

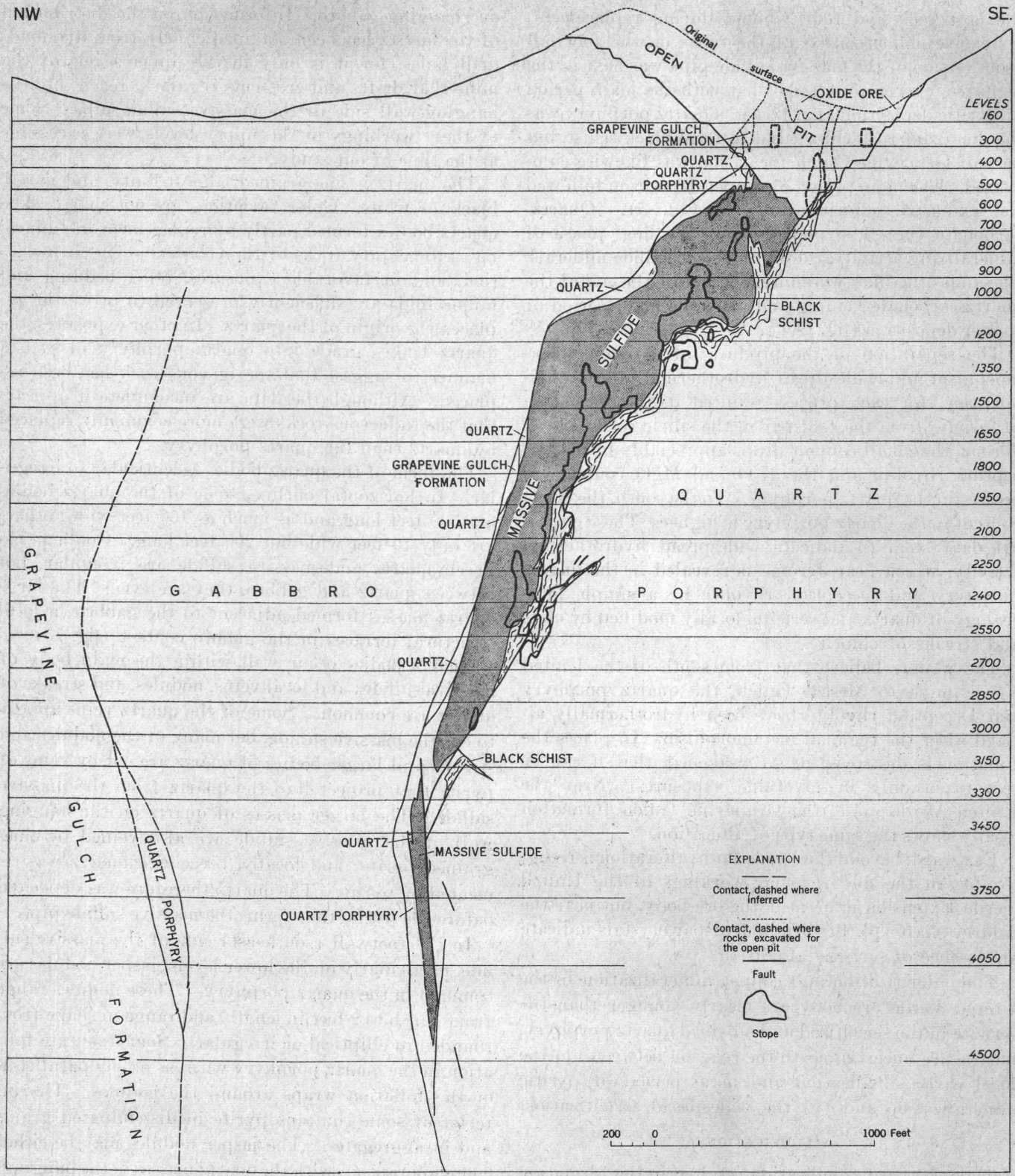


FIGURE 13.—Vertical section through United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.

MASSIVE SULFIDE PIPE

SIZE AND SHAPE

The large pipelike body of massive sulfide formed during the second period of mineralization by replacement of quartz porphyry and Grapevine Gulch formation by pyrite, quartz, and carbonate minerals. Sphalerite and chalcopyrite are present in minor amounts, and locally, sphalerite in significant quantity. The bearing of the axis of the massive sulfide pipe is about N. 20° W. and the plunge is about 65° N.

The variations in area of massive sulfide for each level are shown graphically by figure 14. From the surface to the 900 level, the horizontal area ranges from 220,000 to 260,000 square feet. Below the 900 level, it increases to about 500,000 square feet at the 1,500 level. Below this level, the horizontal area diminishes regularly to the 3,300 level, where it is about 60,000 square feet. Below the 3,300 level the horizontal area decreases at a smaller rate per level; and for all practical purposes, the massive sulfide bottoms a short distance below the 4,500 level.

The horizontal outline of the massive sulfide pipe varies on each level (fig. 15), and the variations are controlled by the gabbro contact on the hanging-wall (northwest) side of the massive sulfide and by a zone of intense foliation that controlled in part the eastern limb. The gabbro contact to the 3,000 level is essentially an inverted trough, plunging 65° N. and has a bearing of N. 20° W. The zone of intense foliation strikes N. 20° W. and dips 75°–80° E. Where these two structural elements converge on the 900 level and higher levels, the sulfide pipe is nearly equidimensional. At the 1,200 level the eastward-dipping zone of intense foliation diverges from the axis of inverted trough in the gabbro and the eastern part of the massive sulfide pipe is elongated in a south-southeast direction along the zone of intense foliation. This eastern limb dips east, parallel to the foliation (fig. 16) and away from the western part of the sulfide pipe adjacent to the gabbro. The western part of this pipe is also elongated in a south-southeast direction because of this trend in the gabbro contact. Along the west margin of the massive sulfide at successively lower levels, the gabbro contact trends more and more southward; on the 2,550 level, it trends southwestward and the western part of the massive sulfide has a similar trend. Because the eastern limb follows the zone of intense foliation, at this level the massive sulfide is crescent shaped horizontally. At greater depth, the two limbs diverge farther, and the massive sulfide narrows in the axial region of the sulfide pipe. On the 3,000 level, it is divided near the axis into two segments, and below the 3,000 level each limb of the arcuate pipe further divides into several segments. On the 3,450 level, the gabbro contact has a straight northeast trend, and the arcuate form of the massive sulfide has disappeared. Instead two small bodies of massive sulfide next to the gabbro are all that appear on the 3,450 level. Below this level, they break into segments and decrease in size so that on the 4,500 level, massive sulfide occurs only in 6 rootlike masses, the largest of which is 15 by 130 feet.

The gabbro, perhaps because it was more massive than the adjacent rocks, resisted mineralization and in the middle and upper levels channeled the solutions below the inverted trough into the adjacent quartz porphyry and Grapevine Gulch formation. Most of the massive sulfide formed in the zone above the 3,000 level where the gabbro dips west, and the inverted trough is present. Below the 3,000 level, the gabbro contact dips east, and massive sulfide abruptly diminishes. Thus on the intermediate levels where the western limb of the crescent body of massive sulfide is linear, the gabbro contact is the structural element to which the massive sulfide conforms.

In summary, the shape of the massive sulfide pipe is chiefly the result of three controls: (1) the overhang of the gabbro above the 3,000 level, (2) the contact between gabbro and the Grapevine Gulch formation or quartz porphyry, and (3) the zone of pronounced foliation. The shape on any given level is a function of the spatial relationship of these structural elements to each other.

RELATIONSHIP TO ADJACENT ROCKS

The contact of the massive sulfide with the adjacent rocks is generally sharp, but in many exposures it is gradational with disseminated pyritic rock. The massive sulfide is in contact with tuffaceous rocks of the Grapevine Gulch formation, quartz porphyry, early quartz masses, and black schist. Most of the black schist probably formed later than the pyritic massive sulfide.

The massive sulfide-early quartz contacts are best exposed in the upper levels of the mine on the hanging-wall side of the sulfide pipe; the relationship is discussed above.

The contact between the pyritic massive sulfide and tuffaceous rocks is best exposed also on the hanging-wall side of the sulfide pipe on the upper levels. Disseminated pyrite is present locally in some of the tuffaceous rocks. Lenses and tongues of massive sulfide as much as a foot wide are in the tuffaceous rocks, parallel to the stratification. In one locality, minor folds in the tuffaceous rocks are duplicated in the adjacent massive sulfide, clearly indicating replacement; at many others, banding in the margin of the massive sulfide parallels the stratification in the adjacent tuffaceous rocks, and in a few places, banding well within the

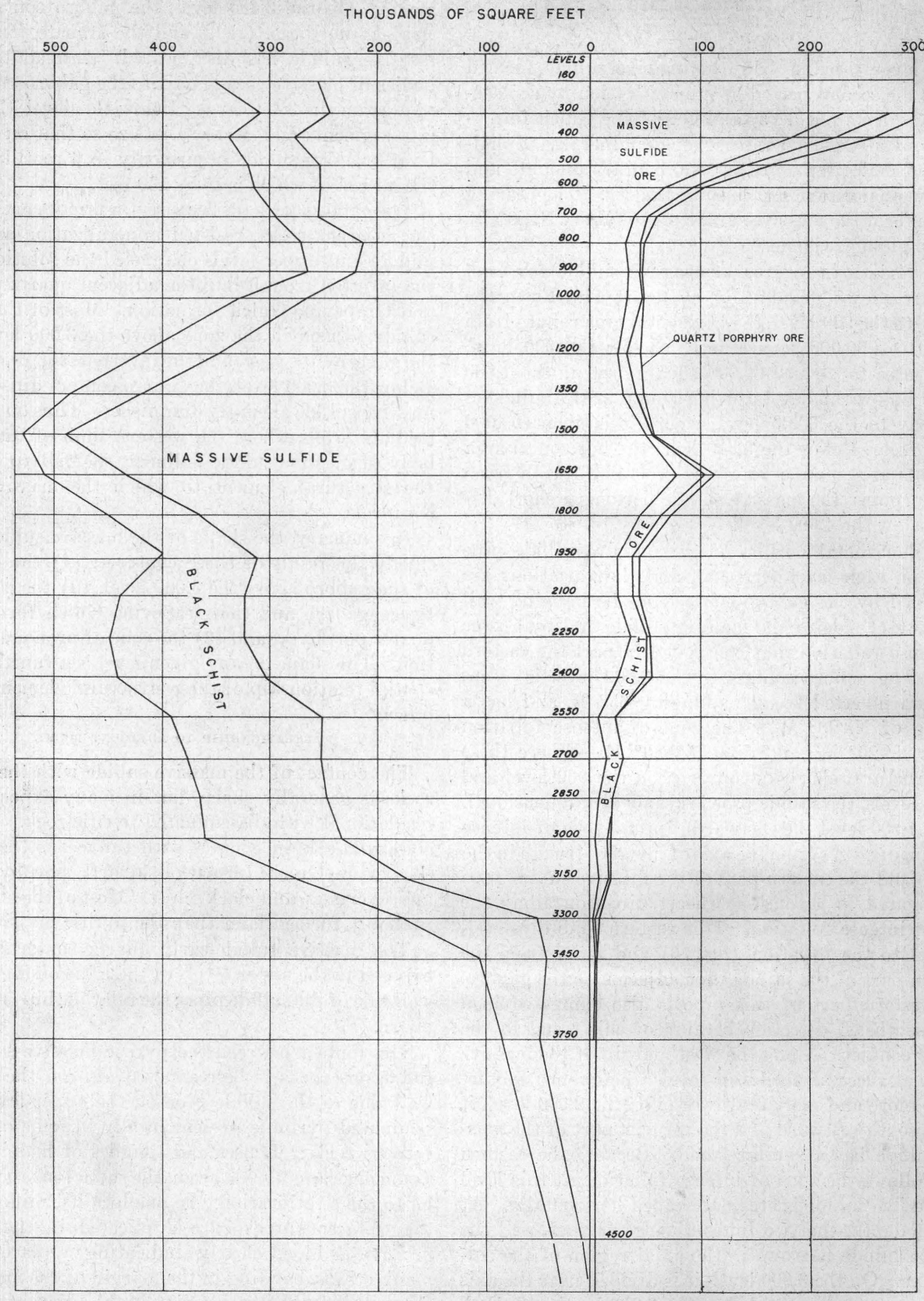


FIGURE 14.—Graphs showing relations of massive sulfide to black schist and types of ore in main ore body, United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.

massive sulfide parallels the bedding in the nearest (100 feet) exposure of tuffaceous rocks. This banding is interpreted as relict bedding following replacement of the sedimentary rocks by the pyritic massive sulfide.

Quartz porphyry generally is foliated where it is in contact with the massive sulfide. In many exposures, the massive sulfide contact is conformable with the adjacent foliation, but in many places, it cuts across the foliated quartz porphyry. Tongues of massive sulfide in part parallel the foliation, but some tongues crosscut it. On the 3,150 level, the quartz porphyry is brecciated, locally, at the contact with the massive sulfide, and pyrite is present between the quartz porphyry fragments. In a few exposures, the massive sulfide grades into pyritic disseminations in the quartz porphyry.

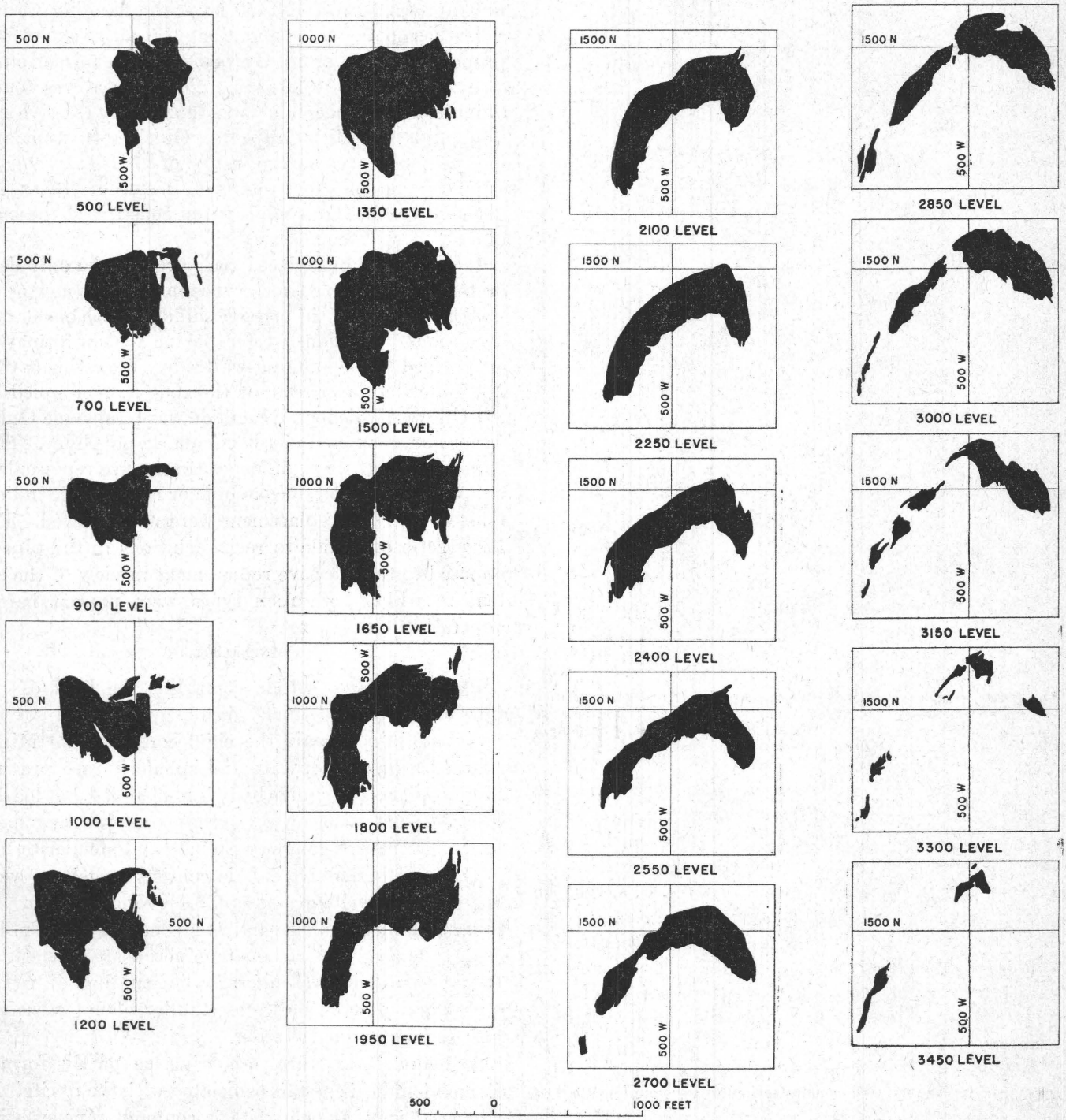


FIGURE 15.—Sketch map showing outline of massive sulfide and associated quartz on various levels, United Verde mine.

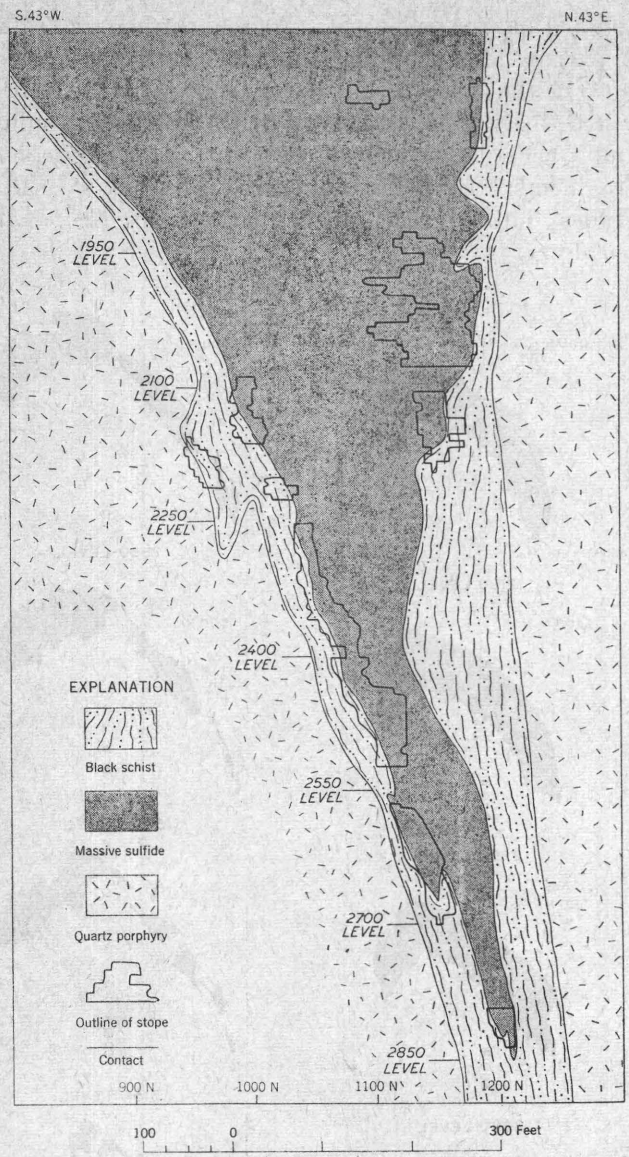


FIGURE 16.—Vertical section looking northward across eastern limb of massive sulfide, United Verde mine. Shows eastern dip of massive sulfide controlled by intense foliation in host rock, and distribution of ore shoots along margin of massive sulfide.

The contact between black schist and pyritic massive sulfide is variable. Generally, the massive sulfide essentially conforms with the foliation in the black schist, and where the dip of the contact changes, the foliation also changes in dip. In many exposures, however, the pyritic massive sulfide truncates the foliation of the black schist. Some contacts are sharp and others intertongue with pods of pyrite parallel to or discordant with the foliation. Black schist stringers as much as 1 inch wide are preserved within the massive sulfide adjacent to the quartz porphyry.

How much of the massive sulfide zone was originally quartz porphyry and how much was Grapevine Gulch formation? The distribution of these rocks within the massive sulfide pipe and adjacent to it provides the only evidence. Quartz porphyry forms the footwall, except for a little of the Grapevine Gulch formation along the southwestern part of the sulfide pipe on the 1,650 level (pl. 7). On the hanging-wall (north) side of this pipe, some approximations can be made. From the 600 to 2,100 level, the north side is fairly well defined; and gabbro, Grapevine Gulch formation, and quartz porphyry are in contact with it in the proportion, 5:3:1. From the 2,250 to the 3,150 level, the hanging wall is a gentle smooth curve; and along the entire extent, the proportion of gabbro to Grapevine Gulch formation to quartz porphyry, is 1.3:4:1. No evidence was found to indicate that the gabbro was replaced by massive sulfide; and the relict Grapevine Gulch formation and quartz porphyry in the ratio of 3:1–4:1 suggest strongly that on the hanging-wall side of the sulfide pipe, Grapevine Gulch formation constituted the bulk of the rock that was replaced.

Inclusions of unreplaced rock provide the only data on the distribution of rock types in the zone now occupied by the core of the massive sulfide. With one or two exceptions, inclusions of Grapevine Gulch formation are limited to the hanging-wall side. We estimate that not more than a quarter of the zone now occupied by the United Verde massive sulfide was Grapevine Gulch formation, the remainder being quartz porphyry. This conclusion is of no significance if selective replacement occurred. However, megascopic or microscopic indications of selective replacement were not observed. The large ratio of sulfide to rock inclusions in the pipe is opposed to the selective replacement in view of the extent to which both rock types were present before mineralization.

MINERALOGY

Pyritic massive sulfide comprises the bulk of the pipelike body, of which pyrite, quartz, and several carbonate minerals are the chief constituents. Minor quantities of chalcopyrite and sphalerite are present. In most places the sulfide body is nonbanded, but locally it has banded facies owing to the parallel arrangement of the pyrite and gangue or pyrite and sphalerite.

The pyritic massive sulfide consists largely of very fine grained pyrite grains (pl. 8*A*), ranging from 0.04 to 0.2 millimeter in size and averaging about 0.1 millimeter. Quartz and carbonate minerals are interstitial. In some specimens a microscopic banding of pyrite grains and gangue indicates a relict planar structure such as bedding or foliation.

Arsenopyrite is rare, occurring as partly formed diamond-shaped crystals associated with the pyrite. In a general way, arsenopyrite is contemporaneous with

the pyrite, although individual grains of arsenopyrite may be slightly younger than contiguous pyrite.

The ratio of quartz gangue to the carbonate gangue varies, and in some specimens, the gangue is largely quartz whereas in others it is carbonate minerals. The quartz forms grains about the same size as the pyrite grains, and in part is in single irregular grains between pyrite or carbonate, but locally the quartz forms aggregates. In a few specimens, particularly where the pyrite has been fractured and healed by later quartz, the quartz consists of finely granular aggregates or finely columnar or fibrous crystals. Some of the columnar crystals are normal to faces of single pyritic crystals, but some are subparallel for short distances and show no relationship to individual pyritic crystals. The quartz may be residual from the replaced rock, except for veinlets that cut the pyrite, but the vein quartz may be redistributed original quartz.

The carbonate minerals form single crystals and aggregates, either pure or intergrown with quartz. The composition of carbonate minerals range from dolomite to ankerite, according to the range of the omega index of refraction, from 1.68 to 1.72.

Specimens of the pyritic massive sulfide taken from near the black schist contact on the footwall of the sulfide pipe contain streaks and nests of chlorite, associated with quartz and dolomite. The chlorite crystals are larger than those in typical black schist and in part are in subparallel arrangement but commonly are in random orientation and individual chlorite crystals penetrate pyrite. This relationship gives no clue to relative age: the pyrite and chlorite may be contemporaneous; earlier pyrite may be partly replaced by chlorite, or older chlorite may have been inert to the pyritic solutions.

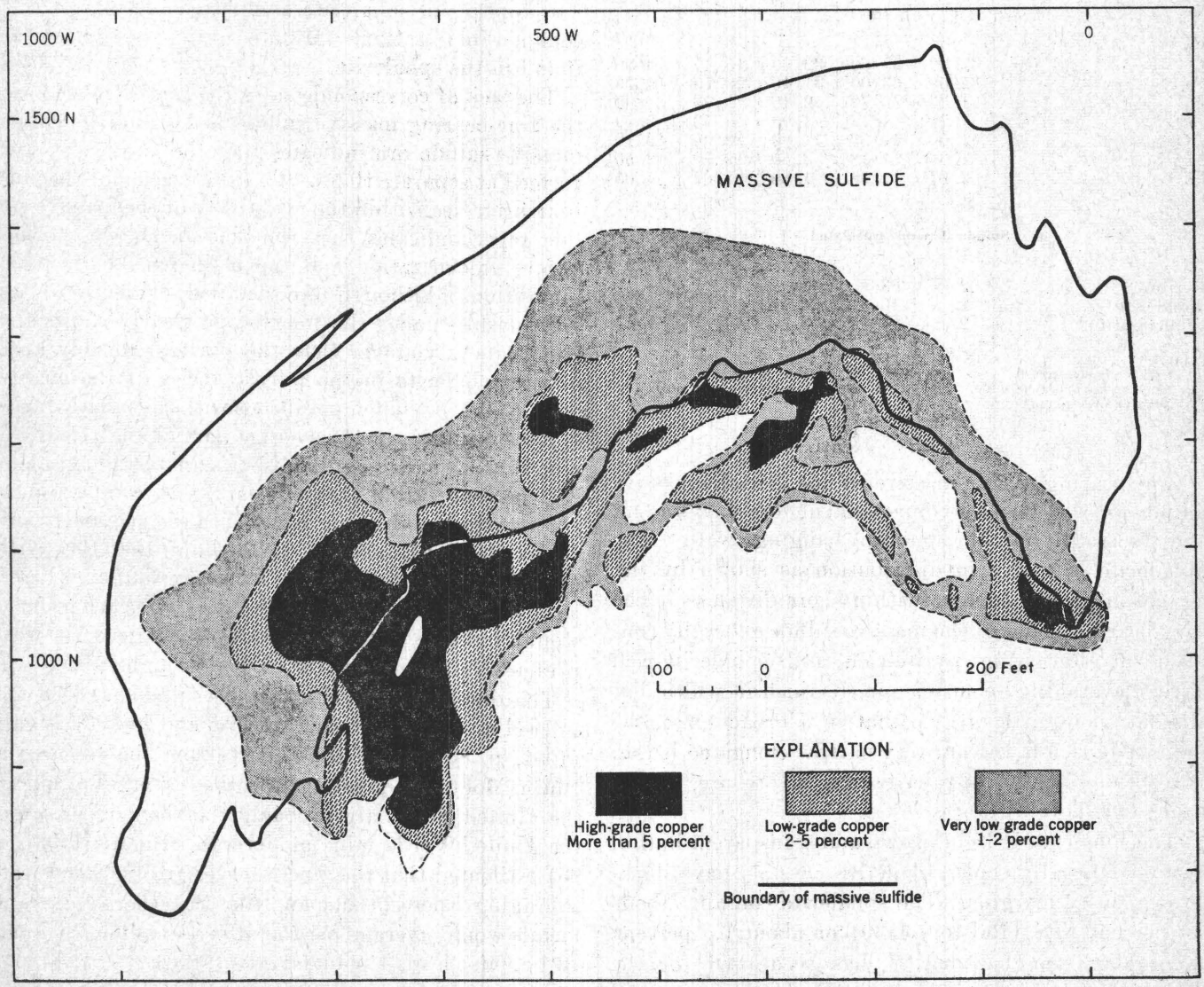


FIGURE 17.—Sketch map showing distribution of copper on a typical level, United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.

Most polished surfaces of the pyritic massive sulfide show small amounts of sphalerite and chalcopyrite interstitial to the pyrite. Possibly some copper and zinc were deposited with the pyritic massive sulfide, but the distribution of copper (fig. 17) is spatially related to the footwall side of the pipe, and the copper content decreases toward the hanging-wall (north) side, indicating that the chalcopyrite in the pyritic facies may be younger than the formation of the pyritic massive sulfide pipe.

Reber (1938, p. 48) has given an average partial analysis of the pyritic massive sulfide (table 21, analysis 1), which shows that sulfide minerals average only 66 percent, the remainder consists of quartz, dolomite and ankerite, and locally, chlorite.

TABLE 21.—*Average partial analyses of United Verde massive sulfide and types of copper ores, in percent, from Reber (1938)*

	1	2	3	4	5
S	37.00	38.00	32.60	15.40	10.80
Fe	35.00	33.00	31.40	20.70	15.70
Zn	1.33	6.72	2.40	.70	.90
Cu	.70	1.16	4.99	4.73	2.79
SiO_2	11.00	11.50	11.80	22.90	40.10
Al_2O_3	1.39	1.18	3.50	12.70	12.50
CaO	2.07	.78	1.80	1.22	1.05
Sulfide minerals calculated					
Pyrite	62	63	50	19	14
Sphalerite	2	10	4	1	1.3
Chalcopyrite	2	3	14.5	14	8

1. Pyritic massive sulfide.
2. Zinc-bearing massive sulfide.
3. Massive sulfide copper ore.
4. Black schist copper ore.
5. Quartz porphyry copper ore.

ZINC-BEARING FACIES

Zinc-bearing pyritic massive sulfide is in the massive sulfide pipe between the copper-bearing ore shoots along the footwall and the gabbro hanging wall. The sphalerite is spotty in distribution as shown by the variation in zinc content within short distances. The zinc-bearing facies of the massive sulfide generally consists of alternating pyrite-rich and sphalerite-rich layers, averaging about one quarter inch in width, but discontinuous in length so that in a broad sense, the sphalerite-rich layers are very narrow elongated lenses. Exceptionally, the sphalerite-rich layers are 1 inch wide for short distances.

The sphalerite is light brown and transparent in thin section. Locally some sphalerite crystals have dark-brown dusty margins. The sphalerite contains about 10 percent FeS (Ralston, 1930) or about 6.5 percent Fe. The term "marmatite" has been used for the United Verde zinc sulfide, but Palache, Berman, and Frondell (1944, p. 2122) have defined marmatite as containing more than 10 percent of Fe.

Microscopic studies of polished surfaces show that generally the pyrite is older than the sphalerite, for sphalerite veinlets cut fractured pyrite. On some polished surfaces, the fine-grained pyrite crystals are preferentially oriented, arranged in chainlike trains, and separated by quartz-carbonate gangue, also arranged parallel to the pyrite. Presumably this preferred orientation of pyrite crystals is a relict structure after foliation or bedding. The sphalerite veinlets truncate this older structure at several angles. In places, the sphalerite veinlets contain a gangue of quartz and carbonate. In other specimens, sphalerite is interstitial to pyrite, indicating that the original gangue minerals were replaced.

Chalcopyrite generally forms irregular grains surrounded by sphalerite, but in a few places the chalcopyrite appears in sphalerite as microscopic blebs with a common orientation. Minute grains of galena are found in the sphalerite.

The lack of correspondence in the spatial relations of the zinc-bearing massive sulfide facies and the copper massive sulfide ore indicates that the two facies were formed at separate times. The localization of the zinc-bearing facies within the main body of the massive sulfide pipe, indicates that the zinc facies was formed before chloritization and the main period of copper deposition. Although the detailed evidence of age relationship under the microscope clearly shows that sphalerite is younger than the pyrite, probably much of the sphalerite in the pyritic facies of the massive sulfide must be contemporaneous with the pyritic facies.

Reber's (1938, p. 48) average partial analysis of the zinc-bearing massive sulfide shows an average grade of 6.72 percent (table 21, analysis 2). A comparison of the average zinc-bearing facies with the massive pyritic facies (table 21, analyses 1, 2) shows that the pyrite and silica content of the two facies are similar, whereas the lower content of CaO in the zinc-bearing facies would indicate that the sphalerite, in part at least, preferentially replaces the carbonate minerals.

The zinc-bearing pyritic massive sulfide has been considered an important zinc reserve, and in 1931 Ingalls (1931, p. 83) stated that apart from the reserves at Butte, Mont., the greatest unworked reserves of zinc in the United States are probably the zinc-copper ore of the United Verde mine at Jerome. Ingalls (1931, p. 63) estimated that the zinc reserves are 12,000,000 tons averaging 7 percent zinc and that the other recoverable metals would average per ton of ore as follows: copper, 12 pounds; lead, 4 pounds; silver, 2 ounces; gold, 0.02 ounce.

In 1930, serious attention was given to the recovery of the zinc from the massive sulfide (Ralston, 1930a) but the small size of the pyrite, sphalerite, and chalcopyrite grains requires extremely fine grinding for separation; on the average the ore must be ground to pass 800-mesh to obtain more than 90 percent liberation of copper and zinc minerals from the pyrite. As 300-mesh is about the commercial limit for grinding, the grains in part consist of aggregates of sphalerite and chalcopyrite. Tests made by Ralston produced zinc concentrates containing 45 percent zinc and 2 percent copper. The copper concentrate from the zinc-bearing massive sulfide contained about 15 percent copper and from 8 to 10 percent zinc.

According to Yates [6] the development of the zinc-bearing massive sulfide disclosed that it was irregular and discontinuous and considerable zinc-bearing massive sulfide was removed in the open-pit operation. The remainder of the best grade of zinc in the massive sulfide is between the gabbro hanging wall and filled stopes and along the footwall side. Much of the zinc-bearing massive sulfide is now slowly subsiding owing to compaction of the fill in the stopes, and any plans for future mining of the zinc must take into account this mine subsidence.

Probably the estimate of grade and tonnage of the zinc by Ingall is optimistic, but no accurate figure can be determined because of the scarcity of data to block out irregular and discontinuous zinc ore shoots. But it is true that the reserves of zinc-bearing massive sulfide are large. When it is economic to mine 4–5 percent zinc in massive sulfide, considering the fine intergrowth of zinc, copper, and pyrite, and the mine subsidence, the zinc-bearing massive sulfide in the United Verde mine will receive serious consideration.

BLACK SCHIST

ORIGIN

Large masses of nearly pure chloritic rock are on the footwall and locally east of the massive sulfide pipe. The chloritic rock is greenish black to black and has been called "black schist" by the mine operators. The black schist largely formed from quartz porphyry, but on the upper levels on the east side of the massive sulfide pipe, tuffaceous rocks north, east, and south of black schist indicate that considerable Grapevine Gulch formation was replaced by chlorite.

In the large masses of black schist where all relict minerals and textures are gone, it is impossible to determine the original rock with certainty, but the unreplaced rock at the margins provides some information. On the fringes of the masses of the black schist, unreplaced rock is cut by fractures containing chlorite, and at fracture intersections, small masses of black schist are present. Where quartz porphyry is the replaced rock, relict quartz phenocrysts in a base of chlorite are present several inches from the unreplaced quartz porphyry. Thin sections reveal that the quartz phenocrysts have ragged edges and are partly veined by chlorite. Near the margins of the black schist, the groundmass of the quartz porphyry consists of quartz granules and chlorite indicating that sericite and feldspar are the first minerals in the porphyry to be replaced by chlorite.

STRUCTURE AND MINERALOGY

The black schist, generally, is poorly foliated and breaks with irregular surfaces parallel to the foliation. Some blocks are massive and others splintery. Microscopic studies show that the black schist consists of a very fine grained aggregate of pale-green chlorite, in places having a preferred orientation and in other places, forming a mat. The diameter of the individual crystal ranges from 0.002 to 0.01 millimeter. Locally, chlorite crystals from 0.02 to 0.5 millimeter in diameter form aggregates, commonly divergent from the foliation.

Zircon crystals, ranging from 0.001 to 0.002 millimeter in length, are commonly surrounded by pleochroic haloes as large as 0.006 millimeter across. A few larger zircon crystals, 0.07 millimeter long, have pleochroic haloes 0.01 millimeter across. Rutile crystals are common in the black schist, 0.01 millimeter long, and arranged in clusters or in trains of disconnected granules. Blue tourmaline was observed in one thin section. Ankerite in individual crystals and in aggregates is in some thin sections. Quartz grains are rare in the solid masses of black schist except as younger crosscutting veinlets.

CHEMICAL COMPOSITION

The high chlorite content of copper ore associated with black schist added to smelter difficulties which necessitated many complete and partial chemical analyses by the United Verde smelter analysts. These chemical data have been made available for our study (table 22). Three composite samples of black schist were carefully collected, two from the 2,400 level and one from the 3,000 level of the mine. A sample of quartz porphyry was collected on the 2,100 level, 1,600 feet south of the massive sulfide pipe. Using bulk specific gravity determinations, it is possible to calculate the amount of milligrams per cubic centimeter of the main analyzed constituents, and these are plotted graphically (fig. 18) to show the appreciable loss in SiO_2 and CaO and marked gain in Al_2O_3, FeO, MgO, and H_2O in the conversion of quartz porphyry to black schist. This graph

[6] Yates, P. F., 1946, Bottoming of the United Verde sulphide pipe: Mimeographed paper distributed at meeting of Arizona section, Am. Inst. Min. Metall. Eng., Tucson, Ariz., October 1946.

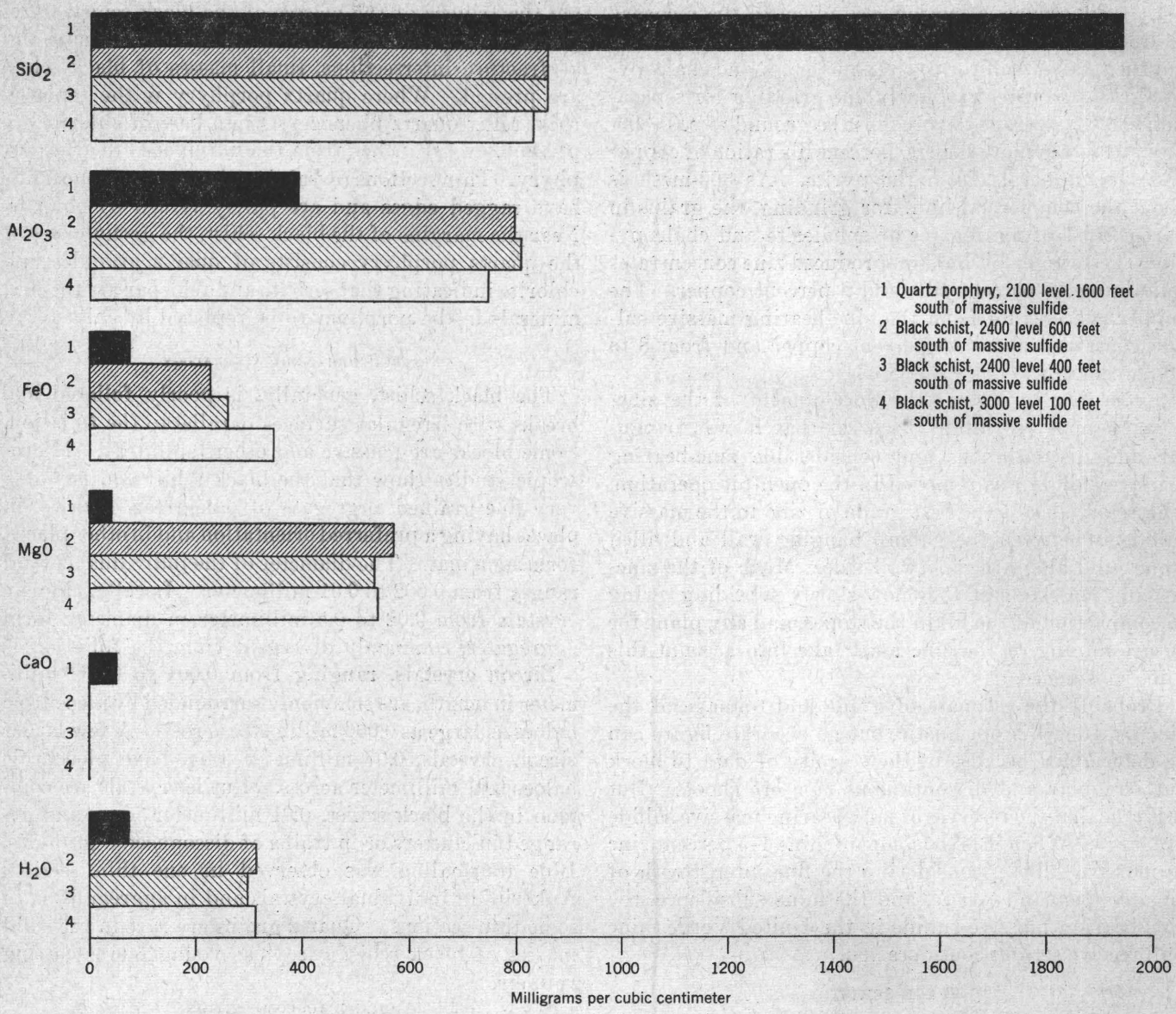


FIGURE 18.—Graph showing losses and gains of principal constituents in the formation of black schist from quartz porphyry.

also indicates that the iron content in the black schist increases toward the massive sulfide.

There is some variation in the ratio of SiO_2, Al_2O_3, FeO, and MgO in the black schist, and these variations are shown graphically in figure 19 by means of Niggli's *al* and *si* values, and *mg* ratios (Grubenmann and Niggli, 1924, p. 29). These values are obtained by calculating the molecular ratios of SiO_2, Al_2O_3, FeO, and MgO from the weight percentage of the chemical analyses. The molecular ratios of Al_2O_3, FeO, and MgO are recalculated on the basis of 100 percent so that an *al* value of 26 indicates that the Al_2O_3 content is 26 percent of the total Al_2O_3, FeO, and MgO molecules. The *si* value is the percent of the molecular ratio of SiO_2 to the total Al_2O_3, FeO, and MgO molecular ratios. The *mg* value is the molecular ratio of MgO to total FeO and MgO, which means that for *mg* in excess of 0.5, MgO is in excess of FeO.

The *al* and *mg* values are plotted against the *si* value (fig. 19) and they show, in general, a decreased *al* value with decreased *si* value, and in most of the analyzed black schist from the United Verde mine, *mg* is in excess of 0.5, indicating that MgO is in excess of FeO.

CHLORITE MINERALS

The variations in the chemical composition of the black schist are due in part to the variety of chlorite forming the rock, and in part, to variations in amounts of constituents such as quartz and carbonate. The differences in composition of the chlorite are most significant, because small amounts of accessory minerals are normally present. As discussed in the section on

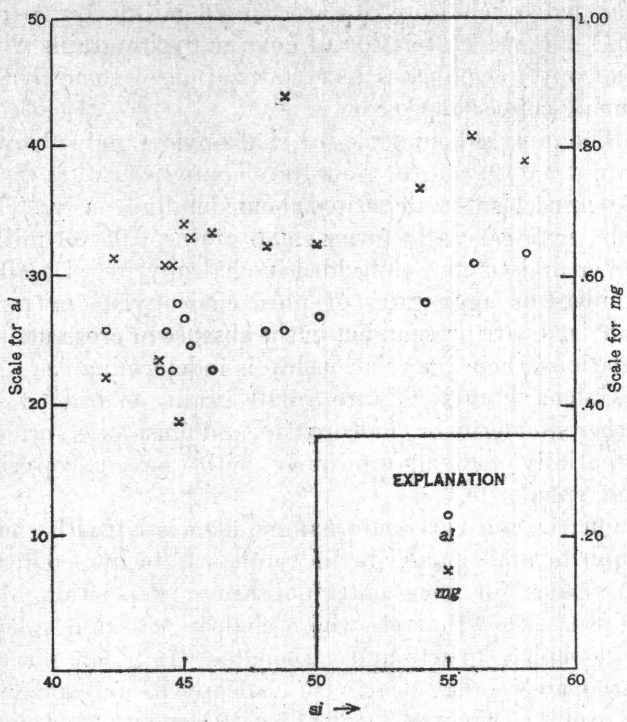


FIGURE 19.—Diagram showing Niggli values for analyzed black schist, United Verde mine.

TABLE 22.—*Chemical analyses of quartz porphyry and black schist, United Verde mine, in percent*

[Analyses furnished by Phelps Dodge Corp.]

	1	2	3	4
SiO_2	72. 08	30. 88	30. 92	29. 71
Al_2O_3	14. 57	28. 77	29. 30	26. 66
FeO[1]	2. 86	8. 42	9. 44	12. 45
MgO	1. 48	20. 64	19. 36	19. 42
CaO	1. 98	Tr.	. 04	. 03
H_2O	2. 67	11. 44	10. 06	11. 28
CO_2	1. 35			
S	. 06	. 02	. 01	. 55
Cu	. 02	Tr.	Tr.	. 04
Zn	. 04	. 14	. 12	. 27
Total	97. 11	100. 31	99. 25	100. 41
Specific gravity	[2] 2. 70	2. 78	2. 78	2. 82

[1] Iron determined as Fe, calculated as FeO.
[2] Average of 4 samples of quartz porphyry collected in the mine.

1. Quartz porphyry, 2,100 level, 1,600 feet south of massive sulfide.
2. Black schist, 2,400 level, 600 feet south of massive sulfide.
3. Black schist, 2,400 level, 400 feet south of massive sulfide.
4. Black schist, 3,000 level, 100 feet south of massive sulfide.

mineralogy, p. 93, the varieties of chlorite that have been recognized in the black schist in the United Verde mine are: clinochlore, prochlorite, diabantite, ripidolite, and brunsvigite. Most of the black schist is composed of clinochlore-prochlorite, but even with these varieties, the ratio of MgO to FeO varies appreciably.

Where veinlets of quartz, carbonate, or sulfide, or several of these minerals, cut the black schist, chlorite flakes adjacent to the veinlets are from 10 to 20 times larger than the fine-grained chlorite in the black schist. The optical properties of the coarser crystals are identical with those of the adjacent chlorite, indicating similar composition and mineralogy. Along some of the quartz-carbonate veins, pale-green chlorite crystals 1 millimeter or more in diameter are arranged normal to the vein walls. These flakes are of the same mineralogical variety as the adjacent host, and in most places are clinochlore or prochlorite.

RELATIONSHIP TO MASSIVE SULFIDE

The age relationship of the black schist to the massive sulfide presents a problem that cannot be answered easily. Reber (1922, p. 16) in his first report on the Jerome district, suggested that the black schist was present before the pyritic sulfide was deposited, but later (Reber, 1938, p. 45), he stated that the period of chlorite deposition marked a definite break in sulfide deposition, and that chlorite was deposited after the massive sulfide pipe. Lindgren (1926, p. 33) recognized that the chlorite preceded the introduction of chalcopyrite, but he concluded that the chlorite accompanied the sulfide in many places.

That the pyritic massive sulfide pipe formed before the black schist is shown by the unreplaced quartz porphyry and tuffaceous rocks (see 600 level, pl. 7) within the massive sulfide body and between the north side of the massive sulfide pipe and the gabbro. These unreplaced rocks within the massive sulfide are more common in the upper levels where the pipe is nearly circular.

Limitation of black schist to the footwall side of the massive sulfide pipe indicates that the solutions that added FeO, MgO, and Al_2O_3 and removed SiO_2 were restricted to the footwall. The distribution of the black schist clearly reveals replacement of quartz porphyry and tuffaceous rocks along favorable structures, chiefly foliated zones and fractures.

Along the margin of massive sulfide and black schist, pyrite has been deposited on foliation planes and fractures in the black schist. Wisps of unreplaced black schist are present within the outer several feet of massive sulfide. However, no chlorite is present toward the center of the massive sulfide pipe. In a few specimens from the margin of the massive sulfide, the transition from a chloritic gangue to a normal quartz-dolomite gangue occurs within the field of a single thin section. The chlorite in the massive sulfide is a little coarser than in most of the black schist, and is randomly oriented. This lends support to the belief that there may have been a slight overlap between the periods of pyritic and chloritic replacement.

Chlorite was observed in a few exposures underground on the hanging-wall side of the massive sulfide pipe, but no masses of black schist were found. On the 600

level on the north side of the massive sulfide, chlorite was seen in two places: narrow seams of chlorite line fractures in the tuffaceous rocks, and at the contact of the massive sulfide, folded tuffaceous rocks were selectively replaced by pyrite and chlorite. Beds from 1 to 2 inches wide were replaced by pyrite and the intervening one-quarter inch beds were replaced by chlorite. The folded structure of the tuffaceous rocks is clearly preserved. The local chlorite on the hanging wall of the sulfide pipe shows either that some chlorite formed early, and perhaps contemporaneously with the early pyrite of the masive sulfide pipe, or that there was some leakage through the massive pyrite to the hanging-wall side during the main period of chlorite replacement.

COPPER MINERALIZATION

The introduction of copper and some zinc into the pyritic massive sulfide, black schist, and quartz porphyry followed the main period of chloritization, and most of the commercial copper ore formed during this period of mineralization. Intramineralization fracturing is indicated by the appearance of chalcopyrite in numerous intersecting veinlets. The spacing and distribution of these intersecting fractures at the time of the copper mineralization were fundamental in the formation of the copper ore shoots.

Three types of copper ore are present in the mine: massive sulfide, black schist, and quartz porphyry. The massive sulfide ore formed the largest and highest grade copper ore shoots and the quartz porphyry the smallest and lowest grade. Figure 14 graphically shows the ratio of these three types of ore. Table 21, analyses 3, 4, and 5, shows the average analyses.

MASSIVE SULFIDE ORE

Microscopic studies of the massive sulfide copper ore demonstrate that much of the chalcopyrite and associated sphalerite were introduced along fractures in the pyritic sulfide (pl. 8 *B*), a conclusion reached by Lindgren (1926, p. 73) and Rice (1920, p. 64). Irregular corroded margins of pyritic crystals indicate replacement of pyrite in the chalcopyrite-rich specimens, for the wider chalcopyrite bands contain very little pyrite. In part, however, chalcopyrite and associated sphalerite are interstitial to the small grains of pyrite, indicating that the replacement of the quartz-carbonate gangue was important. Recrystallization of pyrite is indicated because pyrite crystals lining chalcopyrite veinlets are as large as 0.6 millimeter in contrast to an average size of 0.1 millimeter in the main pyritic facies. In places, these coarser pyrite grains form a comblike structure, with chalcopyrite filling the spaces between. In a few of the chalcopyrite veinlets, the pyrite crystals surrounded by chalcopyrite are 0.5–0.6 millimeter in diameter. The association of coarser pyrite grains with chalcopyrite indicates recrystallization of some pyrite during copper deposition.

The mutual boundaries of chalcopyrite and subordinate sphalerite in the same veinlets indicate that these two minerals were deposited about simultaneously. The sphalerite generally forms small grains, 0.05–0.1 millimeter in diameter, embedded in chalcopyrite. Locally, veinlets or aggregates of pure chalcopyrite or pure sphalerite are present, but in the absence of crosscutting relations, their age relationship is indeterminable.

Galena occurs as rare small grains embedded in either sphalerite or chalcopyrite, and must have formed essentially contemporaneously with the chalcopyrite and sphalerite.

Quartz and carbonate gangue associated with chalcopyrite and sphalerite in veinlets indicates contemporaneous, or penecontemporaneous, deposition, despite the fact that elsewhere chalcopyrite and sphalerite replace quartz and carbonates. In a few places, carbonate veinlets clearly cut chalcopyrite and sphalerite, yet the younger veinlets locally contain chalcopyrite and a little sphalerite.

Bornite is present locally in the footwall side of the massive sulfide, associated with quartz. Bornite, chalcopyrite, and coarse-grained quartz form veinlets in fine-grained quartz. Early pyrite is fractured and partly replaced by chalcopyrite and bornite. Sphalerite is rare and appears older than the bornite.

Specular hematite is present sparingly in the massive sulfide ore, generally as narrow veinlets that contain very little carbonate gangue. Lindgren (1926, p. 73) stated that the specular hematite is younger than the pyrite but older than the other sulfide minerals. Examination of the few specimens of massive sulfide ore in which specular hematite occurs, shows that the specular hematite veins clearly cut chalcopyrite-sphalerite veinlets. Possibly specular hematite may have formed during several stages of mineralization, but because of the small amount, no positive statement can be made.

BLACK SCHIST ORE

The black schist copper ore is characterized by intersecting and branching veinlets of sulfide minerals, composed largely of chalcopyrite and pyrite (pl. 8*C*). In places, the chlorite has been extensively replaced by veins of almost solid sulfide 1 to 2 inches wide. The higher grade copper ore generally contains relicts of black schist between ramifying chalcopyrite veinlets.

Microscopic studies of the black schist ore confirm the importance of fractures in guiding the copper-bearing solutions in the black schist. Most sulfide veinlets

cut across the foliation planes, which are bent or broken adjacent to many of them. These sulfide veinlets in places are in contact with the original fine-grained chlorite. In other places, the chlorite is in random orientation and is much coarser grained adjacent to the sulfide veinlets. Some veinlets contain quartz and sulfide minerals; others contain dolomite and sulfides; whereas others contain both quartz and dolomite gangue.

In general, the mineralogy is simple: pyrite, chalcopyrite, and sphalerite are the important sulfide minerals, and are essentially contemporaneous, as a single veinlet will contain all three in mutual relationship. Some veinlets consist of nearly pure pyrite; others contain only chalcopyrite and sphalerite. Some fracturing of pyrite occurred before introduction of chalcopyrite and sphalerite, but in several places pyrite veinlets cut chalcopyrite. Arsenopyrite is rare and its age relative to the other sulfide minerals is in doubt. Under high magnification, galena and tennantite are observed in minute grains associated with the chalcopyrite and sphalerite.

The pyrite crystals range from 0.2 to 1 millimeter across, and generally are coarser grained than the pyrite in the massive sulfide. The smaller pyrite grains in the black schist are partly parallel to the foliation planes, indicating that the deposition of some of the pyrite has been controlled by foliation planes.

In most specimens containing coarse flakes of chlorite adjacent to the sulfide minerals, individual chlorite crystals project into chalcopyrite and have straight and sharp boundaries. This relationship is difficult to interpret, for it may mean that the chlorite has crystallized contemporaneously with the chalcopyrite, or the chalcopyrite has been deposited about the chlorite, or the chlorite has replaced the chalcopyrite.

Locally, quartz-carbonate veins are commercially important in the black schist ore, because they contain the same sulfide minerals as the black schist ore. Perhaps tennantite is more important quantitatively where the quartz-carbonate veins are abundant.

QUARTZ PORPHYRY ORE

Quartz porphyry ore consists of intersecting and ramifying chalcopyrite veins ranging from microscopic size to one-half inch in width, cutting through the quartz porphyry (pl. 8*D*). Chlorite is common along the margins of many of the chalcopyrite veins. Blebs of chlorite and chalcopyrite, in a linear arrangement, plunging very steeply to the north, are present locally in one quartz porphyry stope. This is the only recognized structural control possibly related to steeply plunging lineation in the foliated quartz porphyry. Elsewhere in the quartz porphyry ore, chalcopyrite veins cut across foliation.

Microscopic studies of the quartz porphyry ore show that some veinlets of chalcopyrite are associated with abundant chlorite, and others contain very little (pl. 8*D*). Where chlorite is abundant, the groundmass of the porphyry contains chlorite interstitial to the groundmass quartz, and sericite is absent. In quartz porphyry ore in which chlorite is sparse, sericite is interstitial to the quartz in the groundmass, indicating that the chlorite selectively replaces sericite in the early stages of chloritization. The sericite flakes that form the foliation in the quartz porphyry lie parallel and chalcopyrite veinlets clearly cut across the sericitic layers proving that the sericite is older than the chalcopyrite.

Chalcopyrite is the chief mineral in the sulfide veins, but pyrite and sphalerite are present. Deposition of all three minerals apparently occurred about simultaneously. Nests of randomly oriented chlorite occur along the sulfide veinlets, and individual crystals of chlorite project into chalcopyrite, similar to the relationship of these minerals in the black schist ore. The general association of chloritic margins along the chalcopyrite veins in the quartz porphyry ore indicates that part of the chlorite was formed at about the same time as the chalcopyrite, but fractured veins healed by chalcopyrite indicate that copper is later than the chloritization.

QUARTZ-CARBONATE VEINS

The last period of mineralization is represented by abundant quartz-carbonate veins and nodules that locally contain several sulfide minerals. These veins are not all the same age, for some are cut by the andesitic dikes and others cut the dikes. Some vein intersections show no displacement, whereas in others, older veins are offset as much as 6 inches. The quartz-carbonate veins cut through the black schist, dating the quartz-carbonate vein formation as younger than the bulk of the chloritization. Some chlorite, however, is contemporaneous with vein formation, as shown by chlorite selvages along the vein margin in quartz porphyry; the chlorite forms pale-green plates ranging from 1 to 2 millimeters in diameter, arranged normal to the vein wall, so that the chlorite on many surfaces appears fibrous. Some of the quartz-carbonate veins in the black schist have the same platy chlorite selvages, but this chlorite may be reworked chlorite from the black schist. In a few places, chlorite veinlets cut the quartz-carbonate veins.

The veins range from a fraction of an inch to 2 feet in width, but average between 1 and 2 inches. Where exposures permit observing the veins along strike, they commonly vary in width, and many gradually narrow

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and disappear within 20-foot distances. Quartz-carbonate nodules ranging from 6 to 8 inches in length and from 1 to 2 inches in width occur in some veins; these nodules are more common in the black schist where they are parallel to the foliation.

Quartz-carbonate veins also occur in the fault-zones along the footwall of the massive sulfide; in places, the vein material is brecciated, but many veins are non-brecciated, showing that no movement occurred after mineral deposition.

The veins are in all the rock types in the United Verde mine, including the gabbro on the hanging wall. They are rare in the quartz next to the gabbro, and in the massive sulfide, but most common in the black schist. Local concentrations of veins, however, are present in the quartz porphyry several hundred feet away from the massive sulfide. The tuffaceous rocks are cut by a few veins, particularly east and northeast of the massive sulfide pipe.

In some of the black schist, closely spaced veins a quarter of an inch wide have a wrinkled appearance, and contorted veins an inch wide occur locally in the quartz porphyry. The cause of the irregularity in the veins is unknown.

The mineralogy of the veins is simple: white granular quartz and dolomite are the two chief minerals. White calcite takes the place of dolomite in some veins, particularly those in the gabbro. Most of the veins are barren of sulfide minerals, but some contain sulfide, chiefly pyrite, chalcopyrite, pale sphalerite, galena, and tennantite. The veins away from the massive sulfide generally contain only pyrite. Rarely are all the sulfide minerals present in a single vein. The tennantite-bearing veins appear to have a consistent affinity for hosts of black schist or copper-bearing quartz porphyry. Some tennantite contains as much as 40 ounces of silver to the ton. In stopes where tennantite occurs in quartz-carbonate veins, the silver content of the ore was doubled and locally trebled over the average of the ore.

RELATIONSHIP OF ANDESITIC DIKES TO COPPER MINERALIZATION

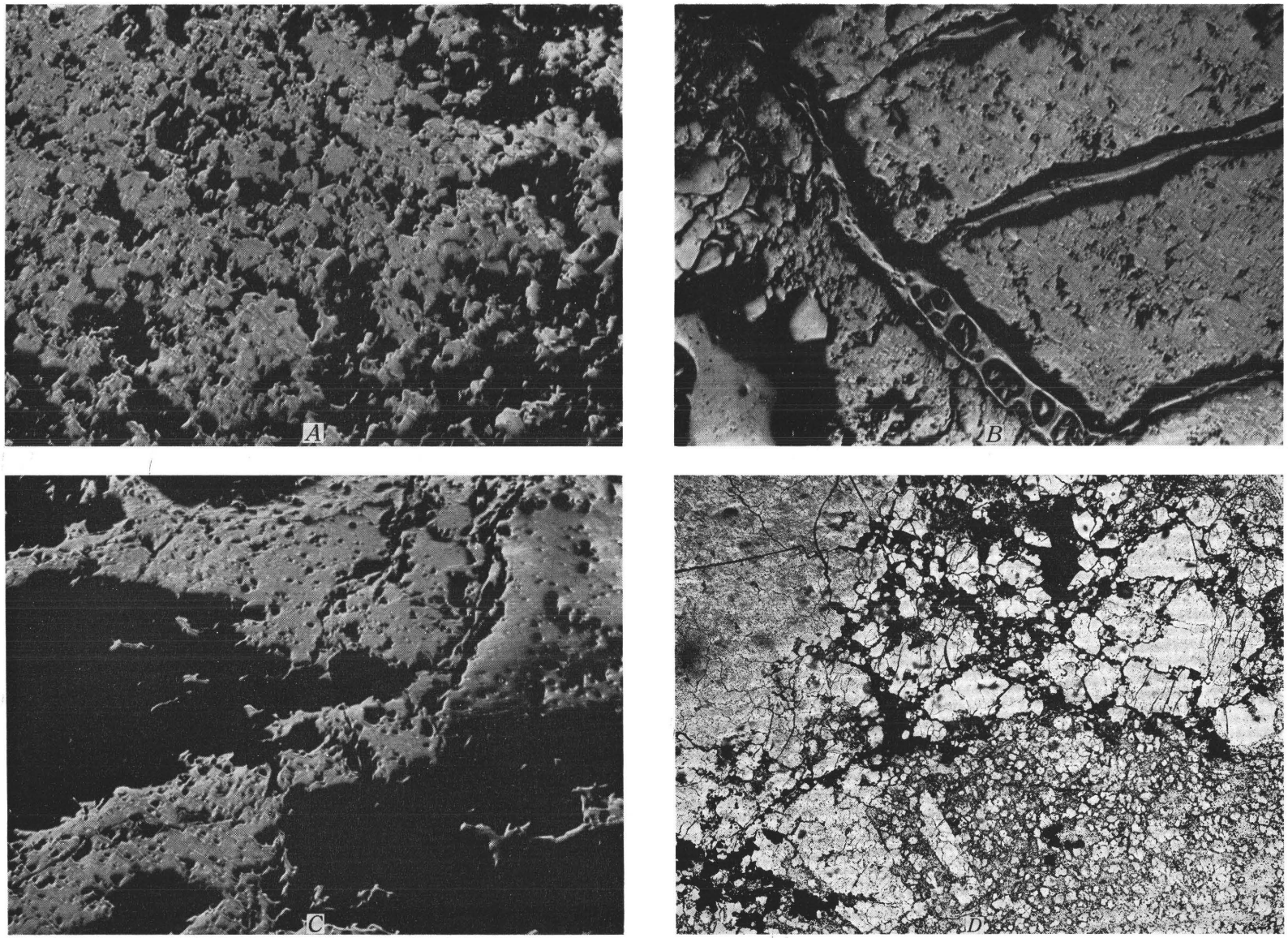
Early exploration entailed driving west or east along the andesitic dikes to intersect the north or northwestward-trending ore shoots, and the operators then believed that the dikes might have had important control in the location of ore shoots. For that reason, the geological staff of the United Verde mine paid special attention to the relationship of the dikes and copper mineralization (Reber, personal communication). Their final conclusions were that the dikes were later than the copper ore shoots. Our studies are in agreement with their conclusions.

It is true that the andesitic dikes in places contain scattered pyrite cubes and in a few places, chalcopyrite veinlets, 2 to 3 inches wide, which cut into the margin of the dikes. They are also cut by many quartz-carbonate veins, and pyrite and chalcopyrite are present in a very few of these veins within the dikes. Clearly, then, some sulfide mineralization occurred after intrusion of the dikes, but a much higher percentage of quartz-carbonate veins carry sulfide minerals outside of the dikes than within, indicating that most sulfide-bearing quartz-carbonate veins are older than the dikes. In one exposure, a quartz-carbonate vein containing pyrite is cut off by an andesitic dike that is cut by barren quartz-carbonate veins.

Where the dikes cut through the massive sulfide, their margins are straight and sharp, regardless of whether the dike cuts a chalcopyrite ore shoot or pyritic massive sulfide. Because fractures in the massive pyrite controlled the copper-bearing solutions to form ore shoots, the absence in the ore shoots of chalcopyrite veinlets in the andesitic dikes indicates that the dikes were intruded after the main chalcopyrite period of mineralization. The fact that the dikes carry many barren quartz-carbonate veins proves that the dikes were fractured.

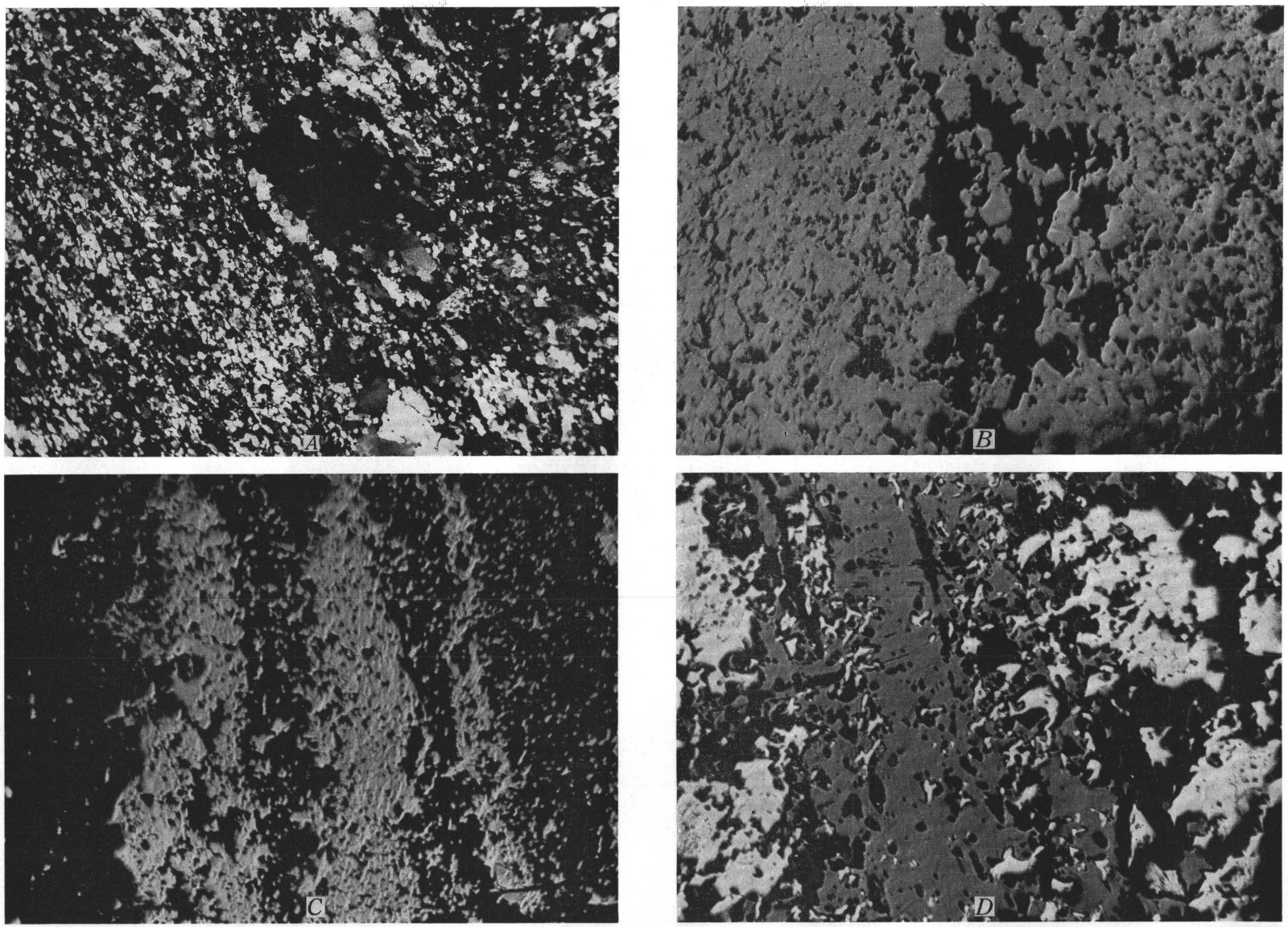
Evidence that the dikes are younger than the main copper deposition seems clearest in the black schist ore shoots where the barren andesitic dikes cut through the chalcopyrite-veined black schist. Here again, the absence of fracture-controlled chalcopyrite in the dikes shows that chalcopyrite mineralization occurred before intrusion of the dikes. Lastly, in one of the quartz porphyry ore shoots, where chalcopyrite with a little chlorite veins the fractured quartz porphyry, an andesitic dike, 1 inch wide, clearly cuts across chalcopyrite veinlets, and a larger dike, 2 to 3 feet wide, contains an inclusion 8 inches long of chalcopyrite-bearing quartz porphyry. The chalcopyrite veinlets in the inclusion are cut off sharply by the surrounding andesite.

The age of the andesitic dikes in relation to the main sulfide mineralization is of some importance as the intrusion of the dikes interrupted the hydrothermal period of activity. Also the relation is of interest because of the debate concerning the relative age of the diabase dikes to the sulfide mineralization at Noranda, Quebec (Price, 1934, p. 138; Wilson, 1941, p. 76–80). At Noranda, northward-trending diabase dikes apparently cut through the massive sulfide bodies and in most exposures, chilled zones of the diabase are in sharp contact with the sulfide mass. In other places, however, the diabase contains stringers of sulfide minerals, and the chilled margins of the dikes are fretted. It can be argued that the chilled selvage of the diabase resisted



PHOTOMICROGRAPHS OF UNITED VERDE SULFIDES

A. Lean pyritic massive sulfide (white) separated by quartz and carbonate minerals (black), × 52. *B*. Massive sulfide ore, showing veinlets of chalcopyrite (low relief) cutting pyrite, × 52. *C*. Black schist ore showing ramifying veinlets of chalcopyrite cutting chlorite (black), × 52. *D*. Quartz porphyry ore (thin section) showing veinlets of chalcopyrite (black) cutting fractured quartz porphyry. Gray material interstitial to quartz (white) in lower right hand corner is chlorite, × 43.



PHOTOMICROGRAPHS OF IRON KING MATERIAL

A. North-end quartz (thin section, crossed nicols) showing cataclastic-granoblastic texture resulting from intense shearing, × 43. *B*. Lean pyritic massive sulfide (white) separated by ankerite and quartz (black), × 52. *C*. Alternating bands of granular pyrite (white) separated by quartz and ankerite (black); probably relict after foliation, × 52. *D*. Massive sulfide ore; sphalerite (gray) replacing pyrite (white) along microscopic shear planes, × 52.

the sulfide replacement, or that the sulfide in the diabase resulted from redistribution of older sulfide, mobilized by the dike. The only purpose of referring to the problem at Noranda is to indicate that essentially unmineralized dikes cutting massive sulfide bodies are pertinent subjects of debate as to age relationship. Straight, sharp dike contacts in massive sulfide bodies may not prove a later age for the dikes. However, the relationship at Jerome leaves no other interpretation but that the dikes are younger than the chalcopyrite.

MINOR STRUCTURAL FEATURES

CHLORITE- AND SULFIDE-BEARING FRACTURES

The "plumbing system" dominated by fractures controlled deposition of chlorite and chalcopyrite, and the foliation planes served merely as distributing channels outward from the fractures. The mineralized fractures have a definite preferred orientation, as illustrated on figure 20, which shows the attitudes of 190 measured mineralized fractures, plotted on the lower hemisphere of an equal-area net. A description of this method of plotting the poles of strike and dip measurements is given by Billings (1942, p. 118–122). The usual method of contouring concentrations could not be used. The dip vector of many of the fractures, plotted as vertical on the net, was not determined because the fractures were observed only on the backs of drifts and stopes. Nevertheless, figure 20 illustrates clearly that all but a very small percentage of the mineralized fractures strike within the range from N. 30° E. to N. 30° W. and that the dips are 60° or steeper toward the east and west. The attitude of the greatest concentration of mineralized fractures essentially parallels the regional foliation, N. 20° W., 80° E. Because foliation preceded fracturing, this coincidence is expectable.

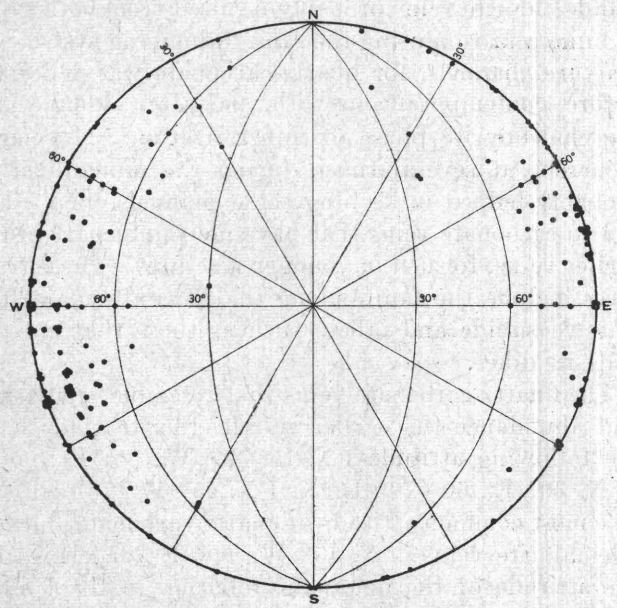


FIGURE 20.—Poles of mineralized fractures in United Verde mine, plotted on lower hemisphere of equal-area net.

Wherever observed, the direction of relative movement on the fractures is reverse. Apparently no consistent age relationship exists, for northeastward-trending fractures do not consistently offset northwestward-trending fractures, or conversely. Presumably this fracture set represents conjugate shears and the reverse movement indicates compression and shortening in a general easterly direction.

Several veins trend N. 30° W. for a short distance and deviate to N. 30° E. indicating that the mineralizing solutions followed first fractures in one direction and then the other. If this deviation took place on a small scale, the resulting vein would strike north. Presumably deviation from N. 30° W. to N. 30° E. on different scales is a partial explanation for the spread in strikes of the mineralized fractures from N. 30° W. to N. 30° E.

The veins produced by the fracture filling are black schist veins, sulfide veins, black schist-sulfide veins, and quartz-carbonate veins. Discussion of the structure of the quart-carbonate veins is reserved for a later section. For all veins measured in the United Verde mine, the proportion of black schist veins to sulfide veins to black schist-sulfide veins is about 1:2:2.

Black schist veins in the quartz porphyry are well displayed on the south wall of the open pit where zones of black schist veins pass northward into tongues of solid black schist that, in turn, form a nearly continuous envelope of black schist along the footwall of the massive sulfide. Similar black schist veins in quartz porphyry are common along the margins of the solid black schist in the underground workings. In the upper levels east of the massive sulfide, black schist veins are present in the tuffaceous rocks of the Grapevine Gulch formation, adjacent to the masses of black schist.

The black schist veins were commonly reopened and pyrite and chalcopyrite introduced into the new fractures, and in some places sufficient copper was added to form ore.

Sulfide-bearing fractures are abundant in quartz porphyry, black schist, and in the massive sulfide in the ore shoots. In a few places, a random network of sulfide veinlets forms a zone parallel to one or the other dominant trend of fractures. Where these zones are observed in the black schist, patches as much as 2 feet across are completely converted to sulfide, chiefly chalcopyrite.

The spatial position and outline of ore shoots reveal clearly that the fracture system was an important con-

trol. All or nearly all ore shoots plunge more steeply than the sulfide pipe. This plunge is probably controlled by the dip of the mineralized fractures and the plunge of their intersection. As the average dip of the mineralized fracture is about 80°, and very few dips are as low as 60°, ore shoots controlled by these fractures plunge steeply, somewhat steeper than 60°. The map of an ore shoot commonly reflects the fracture control by elongation parallel to the dominant fracture direction. In several ore shoots, the maximum elongation from one level to another corresponded to either the northeast or the northwest strike of the fractures.

Premineral fractures partly bound some stopes; locally these fractures are veined and ribboned with sulfide and quartz-carbonate. Sulfide minerals are not limited to one or the other side of these controlling fractures, but the grade in copper may be much higher on one side than the other; and in places, the lower grade may be below the cutoff grade used in mining.

Foliation has acted largely as a distributing element for the mineralizing solutions fed into the rocks by the fractures. Along the south wall of the open pit, this function of foliation is shown clearly by the orientation and growth of the chlorite in the fractures in the quartz porphyry. The chlorite grew outward from the fracture with the basal pinacoid of the chlorite parallel to the foliation in the quartz porphyry (mimetic crystallization). At the intersection of fractures, irregular knots or masses of chlorite formed, and the foliation in the black schist (chlorite) is pronounced and continues with that in the surrounding porphyry.

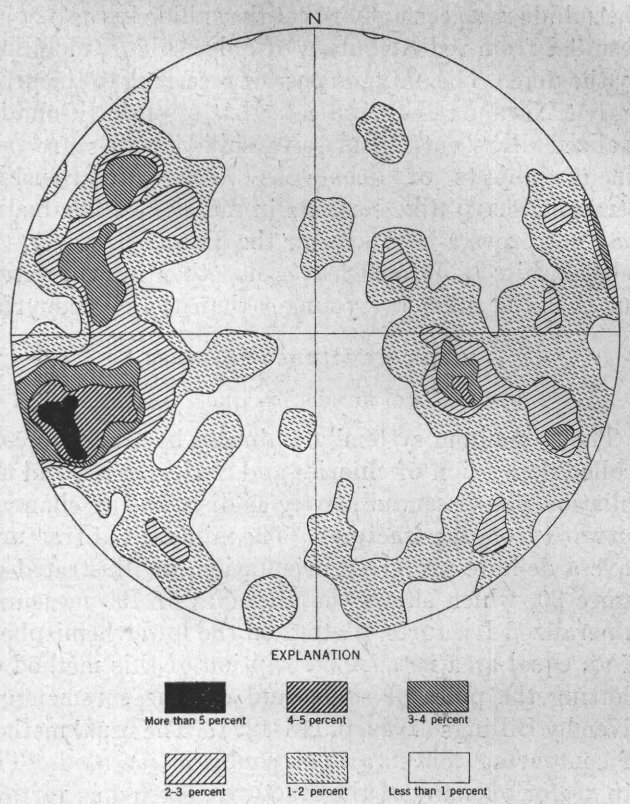


FIGURE 21.—Contour of concentrations of attitudes of quartz-carbonate veins, United Verde mine, plotted on lower hemisphere of equal-area net.

STRUCTURE OF THE QUARTZ-CARBONATE VEINS

Quartz-carbonate was deposited throughout a long period during the formation of the United Verde deposit. Early quartz-carbonate and pyrite together formed the bulk of the massive sulfide pipe before the introduction of the ore minerals. Quartz-carbonate continued to be deposited as veins during the ore-forming mineralization and some quartz-carbonate veins represent the latest hypogene mineralization. The long history of quartz-carbonate deposition accounts for the wide range in attitudes of the quartz-carbonate veins as compared to the sulfide-chlorite veins.

In order to gain a more precise understanding of the fracture pattern that controlled the quartz-carbonate veins, the attitudes of 177 quartz-carbonate veins were measured in the vicinity of the ore deposit. The poles to the planes of the veins were plotted on an equal-area net, and the resulting concentrations contoured; figure 21, which shows two significant features: (1) the quartz-carbonate veins have the same range in attitude and relative intensity of concentration as the earlier sulfide-chlorite veins, and (2) they are also concentrated in a zone that ranges in strike from north to N. 30° E. and ranges in dip from 30° to 80° W. There are also a higher percentage of dips less than 60°. The similarity in attitude between some quartz-carbonate veins and the sulfide-chlorite veins probably resulted from both types of mineralization using the same "plumbing system" of fracture channels, for quartz-carbonate was deposited before, contemporaneous with, and after chlorite and the chalcopyrite phase of mineralization. Movement recurrent at several times during the mineralization probably helped in keeping these channels open. The quartz-carbonate veins that have no counterpart in the earlier veins, formed in younger fractures. These fractures include the faults along the footwall side of the massive sulfide and others such as those that cut the andesite dikes.

The quartz-carbonate veins that resemble in attitude and abundance the earlier sulfide-chlorite veins have the following attitudes: N. 15°–30° W., 80° E.; north to N. 30° E., 60°–70° E.; and N. 25° W., 75°–80° W. The most common attitude of quartz-carbonate and sulfide-chlorite veins is N. 15° W., 80° E., which is also the attitude of the dominant foliation in the United Verde mine. The faults along the footwall of the massive sulfide pipe probably controlled the quartz-car-

bonate veins that strike N. 23° E. and dip 78° NW. The quartz-carbonate veins that strike N. 25°–45° E. and dip 35°–50° NW. also have no counterpart in the earlier veins. The fractures localizing these veins may have resulted from the same deformation that produced the faults along the footwall of the sulfide pipe.

FOOTWALL FAULTS

The footwall faults occur along the footwall zone of the massive sulfide pipe in the intermediate and lower levels of the United Verde mine (see 3,300 level, pl. 7). These faults first appear as definite structural features on about the 1,500-foot level and gradually become more conspicuous and continuous with depth. From the 2,250 to the 3,450 levels, the faults are pronounced, and extend throughout the length of the massive sulfide body from the 2,850 to the 3,150 levels. Below the 3,450 level, the faults diminish in number, but are present on the 4,500 level.

Two distinct trends characterize the faults. The dominant ones strike northeast and dip west at an angle somewhat less steeply than the footwall of the massive sulfide. These faults appear to strike slightly more easterly than the trend of the massive sulfide body, so that the massive sulfide body and faults diverge northward. These northeastward-trending faults occur only where the footwall of the massive sulfide trends northeastward. The subordinate faults strike northwest and dip northeast; they occur only where the footwall of the massive sulfide trends northwestward. The faults do not exist apparently near the upper part of the deposit where the horizontal section of the massive sulfide pipe is about equidimensional, but they appear where the pipe becomes crescent shaped. The northwestward-trending limb of the massive sulfide pipe is not well defined above the 2,100 level, and on this level the northwestward-trending footwall faults are first perceptible. The northwestward-trending limb of the massive sulfide pipe is not as long nor as wide as the northeastward-trending limb; this is reflected in the number, continuity, and apparent strength of the faults. Where the massive sulfide pipe is divided on the lower levels, the faults follow the outlines of individual segments, and as the massive sulfide pipe dies out, the faults disappear.

The spacing of the footwall faults next to the massive sulfide pipe is not fortuitous, and it is possible that the faults came into existence because of lithologic difference between the massive sulfide and the adjacent black schist and quartz porphyry. Stress would more likely be relieved along such contacts.

The age of these faults is debatable. Pyrite in the fault zones indicates deposition of sulfide minerals before faulting. This evidence is manifested in only two places, on the 3,750 and 3,450 levels. The footwall faults displace andesitic dikes in many places, but in one locality, an andesitic dike apparently intrudes fault gouge. Quartz-carbonate forms veins along the faults and in places part of the fault breccia and gouge. If the faults had been in existence during the major period of intrusion of the andesitic dikes, probably at least a few of the dikes would have used them as channels. In summary, these footwall faults probably originated after the andesitic dikes were intruded but during or just before the late quartz-carbonate mineralization.

RECURRENT MOVEMENT

The indications of recurrent movement consist chiefly of earlier veins cut by later ones.

The northwestward-trending foliation is the oldest planar structure recognized, and recurrent movement occurred along it no less than three times. The period of fracturing produced reopenings along the foliation direction, these fractures were chloritized, and later recurrent movement permitted introduction of sulfide and quartz-carbonate. The third period of movement, perhaps more penetrative than the others, produced a weak to nearly imperceptible foliation in the andesite dikes that coincides with the northwestward-trending foliation in the host rock, generally quartz porphyry. The foliation in the andesite dikes was not utilized by mineralizing solutions, not even those depositing the final quartz-carbonate. In one place on the 3,000 level, an andesite dike is offset by a northwestward-trending zone of fracture and foliation that contains appreciable amounts of chlorite. Here, at least then, faulting occurred during this later period of deformation.

In summary, in chronological sequence movement occurred, (1) at the time of folding (northwestward-trending foliation), (2) probably after formation of the massive sulfide pipe but before deposition of chlorite (the period of fracturing), (3) after deposition of chlorite but before deposition by chalcopyrite (ore-forming period), and (4) after intrusion of andesitic dikes and deposition of some quartz-carbonate veins.

ORE SHOOTS

The massive sulfide pipe and black schist fringe on the footwall were completely formed by the time chalcopyrite was introduced to form the copper ore shoots. Fracturing of the massive sulfide, black schist, and quartz porphyry made these rocks permeable for the copper-bearing solutions which spread throughout an irregular zone extending from the southwestern tip of sulfide and along the footwall of the massive sulfide and along the east margin of the sulfide pipe. Shattered zones within the massive sulfide pipe allowed the copper-bearing solutions to rise upward almost to

the hanging wall. The chalcopyrite mineralization formed, in its broadest aspect, a stockwork in which the richer zones of ore are connected by leaner zones, and the ore shoots have assay walls.

The ore shoots generally plunge northward more steeply than the massive sulfide pipe, so that many ore shoots bottom in quartz porphyry or black schist in the footwall of the massive sulfide pipe, and the tops of the ore shoots are well within the massive sulfide (pl. 9). In the upper levels, massive sulfide ore is more common (fig. 14) and in the lower levels black schist ore and quartz porphyry ore are more common.

The outline of ore shoots is controlled in many places by the fracture pattern. North-northwesterly elongation is more common than north-northeasterly; and in a few places, elongation in both directions occurs within a single ore shoot (see stopes 9*R* and 9*Q*, 2,400 level, fig. 22). Elongation along one direction is probably the result of relative abundance, size, and continuity of the fracture. A few ore shoots, particularly in the massive sulfide, are so irregular in outline as to suggest lack of predominant fractures.

The contacts between massive sulfide, black schist, and quartz porphyry exerted some control on the shape of the ore shoots, and below the 1,200 level, most of the ore shoots are located along the contact of massive sulfide and black schist. Stopes on the 1,800, 1,950 and 2,100 levels occupy almost the entire arc of the footwall contact of the pipe. On the 2,250 and 2,400 levels, stopes are common along the footwall of the pipe and along the black schist-quartz porphyry contact (fig. 22); in several places, the stopes reach from one contact to the other. Stopes on the 2,550 level are along both contacts, but those along the contact of black schist with the sulfide pipe are chiefly in the black schist.

Ore shoots along the east margin of the massive sulfide are related to the contact of the massive sulfide pipe; and where this forms a prong extending south-southeastward, ore shoots are located along the east and west contacts (fig. 16).

Black schist was a more favorable host rock than quartz porphyry (fig. 22) and less favorable than massive sulfide. Most of the ore shoots bottom in quartz porphyry, along quartz porphyry-black schist contacts, or in black schist beneath the footwall of the sulfide pipe. The ore shoots in quartz porphyry are generally spatially related to tongues of black chist; and as the ore shoot passes into black schist at higher levels, it widens, and the quartz porphyry on the same level is weakly mineralized. On higher levels, above the black schist ore, the massive sulfide is the favorable host rock, and ore shoots are distributed along the contacts and within the sulfide pipe.

10–O ORE SHOOT

Three ore shoots merit individual description either because of their exceptional size and grade or because

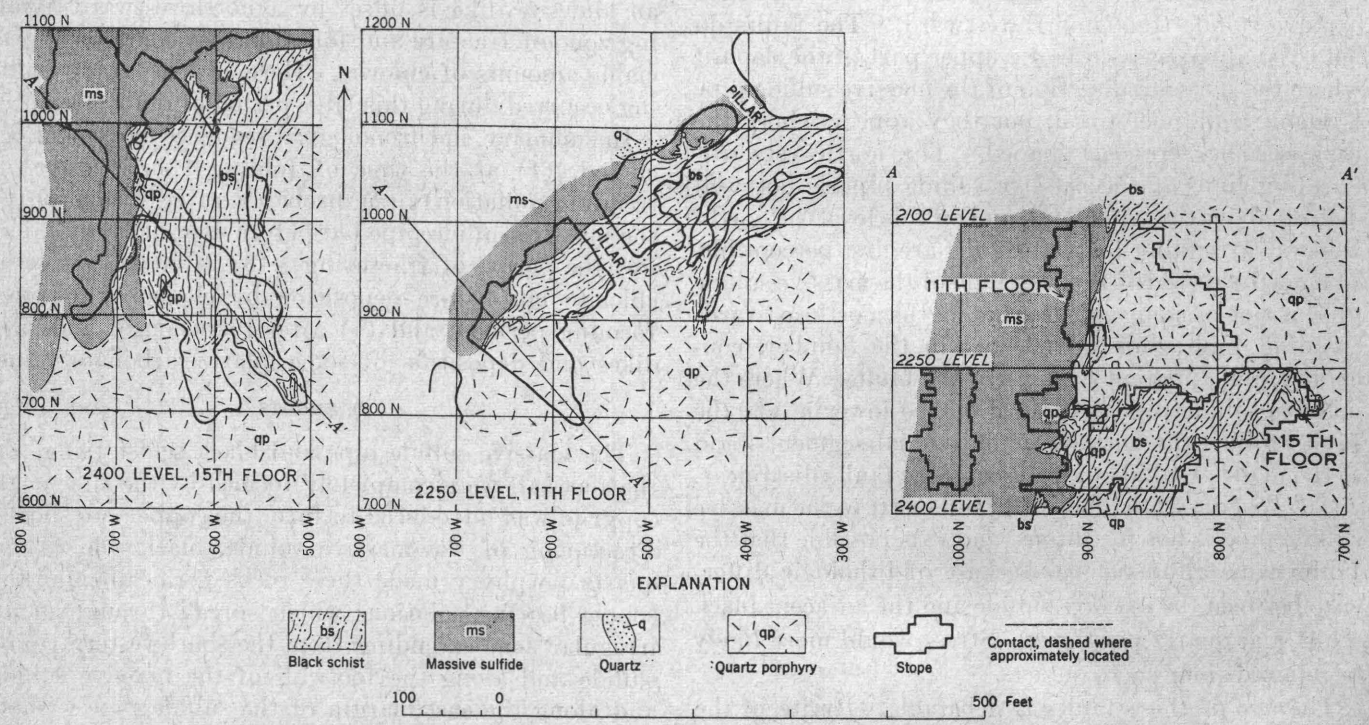


FIGURE 22.—Maps and sections of stopes. Stopes 9*R* and 9*Q*, 2,400 level, 15th floor, show northwest and north-northeast trends in same ore shoot. Stopes on the 2,250 level, 11th floor, show ore shoot including massive sulfide, black schist, and quartz porphyry. Section shows that black schist is more favorable host rock than quartz porphyry for copper mineralization. United Verde mine.

they exemplify a particular kind of structural control. Foremost among these is the 10–O shoot, which was unparalleled in the United Verde mine for size and grade (pl. 9).

The bottom of the 10–O ore shoot was about 40 feet below the 2,250 level, and the top was about 20 feet above the 1,200 level—a vertical range of about 1,100 feet. On the 2,250 level the ore shoot was only 230 feet long and 55 feet in maximum width, but the size increased abruptly upward so that on the 2,100 level it was 800 feet long and ranged in width from 20 to 100 feet. On succeeding levels upward, the size of the ore shoot increased to a maximum on the 1,650 level where the ore shoot was more than 700 feet long and 100 to 250 feet wide. Above the 1,650 level, the ore shoot divided into several parts, most of which were well within the bounds of the massive sulfide pipe. These segments diminish in size with height, and cease at altitudes above the 1,500 level, the largest segment reaching a few feet above the 1,200 level.

A study of the distribution of copper in the 10–O ore shoot revealed that a direct relationship existed between the local dimensions of the ore shoot and copper content. Where the ore shoot had the largest horizontal area, the copper content was highest. The richest ore generally was in a zone in the central part of the ore shoot, and copper content diminished outward from this central core to an assay wall that determined the limits.

WADE HAMPTON ORE SHOOT

The Wade Hampton is the second largest ore shoot in the deposit, reaching its maximum size in the 1,000 and 1,200 levels (pl. 7). It starts in quartz porphyry between the 1,500 and 1,350 levels as three separate ore shoots that integrate upward into the Wade Hampton. The Wade Hampton stope tops on the 8th floor above the 800 level, but the 15–L stope lies directly above the Wade Hampton between the 600 and 500 levels and is probably part of the same mineralized zone. On the 1,200 level, the ore shoot extends along the contact of the massive sulfide and is arcuate. The length along the center is about 260 feet and the maximum width about 80 feet. On the 1,000 level (pl. 7) the ore shoot is within the massive sulfide pipe and is more equidimensional in shape. The maximum dimensions are about 200 by 150 feet. Some of the richest ore in the mine was found in the Wade Hampton. The pattern of distribution of copper in the Wade Hampton ore shoot is similar to that in the 10–O ore shoot.

16–C ORE SHOOT

The 16–C is a long narrow ore shoot in quartz porphyry south of the massive sulfide pipe and the body of black schist (pl. 7, 500–800 S.; 300 E.). It extends from the surface to the 600 level where it has a length of 180 feet and a maximum width of 30 feet. On the 500 level the length is 570 feet and the maximum width 60 feet, but the average width is about 30 feet. The trend of the ore shoot ranges from north-northwest for the southern part to north for the northern part; the dip is nearly vertical (fig. 23).

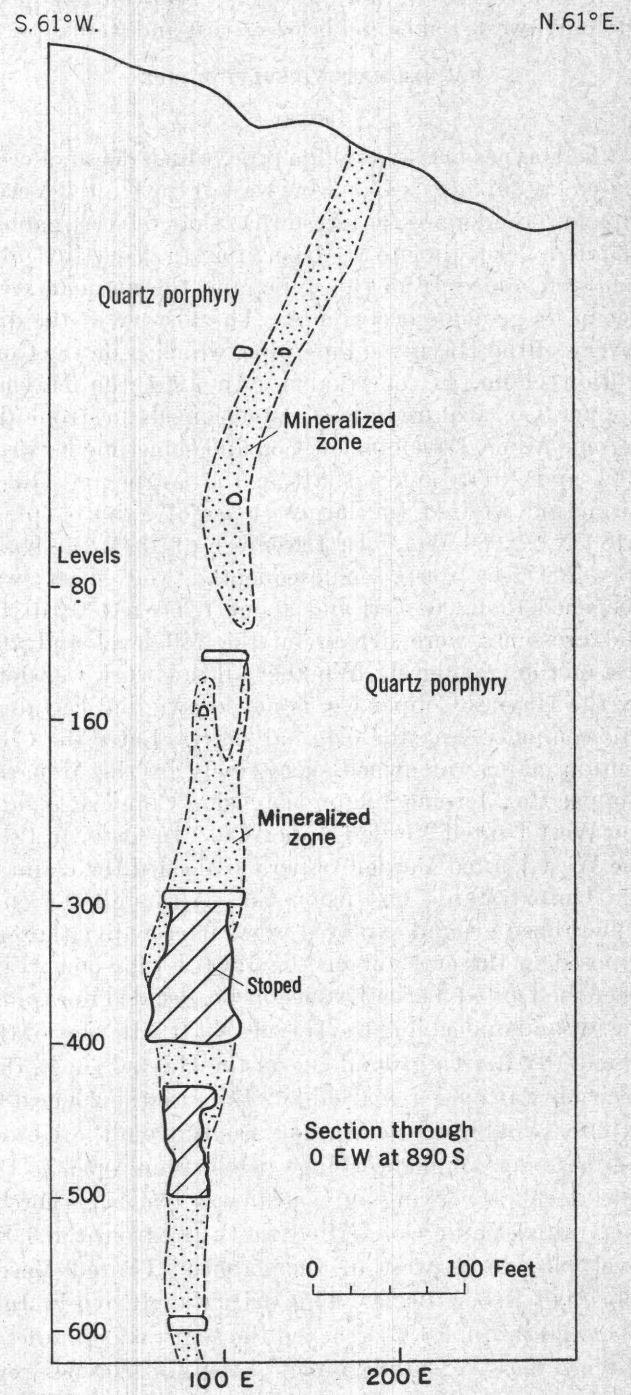


FIGURE 23.—Section through 16–C ore shoot, United Verde mine.

The relationship of the attitude of the ore shoot to mineralized fractures warrants consideration. The ore shoot is made up of many small veins and veinlets of chalcopyrite. These veins strike north-northeast and north-northwest, and dips are east or west, some as low as 60° (fig. 20). But the northern part of the ore shoot strikes north, and the dip is essentially vertical, except for the upper levels. Apparently, the attitude of the ore shoot represents a compromise between northeast and northwest strikes and between east and west dips.

HAYNES MASSIVE SULFIDE PIPE

HISTORY

The Haynes massive sulfide pipe, which does not crop out on the surface, is about due west of the United Verde pipe on the opposite (western) side of the gabbro (United Verde diorite of Reber, 1922). Very little ore has been mined from this pipe, and the main interest lies in its geologic occurrence. The history of the discovery of the Haynes sulfide pipe, which is on the Contention claim, is remarkable. In 1904 the Haynes Copper Co. acquired the Contention claim from the Jerome Mines Development Co., and sometime between 1907 and 1911, sank a shaft in the Grapevine Gulch formation west of the largest mass of gabbro (pl. 5, 1,780 N.; 2,080 W.). In December of 1911, the shaft was 700 feet deep. Subsequently, the shaft was deepened to 1,200 feet, and about 1,700 feet of drifts and crosscuts were driven on the 700 level and 700 feet on the 1,200 level. Whether all this work was done by the Haynes Copper Co. is not known, but certainly this company sank the original shaft. Later the Contention claim was owned successively by the Monarch Copper Co., Jerome Victor Extension Copper Co., and the West United Verde Copper Co. In October 1919, the West United Verde Copper Co. deeded the claim to the United Verde Extension Co. Although the only mineralized ground explored was a narrow pyritic zone exposed in the crosscut on the 1,200 level, some years later the United Verde Extension Co. decided to explore the ground underlying the Haynes shaft. In view of the essentially barren ground cut by the Haynes shaft, this decision was most remarkable. Because the zone to be explored could be reached more easily from the United Verde underground workings, which were open to the 3,000 level, a working-agreement was reached whereby the United Verde Co. drifted westward from the 3,000 level into rocks west of the gabbro (United Verde diorite of Reber, 1922). This drift was started in July 1930 and in June 1931 reached the south margin of the Haynes massive sulfide pipe. After the Haynes pipe had been explored on the 3,000 level, diamond-drill holes from the 3,000 and 3,750 levels of the United Verde mine, and a drift on the 1,500 level further delimited the size and shape and determined the content of base metals.

GENERAL FEATURES

The Haynes pipe is a steeply plunging body of massive sulfide and quartz that is relatively small in cross section compared with the length along the axis (fig. 24). It extends from a short distance above the 2,700 level downward to an undetermined point between the 3,450 and 3,700 levels (United Verde mine levels). On the 3,000 level, the sulfide pipe is triangular, each leg of the triangle being about 120 feet long; the cross-sectional area of the pipe appears to diminish both upward and downward from there. On the 2,700 level, the two maximum dimensions of the irregular-shaped pipe are 70 feet to the north and 60 feet to the east-northeast. With depth the sulfide pipe elongates parallel to the gabbro contact. On the 3,450 level the pipe extends for about 100 feet along the contact and is 40 feet wide.

The bearing of the axis and the plunge of the sulfide pipe are variable. Between the 2,700 and 3,000 levels the axis bears N. 76° E. and the plunge about 67° NE. Between the 3,000 and 3,450 levels, however, the axis bears about N. 60° E. and the plunge about 75° NE.

Gabbro (United Verde diorite of Reber, 1922), quartz porphyry, and Grapevine Gulch formation are the rocks in the vicinity of the Haynes massive sulfide pipe (fig. 24). The gabbro forms the hanging wall (east wall) and quartz porphyry, the footwall (west wall). Actually, gabbro bounds the pipe on the northeast, east, and southeast sides throughout most of the vertical extent, forming an inverted trough in which the massive sulfide pipe lies. Where this inverted trough dies out above the 2,700 level, the gabbro extends over the top of the pipe. Similarly, where the inverted trough dies out in depth, the gabbro underlies the pipe (fig. 24).

The quartz porphyry along the footwall of the pipe is an unusual type—possibly it is actually a quartz-bearing tuff. It consists of numerous quartz phenocrysts in a schistose groundmass containing abundant sericite and chlorite. In places in the quartz porphyry, fragments and the indistinct bands resemble relict bedding. The fragments, bands, and chlorite indicate a sedimentary origin, but the character, distribution, and abundance of quartz phenocrysts(?) indicate quartz porphyry.

The Grapevine Gulch formation is west of the quartz porphyry. It appears typical of the fine- to medium-grained facies found in the United Verde area.

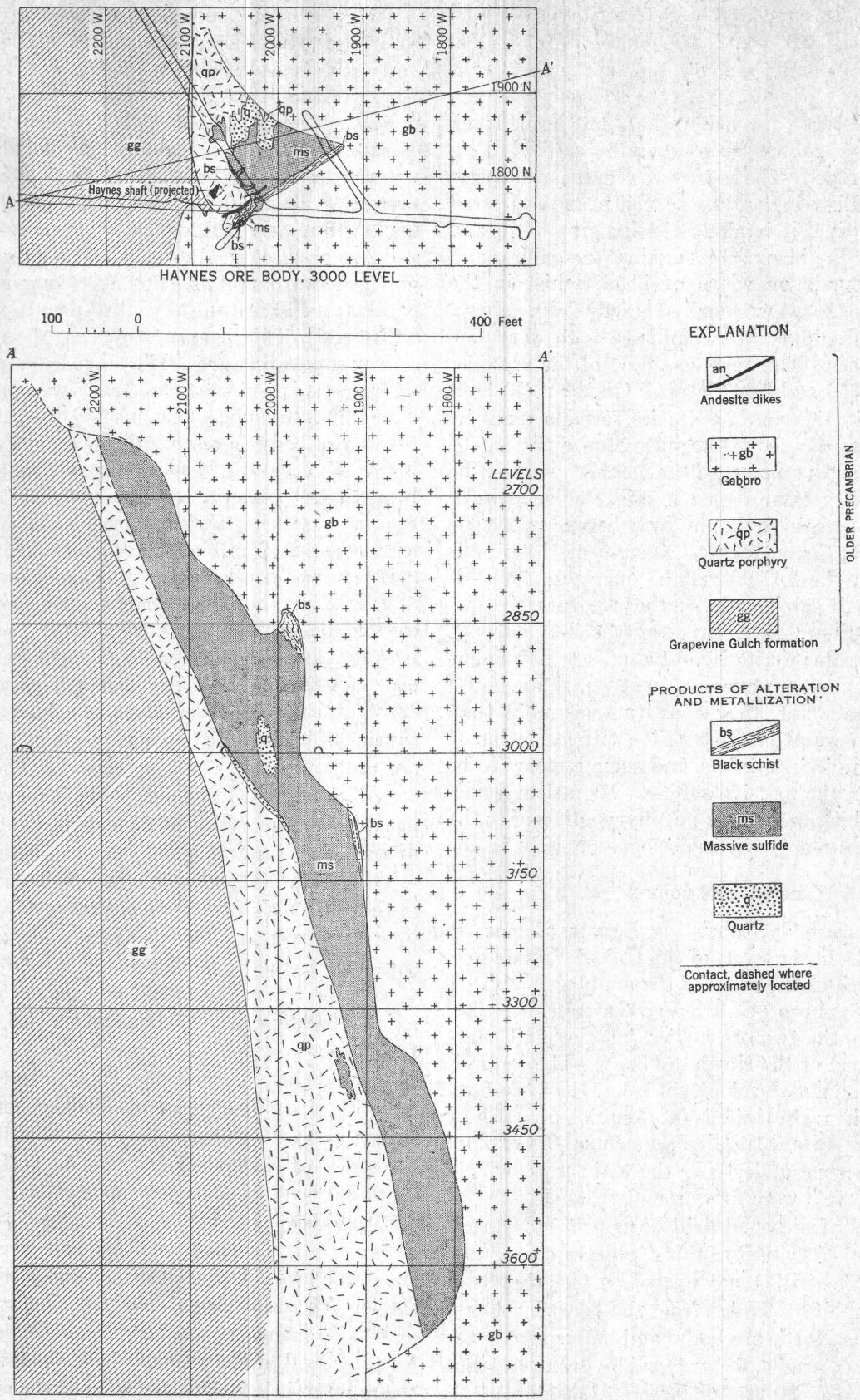


FIGURE 24.—Map and section of Haynes ore body. Furnished by Phelps Dodge Corp.

Black schist is associated with the Haynes sulfide pipe between the 2,850 and 3,000 levels. It lies between the gabbro and massive sulfide along the southeast side of the pipe; and from the 3,000 level downward to a point not far below the 3,150 level, black schist separates gabbro from massive sulfide along most of the contact. There are two significant aspects to the black schist: in contrast to that near the United Verde sulfide pipe, it is along the hanging wall, and the position of the black schist against the gabbro and the small stringers or wisps of black schist in the gabbro indicate that it was derived chiefly from gabbro.

The massive sulfide pipe comprises both quartzose and sulfide facies. The quartzose part of the pipe lies between the 2,850 and 3,300 levels. On the 2,850 level or just below it, the quartzose facies forms a small irregular mass on the north margin of the pipe; on the 3,000 level it makes up most of the northern part of the pipe; and with greater depth it migrates southward along the west margin so that midway between the 3,000 and 3,150 levels the entire west margin of the pipe is quartzose. On the 3,150 level the quartzose facies is limited to a small patch at the southeast corner.

The sulfide minerals are pyrite, pyrrhotite, chalcopyrite, and sphalerite; pyrite is most abundant. Sphalerite and chalcopyrite occur in about equal amounts; where most abundant, they constitute no more than from 10 to 15 percent of the massive sulfide. Some of the chalcopyrite is in veinlets, and some appears to be disseminated in the massive sulfide. Magnetite occurs in the black schist and quartz porphyry adjacent to the pipe, but it was not recognized megascopically in the massive sulfide.

NORTH ORE BODY

The North ore body, consisting largely of massive sulfide, is on the lower levels of the United Verde mine, and to the north of the roots of the main ore body. A small prong of gabbro (United Verde diorite of Reber, 1922) separates the two ore bodies, but the main mass of gabbro is west of the North ore body. The preliminary exploration and development below the 3,000 level revealed that the main United Verde sulfide pipe passed downward into several rootlike segments. To explore adequately the zone underlying the main pipe, widely spaced lower levels were driven, and from these levels nearly 17 miles of diamond-drill holes probed the rock in search of ore. As a part of this exploration program, some holes were drilled northward to test the rocks north of the gabbro. Holes from the 3,300 level first cut through the North ore body, and subsequently the ore body was explored and developed by drifts and diamond-drill holes on the 3,750, 4,050, and 4,500 levels (pl. 7) and by diamond-drill holes on the 3,450 level.

The North ore body comprises a group of interconnected thin lenses of massive sulfide. It extends vertically from a short distance above the 3,450 level downward to about 150 feet below the 4,500 level. On the 3,450 level (pl. 7) the ore body comprises four lenses of massive sulfide, trending westward or west-northwestward and ranging in length from 130 to 35 feet and averaging about 10 feet in width. On the 3,750 level, the North massive sulfide body comprises one large lens and two or three smaller ones. The largest lens trends northwestward and is 2,700 feet long and ranges in width from 40 feet in the central part to about 5 feet at either end. This lens as shown on the map is gently convex southwestward. The other lenses are small, and their shape is conjectural, as each was cut by only one diamond-drill hole. On the 4,050 level (pl. 7) the North ore body is essentially a narrow lens trending N. 30° W., about 600 feet long and ranging in width from 50 feet near the south end to 5 feet at the north. Similarly, on the 4,200 level the North massive sulfide body consists of only one lens, irregular in shape. The northern and southern parts trend a few degrees west of north, but the central part trends about N. 45° W., the average strike being N. 20° W. It is about 650 feet in length and ranges in width from 70 feet in the southern part to 7 feet in the northern part. Below the 4,200 level, the body divides into several segments. On the 4,500 level (pl. 7) several pods of massive sulfide are crudely alined north-northwest. The mineralized zone is about 500 feet long. One massive sulfide body has a maximum width of 90 feet, and another a length of 275 feet.

The North massive sulfide body is nearly concordant to the structure of the adjacent rocks. On the map, the long axis, whether in one segment or several, nearly conforms to the strike of the dominant foliation in the Grapevine Gulch formation and quartz porphyry and of the bedding in the Grapevine Gulch formation. Locally, however, foliation and the contacts of the massive sulfide are discordant. Changes in the attitude of foliation are reflected in the attitude of the massive sulfide bodies. This parallelism results in variation in the strike of the body from west or west-northwest on the 3,450 level to northwest or north-northwest on the 3,750 and lower levels.

The dips in the massive sulfide body are to the east on the lower levels, but are to the west on higher levels as shown on section through North ore body (pl. 7). The quartz is on the footwall and the black schist is on the hanging wall of the zone on the lower levels, similar to the relation below the 3,450 level of the main pipe (fig. 13). On higher levels in both ore bodies, the reverse

relationship exists. The two massive sulfide bodies have a similar relationship to gabbro in that the gabbro overhangs both.

RELATIONSHIP TO ADJACENT ROCKS

The North massive sulfide body is largely in the Grapevine Gulch formation, at or near the contact with the quartz porphyry. Normal gray quartz porphyry lies to the east of the lenses of massive sulfide, and in part has been chloritized adjacent to them. West of the massive sulfide body are one or two bands of "purple porphyry" that may or may not be intrusive quartz porphyry; the doubtful origin of this rock has been discussed on page 104. This rock is indicated as quartz porphyry on the 3,450, 4,050, and 4,500 levels as shown on plate 7. West of the mineralized zone, excellent bedding is recognizable in diamond-drill cores and to the east, undisputed quartz porphyry is present. In the mineralized zone, the rocks are highly foliated, and the differentiation of quartz-bearing crystal tuff from the quartz porphyry is difficult; the contacts are, in part, conjectural.

The gabbro contact, viewed in a longitudinal section (north-south) overhangs the North massive sulfide body and the massive sulfide disappears in higher levels beneath the gabbro. In transverse sections (east-west), through the North massive sulfide ore body, the gabbro is 500–600 feet west of the massive sulfide body. A cross section bearing N. 65° E. as shown on plate 7 illustrates the overhanging relationship of the gabbro.

The distribution of the Grapevine Gulch formation north and south of the massive sulfide lenses indicates that the massive sulfide has largely replaced the tuffaceous sediments of the Grapevine Gulch formation, a conclusion confirmed in some stopes where unreplaced sedimentary rocks remain within the massive sulfide, and small drag folds are partly replaced and give a banded structure to the sulfide. Some of the contorted banding within the sulfide bodies may reflect folded tuffaceous sedimentary rock.

The highly foliated zone encloses the body of massive sulfide which dips steeply east parallel to the massive sulfide. This zone appears to be the northern extension of the strong zone of foliation that intersects the main sulfide pipe to the south (fig. 25), and caused the southern extension of massive sulfide and black schist in the lower part of the main pipe. The gabbro is younger than the zone of strong foliation and nonfoliated gabbro separates these two mineralized zones. The North ore body of massive sulfide is overlain by the gabbro that possibly served as a barrier, and the rising mineralizing solutions replaced the highly foliated rocks.

The lenses of massive sulfide strike more westerly along their south margin near the gabbro intrusive mass, indicating a drag pattern against the gabbro. In part the northern boundary of the gabbro is a fault (pl. 7), and the bending of the structure may be a drag structure. The possibility, however, cannot be denied that the gabbro body intruded across the strongly northwestward-trending foliated zone, and pushed the foliated zone eastward to produce the drag pattern.

MINERALOGY

The mineralogy of the North massive sulfide body is similar to that of the main sulfide pipe. In the North ore body along the west margin fine-grained quartz forms scattered lenses as much as 20 feet in width, concordant with the structure. Pyritic massive sulfide and zinc-bearing massive sulfide occur east of the quartz, and veins of pyrite in the quartz indicate that the quartz is older. The massive sulfide is in part nonbanded, particularly in the pyritic facies, but is banded where sphalerite is abundant. Quartz-carbonate in veins and nodules, with or without sulfide minerals, are common, particularly on the east side of the mineralized zone.

The nonbanded massive sulfide is similar to the pyritic facies of the main pipe in that small aggregates of pyrite are separated by a gangue of quartz and carbonate minerals. Arsenopyrite is a minor constituent; it occurs in diamond-shaped crystals, partly embedded in the gangue. Sphalerite is interstitial to the pyrite, and their age relationship is uncertain. Carbonate veins cut the sulfide minerals.

The banded ore is rich in clear sphalerite (light-brown) in thin section that veins the quartz or is interstitial to the quartz grains. Much of the sphalerite averages about 0.2 millimeter across. The pyrite forms cubes ranging from 0.05 to 0.04 millimeter across and averages about 0.02 millimeter. Aggregates of pyrite cubes form crude veinlets in the ore. Galena and chalcopyrite are common in isolated crystals in the quartz gangue. Some of the quartz is very fine grained, and is cut by coarser grained quartz veinlets; other quartz forms columns at right angles to pyrite crystal faces, indicating pressure shadows and a growth later than the pyrite.

Chlorite in some of the massive sulfide ore, forms clusters or subparallel bands parallel to the banded pyrite and sphalerite. The chlorite in part is bright green in thin section; it has a very low birefringence, an index of refraction of about 1.635, and negative optic angle. Inclusions of tuffaceous sedimentary rock in the massive sulfide are heavily chloritized, and the chlorite (ripidolite) has the same optical properties as the chlorite in the ore except for a positive optic angle. Chlorite to

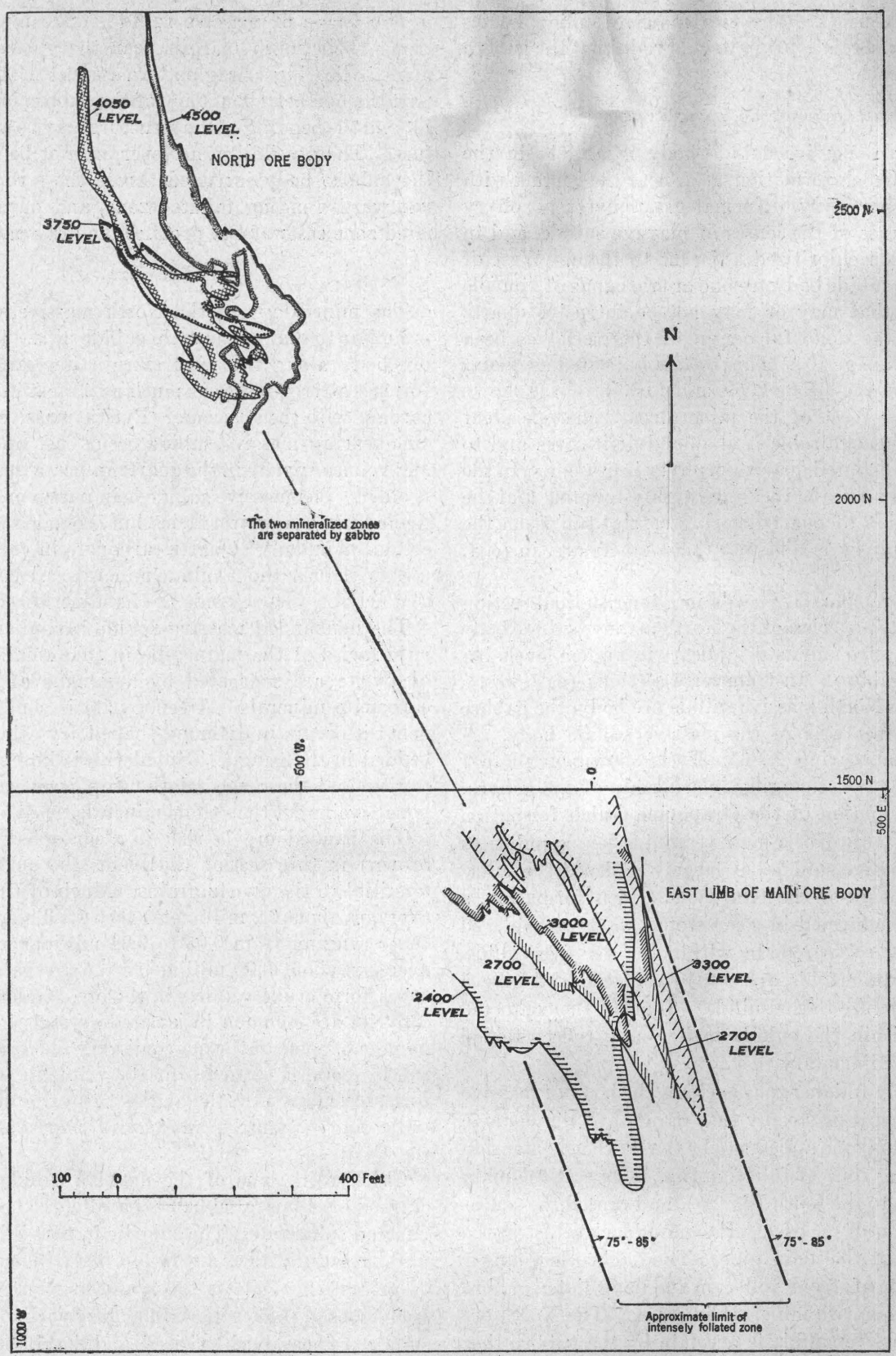


FIGURE 25.—Map showing relationship of zone of intense foliation in main ore body to North ore body, United Verde mine.

the east of the massive sulfide body is like most of the chlorite (prochlorite) south of the main pipe, in that the birefringence is moderate, index of refraction is about 1.610, and the optic angle is positive.

Chalcopyrite and pyrite veins in the black schist in places form minable ore on the east side of the massive sulfide body. This black schist ore is similar to that from the main ore body and does not warrant separate description.

ORE SHOOT

Where the massive sulfide lenses widen appreciably, they are richer in copper and zinc, and one ore shoot was estimated at 85,000 tons of ore containing 4 percent copper and 9 percent zinc (Little, 1950). The minable ore was 180 feet long and ranged from 10 to 60 feet in width. The zinc content was high enough to warrant flotation separation in the concentrator at Clarkdale.

SUPERGENE CHANGES

The oxidized and secondary enrichment zones of the United Verde deposit were removed in the open-pit operations, but Reber (1938) has recorded the pertinent data and the following information has been abstracted in part from his paper.

OXIDIZED ZONE

The gossan of the ore zone formed a blanket about 100 feet in average thickness, penetrating slightly below the 160 level of the mine. The gossan consisted largely of highly colored, soft limonitic material containing lenses and boulders of hard iron oxide. At the southern margin of the gossan, copper carbonate minerals were conspicuous where the mineralized black schist interfingers with the quartz porphyry. Along the northern margin, the primary quartz cropped out against the gabbro. South of the massive quartz and to some extent along the east side of the gossan, brecciated, honeycombed quartz was exposed.

High-grade gold-silver ore was mined from parts of the soft gossan, and high concentrations of native silver were found at several places immediately overlying the sulfide minerals. Almost all of the soft gossan was of commercial grade in the open-pit operation, and portions of the low-grade gold-silver ore were high enough in silica for converter flux.

The following analysis is of oxide ore shipped during 1918 (Reber, 1938, p. 50):

	Percent
Cu	1.42
Fe	31.5
Zn	.2
SiO_2	34.1
Al_2O_3	5.7
S	4.2

SECONDARY ENRICHMENT

In places, unenriched massive pyrite was just below the oxidized ore, but Reber believed that chalcocite enrichment affected the chalcopyrite ore as deep as the 500 level, and possibly much of the highest grade ore on the 300 level was chalcocite. Detailed records of the early mining are lacking, and during the open-pit operations most of the highest grade pillar ore was found as crushed and broken fragments in places mixed with old stope fill. Much chalcocite and considerable bornite were present in the most crushed material from the pillars under the 300 and less extensively under the 400 levels.

Bornite and steely chalcocite were found as lumps and boulders in loose material that showed evidence of intense fire action. Thirty years elapsed from the time of the first mine fire to the opening of the pit; and during much of this time, the material was extremely hot. Boyd [7] concluded that all or nearly all the bornite and probably much of the chalcocite were formed by fire action. Boyd produced bornite by maintaining chalcopyrite at a temperature of 500°C in a reducing atmosphere for 4 to 5 hours. By additional heating, some bornite changed to chalcocite. There is a possibility therefore that much of the chalcocite in the upper levels of the mine was not due to secondary enrichment.

A disseminated ore body in the quartz porphyry, south of the main ore zone and about 500 feet in vertical dimension, contained about 1,000,000 tons of 1.5 percent copper. This ore body was formed by secondary enrichment of very lean disseminated pyrite. The scant leached capping, and minor oxidation throughout the ore body indicate that this chalcocite enrichment was not related to the present erosion cycle.

The distribution and amount of gold and silver in the open pit compared to that in primary ore, led to the conclusion that gold and silver were enriched. Possibly the precious-metal enrichment had some relationship to the fire-zone conditions.

UNITED VERDE EXTENSION MINE

No detailed studies were made by the Geological Survey of the United Verde Extension mine. During the fall of 1947, this mine was opened by lessees to the 1,300 level, which is drained by the Josephine haulage tunnel. The levels below the 1,300 level were not unwatered in 1947. G. W. H. Norman of the Mingus Mountain Mining Corp. prepared geologic maps of the accessible underground workings, and compiled geologic maps of the inaccessible workings, using older reports, hand specimens, and diamond-drill cores on file for guid-

[7] L. M. Boyd, 1935, Microscopic examination of certain ores from the United Verde fire stopes: Unpublished thesis, Colo. School of Mines.

ance. Anderson of the Geological Survey and Dr. Norman spent several days observing critical exposures of the underground workings. Through the kindness of Arthur Notman, president of the Mingus Mountain Mining Corp., the maps prepared by Dr. Norman were made available for study, and some of these are included in this report. Notman also permitted Dr. Norman to help prepare the section on the geology of the United Verde Extension mine and the relationship of the United Verde and United Verde Extension ore bodies.

Much of the information on the ore bodies within the United Verde Extension mine has been obtained from the reports by Lindgren (1926) and Reber (1938). The data on production and dividends were taken from annual reports to stockholders.

PRODUCTION AND DIVIDENDS

The United Verde Extension mine produced 3,878,825 dry tons of ore which yielded 793,331,100 pounds of copper, 6,449,156 ounces of silver, and 152,756 ounces of gold. Production started in 1915 with 9,275 tons of ore, which yielded 5,138,989 pounds of copper, 19,393 ounces of silver, and 183 ounces of gold. The following year, 1916, production increased to 77,461 tons of ore, producing 36,402,972 pounds of copper, 128,467 ounces of silver, and 2,570 ounces of gold.

Peak annual production of the United Verde Extension mine in terms of copper, was reached in 1917 when 63,879,506 pounds of copper were obtained from 115,064 tons of ore. The ore averaged about 27.5 percent copper in contrast to the average grade of 23.5 percent copper in 1916. The grade of the copper ore decreased after 1917, dropping to 15.5 percent in 1919, and 12.8 percent in 1920. Subsequent lowering in grade was more gradual, and in 1929, when the peak annual tonnage from the mine was reached, 358,654 tons, the average grade was 8.6 percent copper. From 1933 to 1938 when the mine was closed, the average grade of the ore ranged from 6 to 7 percent copper. During this period, smelting ore was selected, averaging better than 7 percent copper, but the ore sent to the concentrator ranged in copper content from 4.02 percent in 1934 to 1.48 percent in 1937.

The gold content of the United Verde Extension ore averages about 0.040 ounce to the ton, although the first 5 years of production was under the average, dropping to 0.0144 ounce to the ton in 1917. In 1933, the gold content averaged 0.055 ounce to the ton from a mine production of 241,555 tons. In 1938, the last year of operation, only 18,196 tons were mined and the gold content averaged 1.08 ounces to the ton. This high gold content resulted from mining ore from the "gold stope," a small oxidized ore body between the 950 and 1,100 levels (pl. 10).

During most of the production from the mine, the silver averaged less than 2 ounces to the ton. The highest yearly average was attained in 1918 with an average grade of 4.18 ounces to the ton. In 1934, the average grade was 1.33 ounces to the ton, the lowest yearly average in the history of the mine.

The United Verde Extension Mining Co. authorized 1,500,000 shares of stock, of which 450,000 shares were not issued, leaving 1,050,000 shares of issued stock with a par value of 50 cents per share. The company started paying dividends in 1916 and their record is given in table 23.

TABLE 23.—*Dividend record, United Verde Extension Mining Co.*

Year	Value	Year	Value
1916	$1, 050, 000	1928	$2, 100, 000
1917	2, 992, 500	1929	3, 939, 500
1918	4, 725, 000	1930	3, 150, 000
1919	2, 362, 500	1931	1, 575, 000
1920	2, 100, 000	1932	603, 750
1921	1, 050, 000	1933	420, 000
1922	1, 312, 500	1934	1, 155, 000
1923	3, 675, 000	1935	2, 467, 500
1924	2, 625, 000	1936	787, 500
1925	2, 362, 500	1937	1, 050, 000
1926	3, 150, 000	1938	2, 730, 000
1927	3, 150, 000	Total	50, 531, 250

The mine was closed in May 1938, and the dividends paid in that year represented the final liquidation of the property. The total sum paid out to stockholders, $50,531,250, represented $48.125 per share of issued stock.

DEVELOPMENT

Two vertical concrete three-compartment shafts, the Edith and Audrey, are 200 feet apart and 1,000 feet north of the main ore zone. These shafts were used to transport men and supplies, and in the early stages of mining, to hoist ore. Later, a haulage adit, the Josephine tunnel, 2¼ miles long, was driven and connected with the Edith and Audrey shafts on the 1,300-foot level. Ore was hoisted in the Audrey shaft from lower levels to the 1,300-foot level. The Josephine tunnel is 10 by 10 feet in cross section and contained a standard-gage railroad track that connected with the smelter at Clemenceau, 5 miles from the portal of the adit.

Levels were run on the 550-, 800-, 950-, and 1,100-foot elevations, and on 100-foot spacings from the 1,100 to the 1,900 levels inclusive. The elevations refer to the collar of the Daisy shaft, the first exploratory shaft on the property. The elevations below the Edith and Audrey shafts are about 200 feet less.

Production from the main ore body of chalcocite ore came from the 1,300, 1,400, 1,500, and 1,600 levels. Mixed chalcocite-cuprite-malachite ore south of the main ore body was mined from the 800, 950, 1,100, and 1,200 levels. North of the main ore body mixed chalcocite-cuprite-native copper ore was mined from the 700, 800, 950, 1,100, 1,200, 1,300, 1,400, 1,500, and 1,600 levels. The northern ore body of fluxing primary sulfide ore was mined from the 1,500, 1,600, 1,700, 1,800, and 1,900 levels. Secondary ore containing malachite and chrysocolla in the gravel of the Hickey formation was mined from the 300 to above the 700 levels. A small production of malachite and chrysocolla ore in limestone came from above the 550 level.

The main sulfide ore body was mined by square sets, tightly filled with waste, and almost complete extraction with little dilution was achieved (D'Arcy, 1930).

GEOLOGY

By G. W. H. Norman, C. A. Anderson, and S. C. Creasey

OLDER PRECAMBRIAN ROCKS

Paleozoic sedimentary rocks and Tertiary gravel and lava flows cover Precambrian rocks east of the Verde fault in the underground workings of the United Verde Extension mine.

The Deception rhyolite is the main unit of the Ash Creek group exposed in the mine, although a narrow screen of fissile schist separating the Deception rhyolite and gabbro may represent metamorphosed sedimentary tuff of the Grapevine Gulch formation. Quartz porphyry including a spherulitic facies and gabbro are associated intrusive rocks.

Deception rhyolite

The Deception rhyolite exposed in the United Verde Extension mine consists of two facies. The western one is a light-gray to white massive rhyolite like much of the Deception rhyolite exposed in the Deception Gulch except that intense hydrothermal alteration is lacking. Some interbedded tuffaceous sedimentary rocks were noted on some levels. The eastern facies consists of a rhyolitic fragmental rock containing alternating beds of coarse breccia, fine breccia, and coarse tuff. The fragments are dominantly pink rhyolite containing sparse quartz and feldspar phenocrysts in a microcrystalline groundmass.

These two facies of rhyolite are separated by andesitic flows and intercalcated tuffaceous sediments. Locally the andesite is amygdaloidal and has faint fragmental structures. Much of the andesite is massive, and it may possibly be an intrusive mass. No chilled selvages, however, were observed and the andesitic rocks appear generally similar to some of the flows and fragmental rocks interbedded with the Deception rhyolite in the Mescal Gulch area.

The total thickness of the Deception rhyolite and andesitic member on the 1,400 level northwest of the Audrey and Edith shafts (pl. 5) is probably about 750 feet, of which approximately 450 feet is the eastern rhyolitic breccia. On other levels the thickness of the rhyolite breccia also varies: on the 1,100 level near the Audrey shaft, it is only 90 feet, and on the 1,200 level, it is 140 feet. The andesitic member is about 200 feet thick on the 1,200 level, but elsewhere in the mine is thinner. The western exposures of massive rhyolite are about 100 feet thick.

Foliation is intense along the west margin of the Deception rhyolite, but decreases in intensity to the east, where relict textures and structures are easily recognized. The andesitic flows show little orientation of secondary minerals in thin section, and a pilotaxitic arrangement of albitic plagioclase is separated by bright green chlorite and magnetite (or ilmenite). Some thin sections of the andesite show considerable epidote and calcite, masking the original igneous textures.

Grapevine Gulch formation

A band of fissile greenish-gray and grayish-white schist is along the east margin of the gabbro mass, west of the Deception rhyolite. The rock is so intensely foliated that all distinct bedding structures have been destroyed. The interpretation that the original rock was a fine-grained clastic rock is based on fine color-banding similar to that in bedded rocks, and on the fissility of the schist and general similarity to highly schistose units of the Grapevine Gulch formation at outcrops and in the United Verde mine. Evidence of deformation after the formation of the cleavage is shown by contorted foliation associated with quartz veins and silicified zones.

The band of fissile schist may be a thick unit of tuffaceous sedimentary rock interbedded in the Deception rhyolite, or it may be a part of the Grapevine Gulch formation. In the United Verde mine the Grapevine Gulch formation is in contact with the southeast margin of the gabbro (pl. 5), and presumably this formation would continue around the eastern lobe of the gabbro that is downfaulted in the United Verde Extension mine. The geologic maps of the lower levels, such as the 1,700 level (pl. 10) show that the fissile schist extends to the Verde fault, where it could be the faulted extension of the Grapevine Gulch formation from the footwall side. For this reason, the fissile schist is designated Grapevine Gulch formation.

Quartz porphyry

Two large masses of quartz porphyry are in the United Verde Extension mine. One is south of the gabbro and east of the Verde fault; this mass is host rock for much of the massive sulfide ore in the mine. The other large mass is exposed east of the deposit in the Josephine tunnel and on the surface. The surface exposure (pl. 1, 1,368,000 N.; 444,500 E.), has a north-south length of 500 feet or more.

Smaller masses of quartz porphyry are in the gabbro mass east of the Audrey shaft; presumably these masses represent pendants and inclusions of quartz porphyry in the younger gabbro. Dikes, in part parallel to the structure of the Deception rhyolite, intrude the rhyolitic and andesitic members. Many of these dikes have a spherulitic structure; the spherulites now consist of radiating quartz fibers. Some dikes of the spherulitic facies do not contain the conspicuous quartz phenocrysts so characteristic of most of the quartz porphyry.

Gabbro (United Verde diorite of Reber, 1922)

The mass of gabbro north of the ore body and west of the Audrey shaft is cut off by the Verde fault, and presumably this mass is the downfaulted segment of the gabbro exposed at the United Verde mine (United Verde diorite of Reber, 1922). The gabbro in the United Verde Extension mine is more foliated than in the United Verde mine, and on some levels, the foliation is limited to the margin for a width of only 5–10 feet. On other levels, the foliation is much more extensive.

A large mass of gabbro is present east of the Audrey shaft (pl. 5), and a small tongue is exposed in the Josephine tunnel. To the north, its extent is further defined by diamond-drill holes and the A & A workings (pl. 1, 1,370,000 N.; 437,900 E.). This eastern gabbro has a north-south length of 6,000–7,000 feet and a minimum width of 1,600 feet. West of the Verde fault (pl. 5, 3,000 E.; 0 NS) a small body of gabbro, found by underground exploration, was termed the footwall gabbro. Presumably this mass is a faulted segment of the gabbro exposed in the eastern part of the United Verde Extension mine.

Petrographically, the footwall gabbro is similar to the eastern gabbro and both have been altered to the same degree as the gabbro in the United Verde mine. The plagioclase is intensely altered to albite, sericite, epidote, and clinozoisite, and the pyroxene is largely altered to greenish hornblende and chlorite. Relict pyroxene showing an ophitic relationship to plagioclase was noted in one thin section. Bright green chlorite, polarizing to bronzy colors, may be pseudomorphic after biotite. Interstitial quartz is common, and in part, it may represent primary quartz as well as quartz formed by alteration. Leucoxene has replaced primary ilmenite(?) and apatite prisms are common. In some thin sections, carbonate minerals (calcite, dolomite, or ankerite) are present. Possibly these mafic rocks in the United Verde Extension mine were originally dioritic, as indicated by possible original quartz and biotite. The degree of alteration prohibits certainty on this point.

Paleozoic and Tertiary Rocks

The Tapeats sandstone(?) is exposed in the 700 and 800 levels of the United Verde Extension mine, where it ranges from 20 to 40 feet in thickness. The dip is about 5° E. and on the 800 level (pl. 10), Tapeats sandstone(?) is exposed at the station of the Edith shaft. The Tapeats sandstone(?) does not extend westward to the Verde fault for it is cut out by late Tertiary gravel of the Hickey formation (section, pl. 10).

The Devonian Martin formation is exposed on the 800-foot level at the station of the Audrey shaft, and an exploratory drift to the south cuts through it. The Martin is cut out by the gravel of the Hickey formation to the west.

The Mississippian Redwall limestone is exposed at the surface of the United Verde Extension property, but not underground, except in the Josephine tunnel east of the Bessie fault.

The late Tertiary Hickey formation overlies the Precambrian and Paleozoic rocks at the United Verde Extension mine. Basaltic flows crop out at the surface overlying a channel containing about 350 feet of gravel (fig. 26). The gravel unit of the Hickey formation rests on Redwall limestone, Martin limestone, Tapeats sandstone(?), and the Precambrian rocks near the Verde fault. To the west, the gravel and basalt are in fault contact with the Precambrian rocks on the footwall side of the Verde fault.

STRUCTURE

Folds

The regional evidence indicates that the United Verde Extension mine is located in the northward-plunging Mingus anticline (discussed earlier on p. 65). The evidence is less certain as to the location of the mine in relation to the axial plane of the Mingus anticline. Two possibilities should be considered. First, the Deception rhyolite and andesitic unit in the mine are on the eastern limb of the Mingus anticline and in fault contact to the west with westward-facing Deception rhyolite and Grapevine Gulch formation. Second, the Deception rhyolite-Grapevine Gulch formation contact is the downfaulted normal contact observed in the pit at the United Verde mine (discussed on p. 106), and the Deception rhyolite in the United Verde Extension mine is on the western limb of the Mingus anticline.

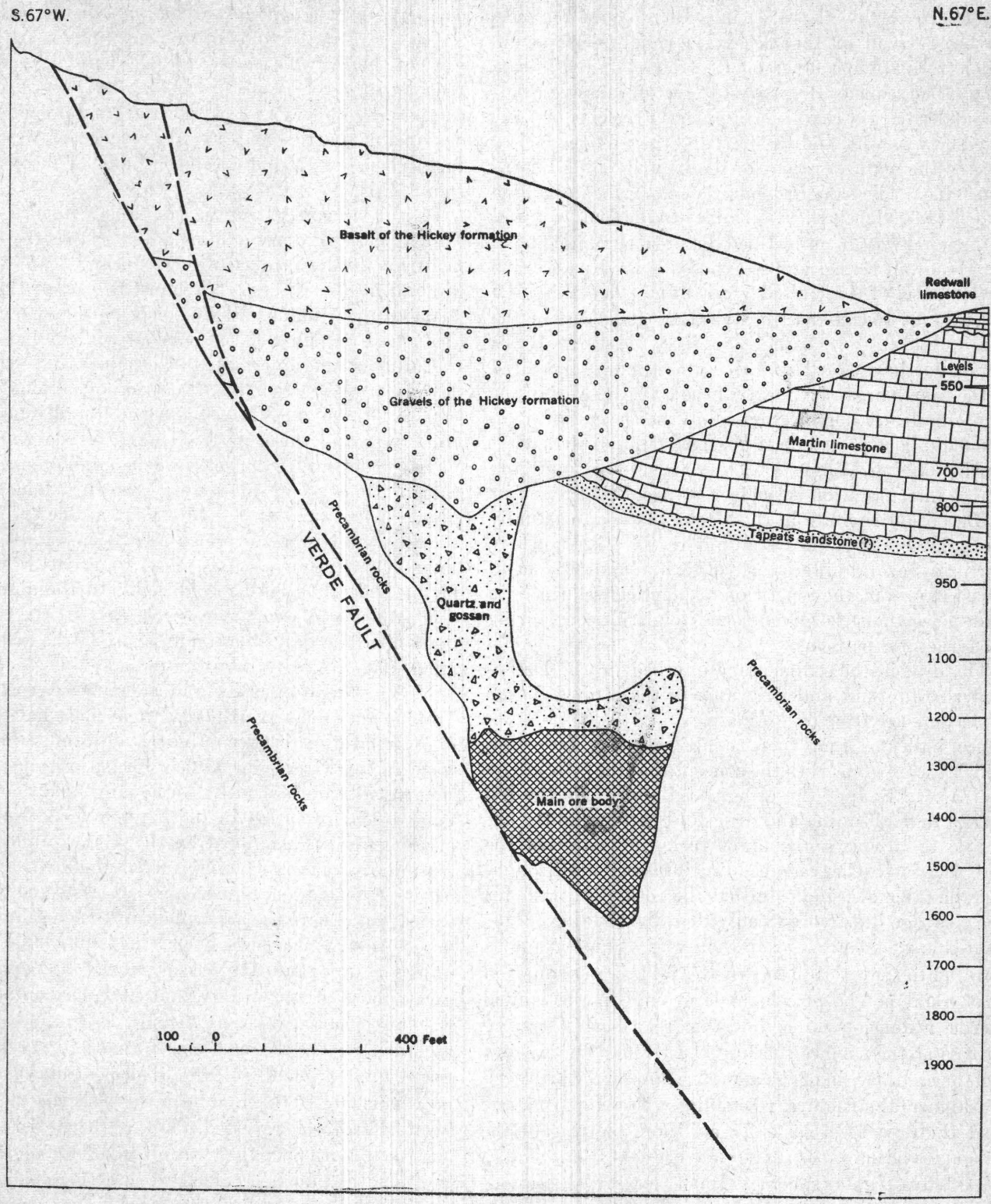


FIGURE 26.—Generalized section through United Verde Extension mine. From Reber (1938).

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The possibility that the Deception rhyolite is on the eastern limb of the Mingus anticline will be considered first. In the United Verde Extension mine, the trend of the Deception rhyolite including the andesitic member is north to northwest (pl. 5) in contrast to the regional northeast trend of the Deception rhyolite-Grapevine Gulch formation contact south of the United Verde mine. The sequence from west to east of massive rhyolite, andesite, and rhyolitic fragmental rocks is best exposed between the Audrey shaft and the western gabbro mass in the main workings of the United Verde Extension mine. The same sequence is cut by the Josephine tunnel (1,300 level, U. V. X. mine) to the southeast and to the northwest in an inclined diamond-drill hole from the Hopewell tunnel and is exposed near the portal to the 500 level of the United Verde mine (pl. 5, 2,400 N.; 150 E.). A line connecting the three areas east of the Verde fault, disregarding the later intrusive rocks, trends northwestward, and the same trend is indicated in the fourth area, the portal of the 500 level west of the Verde fault. A northwest trend of the Deception rhyolite could be expected on the eastern flank of the northward plunging Mingus anticline. The chief objection to this interpretation is the lack of proof that the units of Deception rhyolite in these four places are actually the same, because the Deception rhyolite is known to consist of several different rhyolitic and andesitic members. However, the repetition of identical sequences in four places must be given some weight in considering the structural pattern.

The dips in the sedimentary interbeds in the Deception rhyolite and andesitic unit in the United Verde Extension are from 60° to 80° E.; and the available evidence indicates that these same beds also face east, which is consistent with the concept that the beds are on the eastern limb of the Mingus anticline. On the 950 level, graded bedding and drag-fold patterns in the andesitic tuffaceous sedimentary rocks indicate that the beds face east. Cleavage-bedding intersections throughout the mine also indicate that the rocks face east, for the cleavage dips more steeply than the bedding. This particular evidence is valid only if the cleavage is parallel to the axial planes of the folds, an assumption that could not be proven underground in the United Verde Extension mine, but this relationship can be proven at the surface south of the United Verde mine. On the east margin of the andesitic flow on the 800 level, a fragmental structure, resembling a flow top, suggests that the flow faces east. In addition, plotting of the average lengths of the plagioclase crystals from samples collected at regular spacings from two east-west sections through the flow on the 800 level reveals an asymmetric curve and the steep part of the curve is to the west. This distribution of coarser feldspar to the west indicates that the base of the flow is in that direction.

The absolute proof of the direction that rocks face in a metamorphosed terrane is difficult to determine without persistent recognition of the tops of the beds, and the noses of major folds of known plunge. Although none of the features observed in the United Verde Extension mine constitute proof that the section of Deception rhyolite faces east, the sum of the evidence is convincing, particularly as no evidence of westward-facing beds was found except on limbs of recognizable small folds.

If it is accepted that the available evidence in the United Verde Extension mine indicates that the Deception rhyolite has a northwest trend and that the beds dip and face east, these rocks must lie east of the axial plane of the Mingus anticline or be on the eastern limb of a small anticline, to the west of the axial plane of the Mingus anticline. Small anticlines and synclines have been recognized (see discussion of regional structure p. 69) on the western flank of the Mingus anticline west and south of the United Verde Extension mine. However, if the rocks in the mine represent the eastern limb of a small anticline, this fold should plunge northward at 50°–60°—comparable to folds in the adjacent Grapevine Gulch formation—and the beds should curve around it to the west, and some westward-facing beds should be recognized. In the absence of this evidence, it seems more probable that the rocks in the United Verde Extension mine are to the east of the axial plane of the Mingus anticline.

If the Deception rhyolite in the mine lies east of the Mingus anticlinal axial plane, there must have been a fault separation between this rhyolite and the westward-facing Grapevine Gulch formation and Deception rhyolite in the United Verde mine, and the Grapevine Gulch formation in the United Verde Extension mine, before displacement on the Verde fault and intrusion of quartz porphyry. This suggested fault (pl. 5) is indicated on figure 3 as an eastward-dipping thrust, parallel to the axial plane of the anticline. Appreciable displacement on such a fault would be required in order to bring Deception rhyolite, east of the axial plane of the Mingus anticline, in contact with the uppermost westward-facing Deception rhyolite and younger Grapevine Gulch formation. If this proposed fault dipped west, the displacement would have been normal; if the fault dipped east, the movement would have been reversed. Because there is evidence that the quartz porphyry was intruded during the folding, and the quartz porphyry intrusive bodies show no sign of large-scale faulting, displacement along this proposed fault must have occurred before or during

the folding. There is no evidence of a folded fault. Because the eastward-dipping foliation in this immediate area is essentially parallel to the axial planes of the minor folds, it seems logical to suggest that a fault formed during the compression would be essentially parallel to the axial plane of the folds—that is, dip to the east, and be a thrust fault (fig. 3).

The concept of a faulted anticline requires that some attention be paid (1) to the outcrops of Deception rhyolite west of the Verde fault and north of the Haynes fault, and (2) to the exposures of the Grapevine Gulch formation in the United Verde Extension mine. At the first locality, the rhyolite and andesite trend northwest, similar to that in the United Verde Extension, and the sequence is similar—from west to east, massive rhyolite, andesite, and rhyolitic fragmental rocks; the two outcrops possibly were connected before the displacement along the Verde fault. At the second locality, Grapevine Gulch formation west of Deception rhyolite in an eastward-facing section is stratigraphically out of place. The proposed thrust will explain the relationship at both localities by separating westward-facing Grapevine Gulch formation on the west from eastward-facing Deception rhyolite on the east (pl. 5).

Although a comparable thrust fault parallel to the axial planes of the major folds has not been found elsewhere in the Ash Creek group, it must be admitted that deformation is more intense in the vicinity of the United Verde and United Verde Extension mines as shown by the intense foliation in the quartz porphyry, Deception rhyolite, and Grapevine Gulch formation.

The second possibility, which is perhaps a simpler explanation, is that the Deception rhyolite in the United Verde Extension mine is on the western limb of the Mingus anticline. To visualize this structure, it must be assumed that the rocks moved to their approximate relative position before displacement on the Verde fault. This interpretation is most easily understood by matching the faulted lobe-shaped mass of gabbro exposed on the 1,400 level of the United Verde Extension mine to the parent mass of gabbro at the surface of the United Verde mine. (Pl. 5, visually place contact at 100 N.; 3,420 E. to 1,100 N.; 900 E.) Then the Deception rhyolite-Grapevine Gulch formation contact exposed east of the United Verde mine would sweep around the east margin of the gabbro, and connect with the contact west of the Deception rhyolite buried beneath the mine dump north of the Haynes and west of the Verde fault (pl. 5, 2,400 N.; 200 W.). Information obtained in the United Verde mine indicates that the Deception rhyolite-Grapevine Gulch formation contact is overturned and dips to the east, south of the gabbro mass. The east dips of the Deception rhyolite in the United Verde Extension mine may then represent local overturning of the westward-facing limb of the Mingus anticline, as reflected by the eastward bulge of the Deception rhyolite-Grapevine Gulch formation contact along the east margin of the gabbro. This second explanation demands that the evidence be ignored that indicates that the Deception rhyolite faces east in the United Verde Extension mine.

The choice of interpretation depends therefore upon the reliance to be placed upon the evidence of eastward-facing beds in the United Verde Extension mine. The faulted anticlinal interpretation could only be proved satisfactorily if the cover of younger rocks was removed and detailed mapping of Precambrian rocks revealed truncation of members in the Deception rhyolite by the contact of Grapevine Gulch formation to the west or vice versa. We believe that the evidence is sufficiently good to favor the eastward-facing Deception rhyolite in the United Verde Extension mine, so that the faulted Mingus anticlinal interpretation is our preference; but the skeptic is justified on the evidence available to prefer the second interpretation, namely, that the Deception rhyolite-Grapevine Gulch formation contact is the normal depositional contact. In either interpretation, the contact has been destroyed locally by intrusive quartz porphyry and locally bulged eastward by the intrusive gabbro.

Faults

Near the United Verde and United Verde Extension mines, the Verde fault is complex; the fault zone being about 100 feet wide. The main footwall break contains heavy gouge, and much of the movement may have been along this break (Reber, 1938, p. 51). Some movement is distributed over a braided system of small hanging-wall breaks that are difficult to map accurately. The fault dips east at 60° and the eastern block has been dropped. The only accurate measurement of the vertical separation is obtained by displaced Paleozoic rocks; it is about 1,500 feet. Against the footwall side of the fault, between the footwall and hanging-wall strands (pl. 5, 1,300 N.; 850 E.), displaced Tapeats sandstone(?) crops out and indicates only 690 feet of vertical separation (fig. 27) of this block of sandstone caught in the fault zone. The late Tertiary basaltic flows and gravels of the Hickey formation are displaced by the Verde fault (fig. 26), proving that much of the displacement took place after their accumulation, dating the fault as probably late Tertiary.

Precambrian movement along the Verde fault, following the period of metallization and some chalcocite enrichment, has been suggested by Ransome (1932) and Reber (1938, p. 51), but this phase of the history of

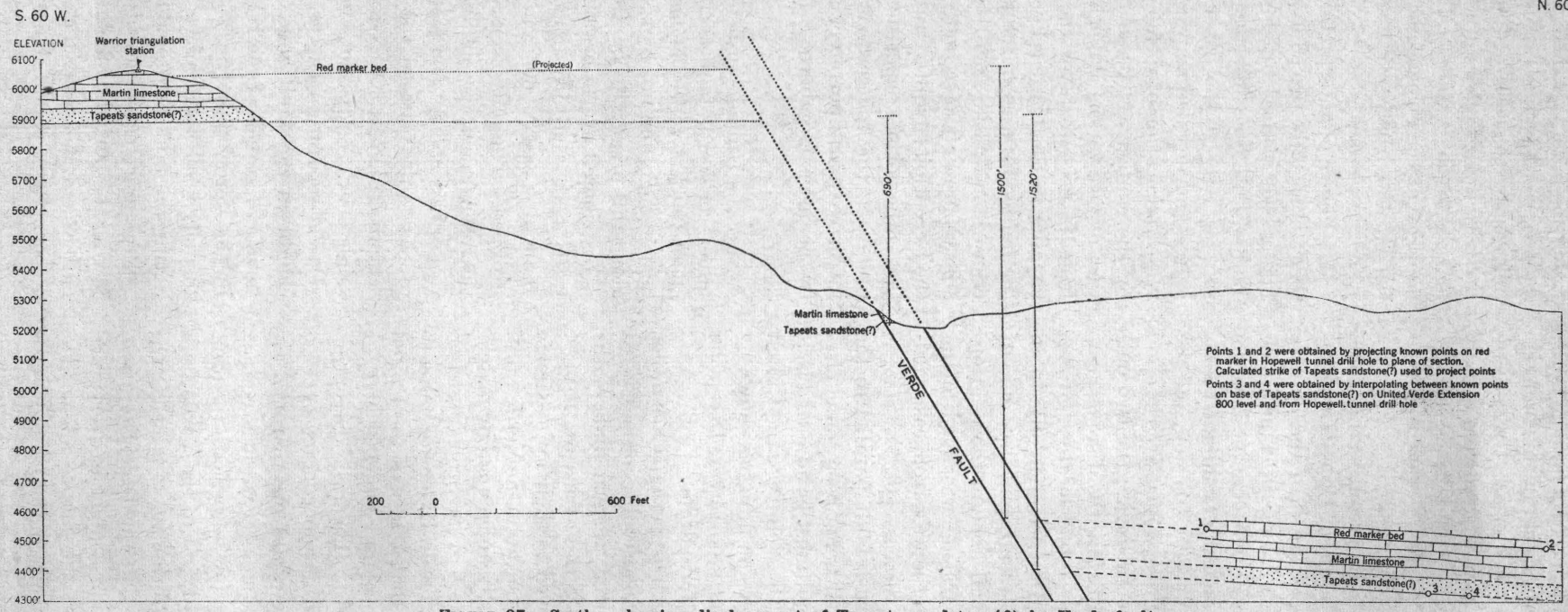

FIGURE 27.—Section showing displacement of Tapeats sandstone(?) by Verde fault.

the Verde fault will be discussed in the next section of this report.

The Florencia fault trends east-northeast and dips 70°–75° S. To the west, it is cut off by the Verde fault (pl. 5). On the 800 level (pl. 10) the Florencia intersects the Verde south of the gabbro, and cuts through the ore zone. On this level, the distribution of the Tapeats sandstone(?) shows that the south side dropped about 40–50 feet, proving that the fault is normal. On lower levels of the mine, the Florencia lies along the south side of the main ore body (see 1,300 level, pl. 10), as if the fault were older than the period of mineralization (which is in the Precambrian) and localized the ore body. The small displacement of the Tapeats sandstone(?) on the 800 level (pl. 10) may represent later adjustments along the fault in Tertiary time. No trace of the Florencia has been found on the footwall side (west) of the Verde. It is tempting to correlate the Florencia with the Hull because of similar attitudes. The probable bearing and plunge of the net slip on the Verde, however, is not permissive.

HYDROTHERMAL ALTERATION

The quartz-sericitic alteration common in the quartz porphyry and Deception rhyolite, in and south of the United Verde mine, is present only between the Verde fault and the ore body in the United Verde Extension mine. The volcanic rocks east of the ore body contain fresh feldspar, and relict textures and structures are recognized easily.

Black schist formed by chloritic alteration is predominant in quartz porphyry and Deception rhyolite south of the ore body in the United Verde Extension mine. According to Reber (1938, p. 52) minor quantities of schistose gabbro as well as quartz porphyry are replaced by black schist north of the ore body.

Jasper is common in the United Verde mine, and some of the hard massive quartz containing small chalcocite ore bodies in the northern part of the upper levels of the United Verde Extension ore body may be a product of similar silicification. The rocks of these upper levels have been appreciably oxidized and leached. How much of the silica remaining is primary silicification, and how much is residual from massive sulfide is unknown.

ORE BODIES

Form and distribution

The main chalcocite ore body is located about 900 feet south of the Edith and Audrey shafts (pl. 5). The top of the sulfide zone was located at the 1,240 level. Oxidized ore for 1 floor above, contained about 10 ounces of silver to the ton, whereas the oxidized zone above this floor was essentially barren. The shape of the main ore body in the upper levels is somewhat rounded to elliptical, with the long dimension east-west. On the 1,300 level (pl. 10), the main ore body was about 500 feet long and about 200 feet wide. On the 1,400 level, the length is 440 feet and the width is 270 feet (pl. 5). On this level, the southwest corner of the ore body is cut off by the footwall strand of the Verde fault, and a hanging-wall strand of the fault cuts through the ore body. Below the 1,400 level, the main ore body tapered; and on the 1,500 level, the southwest margin for 160 feet was bounded by the footwall strand of the Verde. On the 1,600 level, the ore body is oriented northwest; the length is 250 feet and the width ranges from 40 to 70 feet. The Verde is to the west of the ore body on this level, indicating the irregular bottoming of a sulfide lens. The main ore body bottomed above the 1,700 level and was mined to the 8th floor above this level.

On the 1,200 level an ore zone, from 5 to 20 feet wide and rich in native copper, extended for 450 feet north of the main ore body along the east margin of the gabbro. An ore body 120 feet long in this ore zone yielded the first significant tonnage of ore mined in 1915. This eastern ore zone was barren northward to the northeast margin of the gabbro, where a second ore body 350 feet long was found (pl. 10, 2,350 E.; 11,900 N.). This northern ore body was stoped from a few floors above the 1,400 level to 7 floors above the 1,100 level.

Above the 1,100 level, an upward lobe of gabbro is separated from the main mass of gabbro by older rock, and some oxidized ore is in the trough between gabbro (pl. 10, gold stope). On the 800 level (pl. 10), a northwestward-trending gossan zone is west of the gabbro and east of the Verde fault, and some oxide copper ore was stoped from above this level (section, pl. 10).

Small veinlike tongues of mineralized rock with northwesterly plunge extend south and east of the main ore body along shear zones in the quartz porphyry. About 400,000 tons of ore was produced from these tongues above and below the 1,100 level.

Northwest of the main ore body and adjacent to the Verde fault, massive unoxidized low-grade sulfide ore was found on the 1,200 level. This body is about 450 feet long and from 60 to 90 feet wide; the west margin is the footwall strand of the Verde and the east margin is one of the hanging-wall strands. This low-grade sulfide zone extended from the 1,300 to the 950 levels. On the east side, a chalcocite ore body 50 by 100 feet on the 1,200 level and extending from below the 1,300 (pl. 10) to above the 1,200 levels was found; it undoubtedly was "dragged ore" along the Verde fault.

On the 1,300 level (pl. 10) a tongue of chalcocite ore cuts the gabbro from the northwest corner of the

main ore body. The connection between this tongue and the main ore body is severed by one of the hanging-wall strands of the Verde fault between the 1,300 and 1,400 levels. This tongue of ore extended below the 1,500 level.

The northern ore bodies on the lower levels appear on the 1,400 level northwest of the main ore body, and are along the Verde fault. These bodies are mainly in contact with the footwall strand of the Verde (pl. 10), and may have been connected originally with the main ore body, but were severed from it by complex faulting. On the 1,900 level, only a mass 40 by 50 feet was found, 70 feet southeast of the footwall strand of the Verde (fig. 29). Thus neither the lowermost part of the main ore body nor the footwall-strand ore bodies bottomed on the footwall strand of the Verde, indicating an irregular bottoming of lenses of ore. These lower level ore bodies were mined for fluxing ore, and probably about half of these were stoped out. The lowest stope was just above the 1,900 level.

The form and trend of the ore bodies were controlled in part by the contact of quartz porphyry with Grapevine Gulch formation and Deception rhyolite, and in part by the margin of the gabbro (Reber, 1938, p. 50). The local schistosity near and in the gabbro trends east in contrast to the northwest trend in the rocks east of the gabbro. The main ore body was elongated in an easterly direction where it had its maximum size, possibly controlled in part by the Florencia fault; but at depth, the trend was northwest, parallel to the regional foliation.

In the lower levels of the mine, the southern contact of the gabbro dips north, and the deeper sulfide lenses generally are closer to the gabbro than in the upper levels. Some of the marginal schistose gabbro was mineralized like the Haynes ore body, in contrast to the nonmineralized marginal gabbro in the main United Verde mine. Reber (1938, p. 51) noted, however, that some of the apparent penetration of the gabbro by sulfide is partly due to faults.

Sulfide zone

The massive sulfide on the lower levels is similar to that in the United Verde mine. Pyritic massive sulfide contained minor quantities of sphalerite and chalcopyrite, represented by 2 to 7 percent zinc and from ½ to 1½ percent copper (Reber, 1938, p. 52). In places the pyritic massive sulfide is banded. Quartz-carbonate minerals separate the sulfide minerals on a microscopic scale, and veinlets of quartz-carbonate cut the massive sulfide.

Some of the smaller ore bodies on the lower levels contained chalcopyrite in similar proportions to the copper-bearing massive sulfide ore in the United Verde mine (Reber, 1938, p. 52).

The rich chalcocite ore that made the United Verde Extension mine famous consisted in part of massive fine-grained orthorhombic chalcocite containing a well-formed pisolitic structure formed by the replacement of pyrite by chalcocite (Schwartz, 1938, p. 26). In places, the chalcocite was sooty and soft (Lindgren, 1926, p. 86). Some unreplaced pyrite was present in the ore, but it was largely concealed by the chalcocite. Native copper and cuprite were abundant in some stopes on the east side of the gabbro, north of the main ore body, and in the southeastern tongues.

Chalcopyrite and sphalerite were absent in the chalcocite ore and must have been replaced by chalcocite. Lindgren (in a private report, quoted by Reber, 1938, p. 53) suggested that chalcopyrite was sparse in the primary ore, but Reber (1938, p. 53) believes that chalcopyrite replacement was most important in parts of the high-grade chalcocite.

The interpretation that the bonanza of chalcocite ore was formed by secondary enrichment has been widely accepted, and Reber (1938, p. 53) has succinctly reviewed the evidence:

> The decrease of chalcocite with depth, the general scarcity of chalcopyrite or sphalerite where chalcocite was most abundant, and the intense kaolinic alteration of the wall rocks, varying with the abundance of chalcocite, conclusively indicate formation by the process of secondary enrichment, which is also confirmed by microscopic evidence.

Enrichment occurred before deposition of the Tapeats sandstone(?) of Cambrian or possibly of Devonian age (see p. 48), for the Tapeats(?) and younger rocks covered the leached capping and gossan. Thus the enrichment is Precambrian or early Paleozoic, or both.

Oxide zone

The bottom of the oxide zone in the United Verde Extension mine is variable, due partly to faulting. It is deepest away from the Verde fault, reaching greater depths over the main ore body than over some of the smaller ones. It is possible that some oxidation occurred when the top of the ore zone was exposed in Tertiary time before the deposition of the gravel of the Hickey beds (section, pl. 10; fig. 26).

Above the 1,200 level, dense fine-grained primary quartz containing a little sulfide was more abundant than at depth. Within this quartz were several irregular lenses of high-grade quartz-chalcocite ore, some with small vertical extent. In the higher ore bodies, malachite with some cuprite was conspicuous (Reber, 1938, p. 53).

Silver enrichment was evident above the top of the massive sulfide; in the average chalcocite ore, the silver content was less than 2 ounces to the ton. The oxidized ore for about 7 feet above the chalcocite ore contained from 10 to 12 ounces of silver per ton, and locally, high-grade masses contained as much as 100 ounces per ton (Lindgren, 1926, p. 86).

The gossan or capping above the main ore body contains much silica, chiefly massive quartz that locally shows repeated brecciation and recementation. Cavernous hard quartz breccia containing appreciable limonite is more abundant than clean quartz. Soft limonitic material, similar to much of the United Verde gossan, occurs only locally, but predominantly close to the top of the sulfide zone. Native copper is abundant in places in the soft gossan. Some of the smaller siliceous ore bodies were capped with 40 to 50 feet of iron-rich gossan, but in a few places the chalcocite merged into massive quartz with no obvious leaching or slumping (Reber, 1938, p. 53–54). Chalcocite, showing only a small percent of malachite, occurred on and above the 800 level within 70 feet of the Tapeats sandstone (?).

The veinlike tongues of ore extending south and east from the main ore body were overlain by 50 to 100 feet of thoroughly leached, kaolinized rock containing some limonitic material. These ore bodies range vertically from near the 1,300 level to a few floors above the 800 level. Oxidized copper minerals were prevalent throughout and, in some of the higher parts of these ore bodies, accounted for more than half of the copper content and entirely masked the finely divided chalcocite. The most abundant oxidized copper minerals were malachite, chrysocolla, and azurite (Reber, 1938, p. 54).

Partly oxidized copper ore yielded nearly one-eighth of the production of the mine. Some ore mined from the gravel of the Tertiary Hickey formation was entirely in the form of oxidized copper minerals, and this ore amounted to only 2 percent of the production. The gravel ore may have formed by ground water carrying copper from the adjacent United Verde ore zone (Reber, 1938, p. 54) or be dragged ore along the Verde fault.

The "gold stope" ore body was a tabular veinlike body along the eastern gabbro contact in the upper levels of the mine that bottomed in a trough in the gabbro (section, pl. 10). The typical ore was fine-grained friable quartz containing almost no residual iron oxide. The maximum length was about 350 feet and the width ranged from 5 to 20 feet. This ore body extended about 200 feet vertically above and below the 950 level. The gold content was variable, and according to Reber (1938, p. 54), may have averaged $10 per ton, with some higher grade sections. Apparently, local conditions were exceptionally favorable to concentrate gold.

RELATIONSHIP OF THE UNITED VERDE AND UNITED VERDE EXTENSION ORE BODIES

By G. W. H. Norman, C. A. Anderson, and S. C. Creasey

Is the United Verde Extension ore body the down-faulted upper part of the United Verde ore body? Debate on this question has been active since the discovery of the United Verde Extension ore body. The simplest structural explanation is that movement on the Verde fault was dip slip, and the Paleozoic rocks on opposite sides of the fault provide the best measure of displacement. This vertical separation (throw) has been measured as about 1,500 feet (fig. 27), and the 1,400 level of the United Verde Extension mine is about 1,500 feet below the surface elevations of the United Verde mine. Reversing the Tertiary displacement of 1,500 feet at right angles to the trace of the Verde fault places the United Verde Extension ore body, as outlined on the 1,400 level, 2,700 feet east-southeast of the United Verde ore body as exposed at the surface (fig. 28, position I). This simple geometry was accepted by Provot (1916) and Rickard (1918, p. 51) as proof that the United Verde Extension ore body was not the faulted top of the United Verde ore body. This explanation overlooks the important fact that although the reversal of the dip-slip movement will result in matching the Paleozoic rocks, the Precambrian rocks on opposite sides of the Verde fault will not match (pl. 5).

The gabbro mass (United Verde diorite of Reber, 1922) north of the United Verde ore body is cut off by the Verde fault and similar gabbro is also cut off north of the United Verde Extension ore body. Matching these two gabbro masses on opposite sides of the fault can be done by reversing the known Tertiary vertical separation of 1,500 feet and moving the hanging-wall block horizontally 2,200 feet to the northwest. Matching the gabbro places the United Verde Extension ore body 1,000 feet northeast of the United Verde ore body (fig. 28, position II). Fearing (1926, p. 770) made this suggestion of pronounced lateral as well as vertical movement to match the gabbro and he concluded that this was proof that the two ore bodies were originally separate before displacement along the Verde fault.

This suggestion of pronounced lateral displacement has the virtue of matching the gabbro and Paleozoic rocks on opposite sides of the Verde fault, but it neglects the fact that the mass of Deception rhyolite on the surface east of shaft 7 (pl. 5; 200 N.; 1,500 E.) on the footwall side (west) is not represented on the 1,400-foot level of the United Verde Extension mine (1,500 feet vertically below the United Verde surface). In addition, at the surface, the quartz porphyry along the west

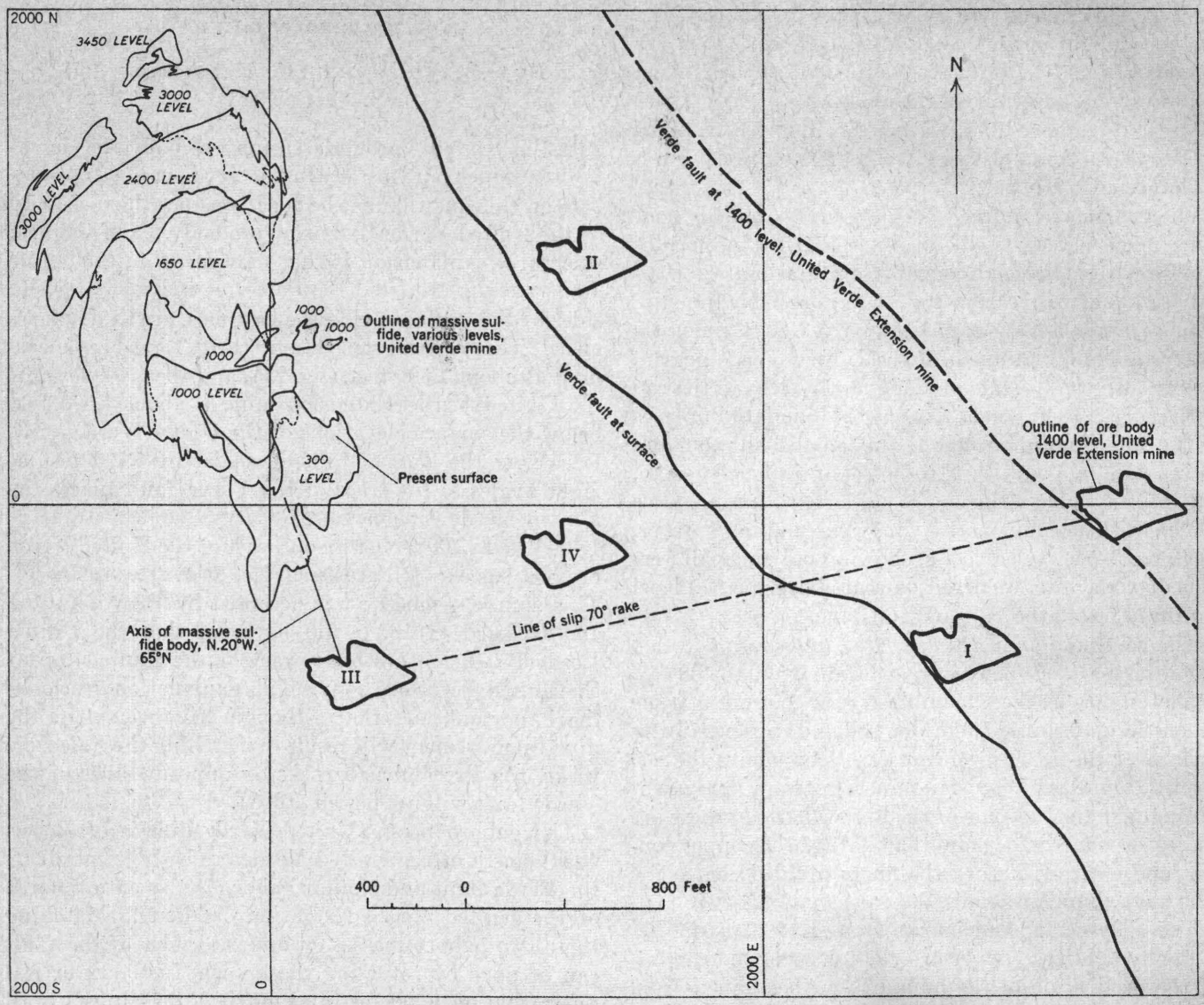

FIGURE 28.—Sketch map showing interpretations of relationship of United Verde and United Verde Extension ore bodies.

side of the Verde fault is exposed for 2,800 feet north of the Hull fault; and on the 1,400 level on the United Verde Extension mine, the quartz porphyry along the east side of the Verde fault is present for less than 1,800 feet north of the Deception rhyolite and south of the gabbro. This lack of coincidence of the Precambrian geology on opposite sides of the fault after reversing the 1,500-foot Tertiary displacement and including sufficient lateral movement to match the gabbro is reason enough to rule out this explanation.

Lindgren (1926, p. 87) was cautious in expressing any positive opinion as to the relationship of the two ore bodies. He did suggest that lateral as well as vertical movement had taken place along the Verde fault and that the lateral movement had not been determined. His conclusions were that the United Verde Extension ore body originally might have been closer to or farther away from the United Verde ore body than is indicated by only dip-slip movement along the Verde fault.

Ransome (1932, p. 21) was emphatic in his belief that the United Verde Extension ore body was formerly the top of the United Verde ore body, and he suggested that the original massive sulfide pipe was severed by a 2,400-foot vertical separation (throw) along the Verde fault in Precambrian time. The United Verde Extension ore body reached its present position by the additional 1,500-foot vertical separation after the deposition of the Paleozoic and Tertiary rocks. Apparently Ransome assumed that the two ore bodies were formerly connected because of the similarity in form and geologic relationship. The amount of Precambrian displacement necessary to connect the two ore bodies was determined by projecting upward the axis of the United Verde ore body (fig. 28) and projecting upward

the United Verde Extension ore body along the plane of the Verde fault—using as a line of slip, a rake of 70° to 75° E. This rake corresponds with the rake of rolls and grooves in the Verde fault plane (Reber, 1938, p. 51). Placing the projected 1,400 level United Verde Extension ore body on the projected axis of the United Verde ore body would give the position of the ore body before faulting (fig. 28, position III). Using this method of determining the total vertical separation (throw), our determination is essentially in agreement with the 4,000-foot total throw as suggested by Ransome, of which 2,500-foot displacement must have occurred before deposition of the Paleozoic rocks. (Reber (1938, p. 51) agrees with Ransome's interpretation.

The evidence seems clear that some Precambrian movement along the Verde fault is necessary to match the Precambrian rocks on opposite sides of the fault, but this cannot be accepted as proof that the United Verde Extension ore body is the severed top of the United Verde ore body. It is true that in general the form of the two ore bodies is similar, and both are south of the gabbro (United Verde diorite of Reber, 1922). However, the United Verde ore body is confined entirely to the Grapevine Gulch formation and intrusive quartz porphyry. The United Verde Extension ore body has replaced Deception rhyolite and quartz porphyry (pl. 5). The fissile schist in the United Verde Extension mine may not be Grapevine Gulch formation, but sedimentary interbeds in the Deception rhyolite. The Deception rhyolite-Grapevine Gulch formation contact at the surface is 700 to 800 feet northeast of the center of the United Verde ore body. The geologic setting, therefore, of the two ore bodies is different in spite of the close spatial relationship to the gabbro, and this fact alone casts serious doubt on the probability that the two ore bodies were connected before displacement along the Verde fault. This doubt appears valid whether the Deception rhyolite in the United Verde Extension mine is or is not facing east, and whether it is or is not separated from the Grapevine Gulch to the west by an earlier thrust fault.

The mass of Deception rhyolite at the surface east of shaft 7 is cut off by the Verde fault, and the trace of the north and south contacts in the plane of the Verde fault is shown on plate 5. The western contact of the rhyolite dips 70° to 80° E., whereas the Verde fault dips 60° E. The contact and fault by upward projection should intersect at about 1,000 feet vertically above the present surface. This same contact in the United Verde Extension mine by downward projection should intersect the Verde fault at about 300 feet below the 1,400 level (1,700 level, pl. 10, U.V.X. 11,225 N.; 7,750 E.). Matching these points of intersection of the western rhyolite contact in the footwall and hanging-wall sides of the Verde should give the vertical separation along the Verde, that is, about 2,800 feet. Subtracting the known Tertiary vertical separation of 1,500 feet, indicates that 1,300 feet of Precambrian vertical separation is required. Because the western contacts of rhyolite used in projection are igneous contacts and intersect the Verde fault at an acute angle, no precision is justified; and perhaps it is best to state that the Precambrian vertical separation measured on the displaced rhyolite contact is about 1,000 feet.

Some features of the gabbro north of the ore bodies can be used to obtain additional data on the amount of required Precambrian movement on the Verde fault in order to match the geology on opposite sides of the fault. The downward projection of the eastern contact of the gabbro in the United Verde Extension mine indicates that it would intersect the Verde at an elevation of about 2,500 feet, corresponding to the 2,600 level of the mine. In the United Verde mine, data are inadequate to state positively at what elevation the lower (underneath) eastern contact of the gabbro is cut by the Verde. The gabbro is clearly in fault contact on the 1,200 level in the United Verde mine at an elevation of 4,300 feet, which is 1,800 feet higher than the point where the gabbro cuts out against the Verde fault in the United Verde Extension mine. Both Tertiary and Precambrian vertical separation are included in this 1,800 feet, 1,500 feet for the Tertiary and 300 feet for the Precambrian. The elevation of the intersection of lower contact of gabbro with the Verde in the United Verde mine may be at lower elevations, corresponding to the 1,500 or 1,650 levels, which would decrease the total vertical separation by from 300 to 450 feet. In other words, the gabbro-Verde fault intersection on opposite sides of the Verde, from the data available, does not indicate as great a total vertical separation, 2,500 feet, as indicated by the Deception rhyolite east of shaft 7.

The eastern gabbro-Verde fault intersection again represents planes meeting at acute angles; no precision is justified, but an approximate displacement is indicated. If a Precambrian displacement of 2,500 feet took place, as suggested by Ransome, the east margin of the gabbro necessarily would have essentially followed the future course of the Verde fault for about 1,000 feet—that is, the eastern contact would have dipped to the east at about 60° in contrast to the dominant westerly dip. Igneous contacts cannot be used with assurance in projections; but nevertheless, using known trends for projection gives some measure of possible displacement.

The quartz porphyry near the United Verde mine appears largely as tongues parallel to the Grapevine Gulch formation and Deception rhyolite (pl. 5). These tongues plunge northward and northwestward, essentially parallel to the plunge of the minor folds in the Grapevine Gulch formation. Possibly the tongue of quartz porphyry in which shaft 7 was sunk (pl. 5) had a comparable plunge to the north, steeper than the southeastern gabbro contact. Furthermore, this tongue probably would narrow at higher elevations. It does seem reasonable to presume that the gabbro mass, projected upward and to the south, would partly cut out the narrow tongue of quartz porphyry at an elevation of about 1,000 feet above the present surface. Then the quartz porphyry on the west side of the Verde fault would have a northwest length north of the Deception rhyolite comparable to that exposed on the east side of the fault on the 1,400 level of the United Verde Extension mine. It would then be possible to match the quartz porphyry-gabbro contacts and both the northern and southern quartz porphyry contacts with Deception rhyolite.

The evidence, although weak because we are dealing chiefly with intrusive igneous contacts, indicates that a Precambrian 2,500-foot vertical separation along the Verde fault is more than is needed to match the Precambrian rocks on opposite sides of the Verde fault. A minimum vertical displacement of about 1,000 feet before deposition of Paleozoic rocks seems necessary. The greater the proposed Precambrian displacement in excess of 1,000 feet, the greater the difficulty in matching the eastern gabbro contact on opposite sides of the Verde fault.

Assuming a vertical separation of 1,000 feet in Precambrian time, and projecting upward the southeastern gabbro contact southeastward at 45°, would place the United Verde Extension ore body (fig. 28, position IV) before any faulting, about 1,300 feet to the east-southeast of the surface outcrop of the United Verde ore body (300 level). Projecting the United Verde ore body upward along the axis of the N. 20° W., 65° N. plunge, until its position was 1,000 feet vertically above the present surface, would bring the United Verde ore body approximately west of the suggested position (fig. 28), of the United Verde Extension ore body before faulting. The greater the Precambrian displacement, the closer the proximity of the two ore bodies would be before faulting.

The question has been raised that if the United Verde Extension ore body is not the faulted top of the United Verde ore body, roots of the separate United Verde Extension ore body should be found on the footwall side of the Verde fault. It should be noted, however, that the United Verde Extension ore body tapered with depth, and on the lowest levels the massive sulfide mass was small and not cut off by the Verde fault (fig. 29). Adding at least 1,000 feet of Precambrian displacement to the known 1,500-foot Tertiary displacement, means that at least 1,000 feet of rocks has been eroded from below the former position of the United Verde Extension ore body. East of the United Verde ore body on the footwall side of the Verde, low-grade copper-bearing quartz porphyry is present that might represent the roots of the United Verde Extension ore body.

Another possibility must be considered; in the discussion so far, it has been presumed that the United Verde ore body continued above the present surface along the projected axis of N. 20° W., 65° N. plunge. This upward projection is an assumption. Beginning at the 1,200 level of the United Verde mine, and appearing at higher levels, small bodies of massive sulfide are east of the main body. On several levels, these eastern bodies are separated from the main ore body by quartz porphyry, Grapevine Gulch formation, and black schist. During mining, however, it was discovered that these eastern massive sulfide bodies connected at higher levels to the main mass, and thus represent downward lobes or rolls of massive sulfide. Possibly in the higher levels of the United Verde ore body, now removed by erosion, the east margin of the United Verde ore body replaced rocks progressively to the east at higher elevations, and eventually replaced Deception rhyolite. Thus the United Verde Extension ore body may indeed represent the higher and easternmost part of the upward projected United Verde ore body. With this interpretation, the United Verde Extension ore body could represent the downfaulted top of the United Verde ore body. No compelling evidence is available to prove or disprove this possibility.

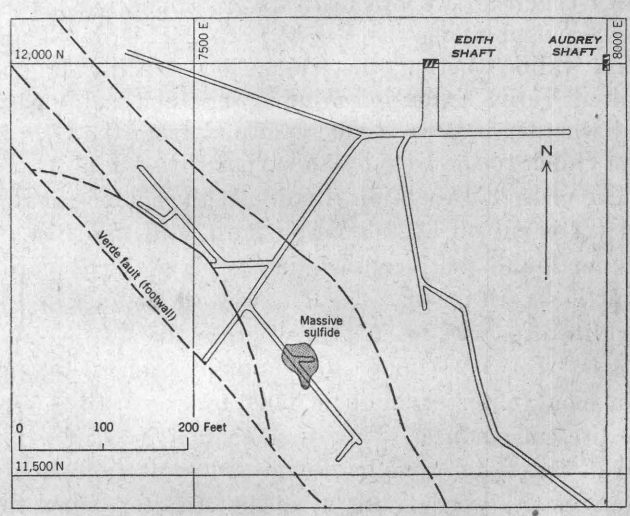

FIGURE 29.—Sketch map showing outline of massive sulfide on the 1,900 level, United Verde Extension mine. Furnished by Mingus Mountain Mining Corp.

The greater quantity of chalcocite ore in the United Verde Extension mine as compared with the United Verde mine has been used as an argument that the United Verde Extension ore body represents the severed enriched top of the United Verde ore body. The history has been so complex that doubt can be raised for this explanation. The evidence is strong that the United Verde Extension ore body was dropped about 1,000 feet in Precambrian time along the Verde fault. By the time the Tapeats sandstone(?) was deposited, the surface above both ore bodies was that of low relief, implying that the United Verde ore body was eroded at least 1,000 feet by Tapeats(?) time. If the two ore bodies were separate before faulting, any appreciable chalcocite in the United Verde ore body would undoubtedly have been removed by the active erosion that would have followed the uplift of the block containing the United Verde mine. Sometime before deposition of the Tertiary Hickey formation, a northward-trending canyon was cut deep into the Paleozoic rocks, and exposed some of the leached gossan in the United Verde Extension mine (fig. 26). The United Verde ore body may have been uncovered during this same period of erosion. After accumulation of the Hickey formation, and Tertiary displacement along the Verde fault, the Paleozoic and Tertiary rocks covering the United Verde ore body were removed by erosion and at least 350 feet of the top of the ore body was eroded judging from the present stream gradients west of the Verde fault. The chalcocite ore in the United Verde mine may have formed during that period of erosion or be relict from the pre-Hickey and (or) Precambrian period of weathering and erosion. Two or three periods of active erosion occurred, therefore, at the site of the United Verde ore body; whereas after displacement along the Verde fault the United Verde Extension ore body was probably only weakly eroded during Precambrian time. Subsequent burial by Paleozoic and younger rocks has protected the United Verde Extension ore body from erosion, except for the fleeting period of time in which the pre-Hickey canyon removed a small amount of quartz and gossan from the leached zone (fig. 26). These facts preclude the explanation that because the chalcocite enrichment in the United Verde ore body is much less than in the United Verde Extension ore body, the latter is the faulted segment of the former.

In summary, the evidence at hand is sufficient to demand movement along the Verde fault before deposition of the Paleozoic and Tertiary rocks. We believe that this Precambrian vertical separation was about 1,000 feet, and that the total vertical separation (throw) was nearly 2,500 feet. Following this interpretation, the United Verde Extension ore body could not be the severed top of the United Verde ore body unless the United Verde ore body shifted eastward at higher elevations above the present erosion surface and did not follow the upward projection along its known axis of N. 20° W., 65° N. plunge.

JEROME VERDE MINE

The Jerome Verde Development Co. owns 28 patented claims north and east of the United Verde Extension mine. The present company was incorporated in 1921 to take over the holdings of the Jerome Verde Copper Co.

The only production came from the Main Top claim and this ore body, known as the Main Top, is a faulted segment of chalcocite ore along the Verde fault and a part of the United Verde Extension ore zone. The ore was discovered in 1917 and mined out by 1920 by the United Verde Extension Mining Co. Lindgren (1926, p. 88) reported that about 10,000 tons of ore mined contained from 8 to 12 percent of copper. Elsing and Heineman (1936, p. 101) state that the production totaled 1,500,000 pounds of copper, $10,000 of gold, and $14,000 of silver, the total value of production amounting to $345,000. If 10,000 tons of ore were mined, the average grade must have been about 7½ percent of copper to match the production of 1,500,000 pounds of copper.

The location of the Main Top ore body is shown on 1,300 level, plate 10 at 11,950 N.; 7,350 E. The ore body was mined from the underground workings of the United Verde Extension mine and stoped from the 1,300 level to the 10th floor above the 1,100 level. The ore body was about 200 feet long and from 10 to 30 feet wide.

Much exploration was done in the Precambrian rocks north of the United Verde Extension mine on Jerome Verde ground during the early exploratory work by the United Verde Extension Mining Co., and in cooperation with this company, but the results were fruitless. Early work on Jerome Verde property was done from the Columbia shaft, 1,000 feet southeast of the Edith shaft in the United Verde Extension mine (Lindgren, 1926, p. 88). The Columbia shaft is 1,061 feet deep and connects with 12,000 feet of workings, mainly on the 1,100 level, which corresponds to the 1,400 level of the United Verde Extension mine. Precambrian rocks were reported at a depth of 426 feet in the Columbia shaft. In 1922, additional work was done north of the United Verde Extension mine without significant discovery.

COPPER CHIEF ORE BODY

DEVELOPMENT AND PRODUCTION

The Copper Chief ore body is 3½ miles south-southeast from Jerome and is accessible by road from Cotton-

wood. A property line divided the ore body; the Copper Chief Mining Co. owned the western part and the Equator Mining and Smelting Co., the eastern part. Phelps Dodge Corp. now owns both properties.

The Equator Mining and Smelting Co. was owned originally by Senator Clark who also controlled the United Verde mine. In 1904–5, about 30,000 tons of sulfide ore were mined and smelted in a plant east of the mine. Production during that period totaled 1,300,000 pounds of copper, having a value of $185,000 (Elsing and Heineman, 1936, p. 101). The Iron King claim was the site of this production, and the mine is known either as the Iron King or Equator. In order to avoid confusion with the Iron King lead-zinc mine near Humboldt, the combined Copper Chief-Equator (Iron King) ore body will be referred to as the "Copper Chief ore body."

The western part of the ore body owned by the Copper Chief Mining Co. consists largely of oxidized ore, mined chiefly for the gold and silver content. Most of the production came during 1916–18 by the Hayden Leasing Co. which had a cyanide mill with a capacity of 125 tons daily. The total output during 1916–18 was 71,849 tons of ore that returned $875,800 or about $12 to the ton (Lindgren, 1926, p. 91). Production from 1916 to 1923 included $530,000 in gold and $372,000 in silver. The combined Copper Chief-Equator production totaled $1,087,000 (Elsing and Heineman, 1936, p. 101). A few hundred tons of sulfide ore was mined and shipped to the Humboldt smelter when the price of copper was high (Lindgren, 1926, p. 92; Reber, 1938, p. 56).

During World War II, R. L. d'Arcy and A. B. Peach had a lease on the Copper Chief ore body and produced oxide ore from both ends of the ore body; the east end yielded some oxide copper ore.

The Copper Chief end of the ore body was developed by a shaft 350 feet deep that connected to an adit whose portal was south of the ore body at an altitude of 5,481 feet. Stopes of oxide ore extended from the 240 level to the surface. Withdrawal of ore and waste has formed a large pit at the surface corresponding to the original extent of the gossan.

The Equator end of the ore body was developed by an adit about 80 feet higher than the Copper Chief adit. The Equator adit, with the portal on the east end of the ore body, is about 500 feet long from the portal to the property line. Stoping extended from 15 feet below the adit level to about 70 feet above. A lower adit, 680 feet long at an altitude of 5,344 feet is to the east; no ore was found in it.

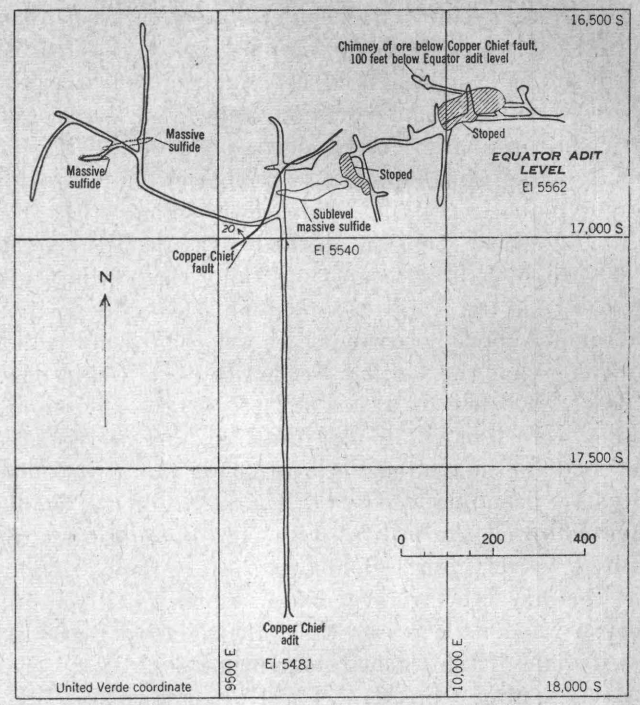

FIGURE 30.—Map showing underground workings, Copper Chief deposit. Compiled by G. W. H. Norman from data furnished by Phelps Dodge Corp.

GEOLOGY

The Copper Chief ore body is massive sulfide that has replaced highly foliated and sheared Shea basalt and tuffaceous sedimentary interbeds. The Shea basalt has been intruded by quartz porphyry, which at the surface forms the north margin of the ore zone where the contact is essentially vertical and some fault movement is indicated. On lower levels, the quartz porphyry contact dips gently north and the ore zone is wholly within foliated Shea basalt, except for small dikes of quartz porphyry south of the ore zone. Granodiorite porphyry dikes trending north-northeastward cut the Shea basalt, quartz porphyry, and ore zone.

The foliation in the Shea basalt strikes N. 65°–80° E. and dips 50°–75° N. Steeper dips were observed at the surface and lower dips within the mine. The quartz porphyry mass north of the ore zone is poorly foliated.

The Copper Chief fault, named by Reber (1938, p. 55), dips west and northwest at a very low angle, and is exposed at the portal of the upper Equator adit and in the Copper Chief adit (fig. 30). The Copper Chief fault is largely a sheeted zone containing some introduced quartz parallel to the sheeting. The northeastward-trending granodiorite porphyry dikes are displaced to the east in the hanging-wall block, and Reber (1938, p. 56) after a careful study of the displaced dikes reported a horizontal displacement of about 300 feet.

ORE ZONE

The primary ore in the Copper Chief ore body is massive pyritic sulfide shaped as an elongate lens trend-

ing N. 80° E. in the highly foliated and sheared Shea basalt, but secondary (oxide) ore extended beyond the limits of the massive sulfide. The surface trace of the ore zone is about 800 feet long; the massive sulfide lens is much shorter and plunges to the east. The massive sulfide replaced the foliated Shea basalt essentially parallel to the foliation so that the southern irregular contact (footwall) dips north. The north margin of the sulfide lens at and near the surface is regular and vertical, controlled by the quartz porphyry contact. On lower levels where quartz porphyry is absent, the north (hanging-wall) margin of the lens dips north parallel to the foliation in the Shea basalt, but is narrower than at the surface.

The western part of the ore body contained a gossan about 60 feet wide and 250 feet long, but the top of the massive sulfide, at a depth of 240 feet, was smaller (Reber, 1938, p. 55). At a depth of 290 feet the sulfide lens was 160 feet long and had a maximum width of 30 feet (fig. 30), and at the 320 level, the massive sulfide bottomed on foliated Shea basalt (Reber, 1938, p. 55).

The top of the gossan plunges eastward and on the surface only a narrow streak of gossan is exposed that follows the quartz porphyry contact to the portal of the upper Equator adit, where a mass of gossan and oxidized copper ore is exposed above and in the Copper Chief fault zone. Massive sulfide is exposed within the adit (fig. 30), within and above the Copper Chief fault; it is strongly brecciated and granulated, proving that the last movement on the Copper Chief fault is younger than the massive sulfide. Near the base of the Copper Chief fault, exploratory workings below the adit level revealed massive sulfide stringers parallel to the foliation in the Shea basalt (N. 65° E., and 60° N.), but the main mass of sulfide was not present below the fault.

According to Reber (1938, p. 55–56) an extensive diamond-drilling program was undertaken on the assumption that the Copper Chief fault cut the ore body; but negative results of this exploration make it fairly certain that there was no downward faulted segment of the Copper Chief ore body, and that the Copper Chief fault merely cuts the bottom of the east end of the massive sulfide lens. Exploration work below the Copper Chief fault did reveal (fig. 30) a small chimney of massive sulfide of about 600 square feet in cross section that trends vertically downward in Shea basalt. Some copper and zinc are reported in the massive sulfide chimney.

In the western workings of the Copper Chief adit, two small thin lenses of massive sulfide were discovered parallel to the foliation in the Shea basalt above the Copper Chief fault (fig. 30). These lenses appear too far north to represent extensions of the Copper Chief ore body and may be downward extensions of gossan exposed on the Wonderful claim west of the Copper Chief ore body, where a little mineralized rock occurs.

The massive sulfide is like the pyritic facies in the United Verde mine and locally contains streaks of chalcopyrite and sphalerite. Galena must be present because of the lead carbonate in the gossan. Chlorite forming black schist is rare and is best observed on the surface at the southwest corner of the pit. The quartz porphyry north of the ore body contains local streaks of chlorite.

A thin zone of secondary enrichment between the massive sulfide and oxide zones contains sooty chalcocite. At the west end of the ore body, the chalcocite zone is from 2 to 5 feet thick, thinning to the east end of the ore body. The chalcocite ore graded downward into sandy pyrite and upward into gossan containing some concentration of copper and lead carbonate minerals. The chalcocite ore is reported to have contained 40 ounces of silver and 0.75 ounce of gold to the ton; the copper content was between 1 and 2 percent.

The oxide zone consists of soft limonitic gossan like that at the United Verde mine, except that the lead content was higher (Reber, 1938, p. 56). The grade of the minable gossan was 0.3 ounce of gold and about 6 ounces of silver to the ton (Lindgren, 1926, p. 91).

RELATIONSHIP OF MASSIVE SULFIDE TO GRANODIORITE PORPHYRY DIKES

The age relationship of the massive sulfide minerals to the granodiorite porphyry dikes cannot be determined with certainty from the accessible underground workings. Some sulfide minerals are clearly younger than the dikes as shown by a quartz vein, 1 inch wide, exposed in the Copper Chief adit, cutting Shea basalt and dike rock. Pyrite is present where the vein cuts the dike, and pyrite and chalcopyrite where it cuts Shea basalt. At the surface, over the eastern part of the ore zone, the granodiorite porphyry dikes contain small stockworks of low-dipping narrow quartz veins containing limonite pseudomorphs after pyrite. When the Copper Chief and Equator mines were more accessible for detailed studies, Reber mapped the underground workings; and from the manner that the dikes cut through the massive sulfide lens, Reber (personal communication) believed that the evidence indicated that the dikes are younger than the massive sulfide lens.

VERDE CENTRAL MINE

The Verde Central mine is on the south wall of Hull Canyon near the Jerome-Prescott highway and about 4,000 feet south of the United Verde mine. The mine was inaccessible in 1952. The mine is owned by Phelps Dodge Corp. The collar of the main shaft is at an

altitude of about 5,500 feet, and the bottom is the 1,900 level at an altitude of 3,584 feet. Besides the development work in the productive area, extensive exploratory work was done to the south on both the 100 and 1,000 levels, and a long crosscut was driven to the west on the 1,450 level.

The Verde Central ore zone is along the east margin of the quartz porphyry in contact with the Deception rhyolite. The contact is irregular and interfingering, and dike-offshoots of porphyry cut the rhyolite. The Verde Central was a blind prospect, and the only outcrop of oxidized copper is at the top of the hill, southeast of the main workings.

Two ore bodies were found in the Verde Central mine. One, at the Deception rhyolite quartz porphyry contact, extended south from the shaft, where black schist contained spotty copper showings. This ore body extended from above the 800 to below the 1,000 levels and was about 200 feet long and 9 feet in average width (Reber, 1938, p. 57).

The other, a much larger ore body is a tabular vein-like body of quartz, pyrite, and chalcopyrite in Deception rhyolite east of the main body of quartz porphyry. The vein strikes northwest and dips east, and although irregular in detail, ranges in width from 5 to 30 feet. On the 1,000 level, the vein ore body is essentially continuous for more than 1,000 feet. According to Reber (1938, p. 57), this ore body extends nearly to the top of the mine, and "is fairly strong at the bottom." Chalcopyrite is subordinate to quartz and pyrite, but ore shoots averaging nearly 3 percent copper could be mined. Below the 600 level, such ore made up a part of the vein.

High copper prices were needed to allow the Verde Central to produce copper profitably, and the net profit on the past production plus the potential profit from developed ore remaining in the mine would fall short of returning the cost of exploratory work (Reber, 1938, p. 57).

The appearance of quartz, pyrite, chalcopyrite, and chlorite (black schist) indicates the presence of mineralized rock like that in the United Verde mine. The restriction of chalcopyrite to vertical shoots in the vein indicates a break between pyrite and chalcopyrite, as in the United Verde mine. But there is not much indication that the black schist followed the main pyrite deposition, as in the United Verde mine; for the most part, the black schist appears to be localized along the quartz porphyry contacts. The low content of precious metals may be related to the scarcity or absence of tetrahedrite, galena, and sphalerite, common as minor constituents in the United Verde deposit (Reber, 1938, p. 57).

SHEA MINE

The Shea mine was owned and operated originally by the Shea Copper Co. The holdings consisted of 16 patented claims and 3 unpatented claims. The property is now owned by Phelps Dodge Corp. The mine is located south of the Copper Chief mine on the north side of an east-striking vein that dips south at an average of 42°; at the surface the dip is steeper. The vein was developed by an incline on the vein, 1,220 feet on the incline or 825 feet vertically. The lower tunnel level reached the surface at about 1,300 feet east of the shaft. About 3,000 feet of drifts are on this level. Work on all levels totals about 7,000 feet (Reber, 1938, p. 58).

Production amounted to $65,000 worth of silver, copper, and gold, obtained from 1,200 to 1,300 tons of sorted shipping ore. Silver accounted for 80 percent of the value (Reber, 1938, p. 59).

The Shea basalt is greatly altered near the Shea vein, and chlorite, sericite, epidote, quartz, and carbonate minerals (ankerite? and calcite?) are common. Northward-trending dikes of granodiorite porphyry cut the Shea basalt and are in turn cut by the Shea vein. The dike rocks are strongly sericitized near the vein.

The vein minerals were deposited in a normal fault that has a probable throw of about 100 feet. The vein crops for about 1,200 feet east of the portal of the main adit; its total length is more than 3,400 feet. The surface exposure ends less than 200 feet west of the main shaft where the low-dipping Copper Chief fault cuts off the Shea vein. Both the most western and the eastern parts of the vein contain lenticular and discontinuous vein quartz. Throughout the mine the vein quartz does not average more than a foot in width, although in the ore shoot, the vein exceeds 5 feet.

A large part of the vein consists of coarsely crystalline quartz containing scattered pyrite and a few bunches of coarse ankerite or siderite and pyrite with some chalcopyrite. Locally, some arsenopyrite and tetrahedrite are present. The tetrahedrite, carrying only a little arsenic, contains 600 ounces or more of silver to the ton. Where sulfide minerals including tetrahedrite are most abundant, the vein has a banded structure. Near the ore shoot, sparse galena is present in quartz veinlets cutting the older quartz.

Mining was limited to one locality, on the 300 level from 300 to 400 feet west of the shaft to where the Shea vein is cut off by the low-dipping Copper Chief fault. Near the top of the ore shoot, native silver was associated with tetrahedrite.

The Copper Chief fault is partly mineralized, and above the Shea ore shoot several tons of barite were exposed in the low-dipping fault. The barite is not

definitely associated with any other vein material and may be younger than the sulfide.

DUNDEE-ARIZONA COPPER CO.

The Dundee-Arizona Copper Co. owns two patented claims covering the ridge east of the town of Jerome and north of the Jerome-Clarkdale highway. The deposit is of interest because chrysocolla-bearing Verde gravel deposits have yielded the only production. From February 1942 to July 1947, at which time premiums on copper were discontinued, Mark Gemmill, in a leasing operation, mined about 100 tons per week of hand-sorted ore that averaged about 4½ percent copper. The ore was shipped to the smelter at Clarkdale.

In 1920, a shaft was sunk 950 feet. The collar is in Verde gravel deposits, and the shaft cut through the Martin limestone, Tapeats sandstone(?) at the 500-foot level and into Precambrian rocks for an additional 450 feet. Lateral exploration on the 950 level amounted to 2,750 feet. Precambrian rock fragments on the dump include quartz porphyry, Deception rhyolite, and gabbro. Apparently no copper was found in the Precambrian rocks.

The base of the Verde gravel has been explored by 3 levels for a length of 700 feet and width of 80 feet. The portal of the lower level, an adit, is 200 feet southeast of the shaft at an altitude of 4,625 feet. To the west, the two higher levels at altitudes of 4,644 and 4,675 feet, connect by raises to the adit level.

The gravel commonly consists of pebbles, cobbles, and boulders of basalt and limestone, resting on the Martin limestone. The ore consists of chrysocolla-bearing gravel, locally forming ore zones from 4 to 5 feet thick. The blue-green chrysocolla coats and replaces the pebbles and cobbles and fills fractures in the underlying Martin limestone. The chrysocolla forms a mammillary coating on the pebbles, associated with some chalcedony. According to Lindgren (1926, p. 90) the copper-bearing solutions, undoubtedly derived from the oxidation of the United Verde deposit, must have carried cupric sulfate with a great abundance of silica; for otherwise, the copper carbonate minerals would have formed by reaction with the limestone.

PROSPECTING OPERATIONS IN THE VERDE MINING DISTRICT

Much exploratory work was done south and north of Jerome in hopes of finding ore bodies like the United Verde and United Verde Extension. Many millions of dollars was spent without encouraging results, and Reber (1938, p. 42) stated "Away from the quartz-porphyry belts, two to three of the prospecting companies developed enough ore to supply the stockholders with specimens." Most of the properties are closed and inaccessible and practically all the information given below is taken from Lindgren (1926, p. 93–97).

Grand Island Mining Co.—This company owned property south of the Shea mine (1,346,400 N.; 447,400 E.) and sank a shaft 520 feet deep with a southwest crosscut on the 200 level. The shaft was sunk where one of the northward-trending granodiorite dikes intersects the low-dipping Copper Chief fault. The fault is locally silicified and mineralized, and a short thick lens produced more than 100 tons of high-grade copper-gold ore. There was little tetrahedrite, and pyrite carried the gold (Reber, 1938, p. 59). The property is now owned by Phelps Dodge Corp.

Green Monster Mining Co.—A large block of claims north and northeast of the Copper Chief is owned by this company. The main shaft is east of the Verde fault and north of the road leading to the Copper Chief mine (1,351,000 N.; 453,000 E.). Precambrian rocks are exposed in the canyon below the Tapeats sandstone(?), and the collar of the shaft is in quartz porphyry. Fragments of quartz porphyry, Shea basalt, and tuffaceous sedimentary rocks are on the dump. Specular hematite veinlets appear in many fragments, but no sulfide minerals were observed. According to Lindgren (1926, p. 94), the shaft is 930 feet deep and on the 500 level, a northwest crosscut contacted the quartz porphyry containing chalcocite and chalcopyrite in a highly leached zone. West of the Green Monster shaft and of the Verde fault, an adit level containing about 800 feet of workings, explored a northeastward-trending gossan. It is reported that some gold was found.

The Cliff claim is near the western boundary of the Green Monster property and north of the Copper Chief mine (1,350,700 N.; 448,200 E.). Two extensive adit levels prospected a small zone of mineralized quartz porphyry at the southern contact with the Shea basalt. The lower adit, inaccessible in 1947, contains about 1,400 feet of workings, whereas the upper adit contains about 800 feet. Two short adits, each 90 feet long, cut the upper eastern part of the mineralized zone, and a short shaft connects to one of these adits. The copper minerals appear to be controlled by an eastward-trending shear zone, dipping 70° N. and from 6 to 8 feet wide. The quartz porphyry locally contains appreciable chlorite in streaks parallel to the shear zone. Copper carbonate is common at the surface and coarse pyrite and some chalcopyrite are on the dumps. Work was discontinued on the property in 1919 (Lindgren, 1926, p. 94).

Jerome Del Monte Copper Co.—This company owns a large group of claims most of which are east of the Verde fault and north of the Green Monster property. The Jerome Del Monte property extends from north

of Mescal Gulch to south of Del Monte Gulch. In 1917, a three-compartment shaft (1,358,300 N.; 451,400 E.) was sunk in limestone to 500 feet, at which depth water was found.

Gadsden Copper Co.—West of the Jerome Del Monte property and about 1½ miles southeast of Jerome, 39 claims are owned by the Gadsden Copper Co. Most of the claims are east of the Verde fault and extend from north of Mescal Gulch to south of Del Monte Gulch. A four-compartment shaft (1,362,000 N.; 445,200 E.) was sunk 1,200 feet deep, which connects 5,619 feet of drifts and crosscuts. The shaft was sunk through limestone and reached the Precambrian rocks at a depth of 400 feet. On the 1,200 level, the rocks are predominantly Deception rhyolite cut by one or two dikes of quartz porphyry. The Verde fault was cut 1,450 feet to the west of the shaft on the 1,200 level. It was reported that ore was found southwest of the shaft that contained from 8 to 10 percent copper, 3 ounces of silver to the ton, and 0.15 ounce of gold to the ton. No information is available as to the size of this ore body; operations discontinued in 1920 (Lindgren, 1926, p. 94).

Verde Combination Copper Co.—Property owned by this company includes a large group of claims south of Mescal Gulch and west of the Verde fault. A shaft, 1,300 feet deep, was sunk west of the Verde fault and a short distance south of Mescal Gulch (1,359,500 N.; 444,500 E.). The shaft was started in Deception rhyolite; and on the dump, fragments of chloritized Deception rhyolite contain seams of pyrite. No copper minerals were observed. Lindgren (1926, p. 94) reports that the underground workings showed a 2-foot fissure vein of 5 percent copper ore.

Pittsburgh-Jerome Mining Co. — This company owned 21 patented claims in the upper part of Mescal Gulch and did prospecting for 16 years, ending in 1921. Phelps Dodge Corp. now owns the property. The exploration work consisted of driving an adit (1,354,500 N.; 439,700 E.) 1,279 feet long, and sinking a connecting shaft, 900 feet deep, 1,000 feet west of the portal. The rocks exposed are the Deception rhyolite, including flow and fragmental rocks and andesitic pyroclastic interbeds. Small clots of intensely chloritized rhyolite are common, particularly in the area that was prospected between the two forks of Mescal Gulch. Malachite in bleached rocks and stringers of chalcocite in rock fragments have been found on the dump.

Cleopatra Copper Co.—This company owned claims in Hull Canyon, south of the United Verde mine; the property is now owned by Phelps Dodge Corp. The principal adit is the Dillon tunnel, 3,200 feet long, which connects to the north with the United Verde 500 level. The rocks on this property are Deception rhyolite and quartz porphyry. North of the Hull fault in quartz porphyry, some copper sulfide is present in irregular quartz stringers and small blebs of massive sulfide. Some ore was shipped and a small smelter was built in the lower part of Deception Gulch west of the Verde fault. A small slag pile is all that remains.

Calumet-Jerome Copper Co.—A small group of claims south of Deception Gulch and to the west of the Verde fault was formerly owned by this company. The property is now owned by Phelps Dodge Corp. A shaft 600 feet deep was sunk just to the west of the Verde (1,361,000 N.; 441,800 E.). It is reported that a crosscut extended west from the shaft on the 600-foot level for 1,800 feet. The underground workings total 7,000 feet. The fragments on the dump are of Deception rhyolite in part chloritized and containing quartz veinlets with pyrite. Lindgren (1926, p. 95) states that chlorite schist rich in chalcopyrite was reported in 1922.

Jerome-Grande Copper Co.—The Verde Grande Copper Co., later reorganized as the Jerome-Grande Copper Co., owned a small group of claims north of Hull Canyon and east of the Warrior fault. The property is now owned by Phelps Dodge Corp. A shaft, 838 feet deep, was sunk east of the Warrior fault (1,361,500 N.; 435,000 E.) and connects with about 2,000 feet of underground workings. The collar of the shaft is in Grapevine Gulch formation. It is reported that the sump is in gabbro and that on the 800 level, a contact of gabbro and quartz porphyry showed some pyrite and chalcopyrite. Lindgren (1926, p. 96) states "The mine was opened with the object of finding the southward extension of the United Verde ore bodies, but this object was not attained."

Arkansas & Arizona Copper Co.—A large group of claims north of the United Verde mine and east of the Verde fault is owned by this company. A shaft (pl. 1, 1,370,000 N.; 438,000 E.), 1,600 feet deep, was sunk through Tertiary basalt and Paleozoic limestone, entering Precambrian rocks at 800 feet. The Verde fault was intersected by a drift, 400 feet west of the shaft. On the 1,600 level, gabbro was found 500 feet southeast of the shaft. Deception rhyolite and quartz porphyry are present elsewhere on this level as well as Tertiary basalt dikes. Reports are conflicting as to whether or not ore was found on the 1,600 level, near the Verde fault.

Jerome Superior Copper Co.—This company owns a very large area of claims north of the Verde Tunnel and Smelter Railroad tract in the Clarkdale quadrangle, and some distance east of the Verde fault. About 3 miles west of Clarkdale (1,374,000 N.; 444,200 E.) a shaft, collared in Verde gravel, was sunk 900 feet. On the dump abundant fragments of quartz porphyry and

Grapevine Gulch formation indicate that the shaft bottomed in Precambrian rocks. It is reported that encouraging showings of chalcocite and chalcopyrite were found, but only pyrite was observed on the dump.

BIG BUG MINING DISTRICT

By S. C. Creasey

The Big Bug mining district, one of the oldest in the region, is bounded on the east by the Agua Fria River and on the north by Lonesome Valley. The district extends south to include the Blue Bell mine (fig. 9), but only the northern part of the district has been studied for this report. The most productive mine and only mine operating in 1952 in the district is the Iron King lead-zinc mine. The McCabe-Gladstone mine was an important early producer of gold and silver, and the Arizona National and Silver Belt mines have produced lead-silver from the same vein. These mines are inactive. The northern part of the Big Bug district has been included in this report because of the increasing importance of the Iron King mine.

The available records (to 1952) indicate that the above-named mines in the Big Bug mining district have produced 321,500 ounces of gold, about 8,500,000 ounces of silver, 3,773 tons of copper, 37,537 tons of lead, and 104,599 tons of zinc.

IRON KING MINE

PRODUCTION

From 1906 to January 1, 1952, the Iron King mine produced about 211,486 ounces of gold, 6,624,101 ounces of silver, 72,573,620 pounds of lead, 209,199,140 pounds of zinc, and 6,345,740 pounds of copper from about 1,640,060 tons of ore (table 24). The average grade of all ore mined on the 400 level and lower levels through 1951 was $7 per ton in gold and silver, 2.21 percent lead, and 6.378 percent zinc. Dollar value of gold generally was about twice that of silver. Samples of the concentrates and tailings were analyzed for selenium by the Geological Survey with the following results: lead concentrate 0.07 percent; zinc concentrate 0.03 percent; pyrite concentrate 0.02 percent; and tailings 0.005 percent.

Table 24.—*Production of recoverable metals, Iron King mine, Humboldt, Ariz.*

[Production data supplied by and published with the permission of Shattuck Denn Mining Corp.]

Year	Tons	Gold (ounces)	Silver (ounces)	Lead (pounds)	Zinc (pounds)	Copper (pounds)
1906–38 [1]	78,452	15,690	313,808	3,138,080	6,276,160	470,700
1938 [2]	13,477	2,317	45,938	404,300	1,078,160	67,400
1939	70,227	9,911	272,604	1,872,680	5,854,020	351,120
1940	65,812	9,239	266,497	1,891,060	7,220,440	329,060
1941	69,159	9,720	331,746	2,320,040	7,617,100	345,800
1942	88,200	11,659	392,458	3,540,100	10,585,560	441,000
1943	73,721	9,167	307,465	3,164,380	10,095,300	220,720
1944	99,164	9,460	308,567	3,611,660	13,623,860	423,820
1945	117,287	13,068	436,506	5,259,640	16,156,180	455,280
1946	115,615	13,065	467,387	5,734,280	16,875,320	485,780
1947	122,368	15,298	533,642	6,194,880	16,925,320	411,820
1948	145,823	17,036	540,548	6,854,120	19,048,100	453,020
1949	175,111	21,432	737,925	8,414,680	23,547,440	546,660
1950	203,063	27,289	904,284	10,645,040	28,220,800	686,460
1951	202,581	27,135	764,731	9,528,680	26,075,380	657,100
Total	1,640,060	211,486	6,624,101	72,573,620	209,199,140	6,345,740

[1] Production before milling.
[2] Last 3 months of milling.

DEVELOPMENT AND MINING [3]

Shaft 1, the oldest on the property, is located between shafts 2 and 5; only the dump is shown on plate 11, as the collar is caved. Shaft 1, 100 feet deep, serves only for ventilation. Shaft 2 (pl. 11, 3,000 N.; 4,050 E.), 650 feet deep has 2 compartments, and although the head frame is still standing, it serves only as an emergency exit. Shafts 3 and 4 are caved, and records do not indicate their location. Shaft 5 (pl. 11, 2,360 N.; 3,830 E.), 200 feet deep, has 2 compartments; it is used only for ventilation. Shaft 6 (pl. 11, 3,390 N.; 4,070 E.) has 3 compartments with steel head frame, and is the main working shaft of the mine for hoisting ore, men, and supplies. A new working shaft, No. 7, was being sunk in 1952; it is north of shaft 6, beyond the limit shown on plate 11.

Levels are established at 100-foot intervals, and in 1952, ore was being extracted from the 1,300 level, development work was in progress on the 1,400, 1,500, and 1,600 levels and a station was being cut on the 1,700 level of shaft 6.

Drifts are along the veins, and raises extend above them to explore and prepare ore for stoping. The first ore was mined in the oxidized zone about the 200 level. Apparently, ore was mined through shaft 2 from the 200 to the 600 level until shaft 6 was ready for hoisting ore. The exploitation of the mine to the 600 level was not difficult. Walls stood well and the ore was mined and drawn by shrinkage stoping without noticeable dilution, and it was not necessary to fill the stopes with waste. As the mine was further explored to the north at lower levels, the hanging-wall (west) side of the veins became weaker, and in 1946, cut and fill methods were adopted for stoping. Waste for stope fill was obtained by blocking caving barren schist over the north end of the 600 level.

For some years, the development drifts also served as haulage drifts, but the cost of maintenance became excessive, and these haulage drifts are now driven in the firm footwall (east) side of the mineralized zone, and crosscuts are driven at 100-foot intervals across the

[3] Most of the information in this section was taken from Kumke and Mills (1950).

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mineralized zone which serves to explore the zone as well as give access to draw points under the stopes.

Mills (1941, 1947) has described the history, ore occurrence, and mining methods; and Hendricks (1947) has reported on the milling practice.

GENERAL FEATURES

The Iron King deposit is a system of 12 westward-dipping massive sulfide veins that are oriented en echelon in a mylonitic sheared zone characterized by a pronounced foliation. The andesitic tuffaceous unit of the breccia facies of the Spud Mountain volcanics forms the wall rock, but on the west or hanging-wall side of the veins, hydrothermal alteration—chiefly silicification, sericitization, and pyritization—modified the rocks, so that in many places the original rock cannot be determined. Reverse strike faults cut the veins; the vertical separation of the faulted veins is generally less than 120 feet.

OLDER PRECAMBRIAN ROCKS

SPUD MOUNTAIN VOLCANICS

The Iron King deposit is in the upper unit of the Spur Mountain volcanics. West of the Iron King deposit, this upper unit comprises interbedded coarse- and medium-grained andesitic tuff, a minor amount of fine-grained tuff, and at least one bed of breccia. The coarser grained tuff abounds in large saussuritized plagioclase crystals or their sheared equivalent, so characteristic of the Spud Mountain volcanics. Undoubtedly, many of these coarse-grained beds originally were crystal tuffs. The interbedded medium-grained tuffs are like the matrix of the crystal tuffs.

East of the Iron King deposit, crystal tuff and conglomeratic beds are absent, and the andesitic tuffaceous unit consists chiefly of alternating beds of medium- and fine-grained andesitic sedimentary tuff. In addition, a rhyolitic tuff and conglomerate bed about 200 feet thick crops out about 500 feet east of the Iron King deposit (pl. 11). The fine-grained andesitic tuff is much more abundant east of the deposit than west. Judging from relicts, the rocks in the zone now occupied by the Iron King deposit (including the hydrothermally altered rocks) did not include any beds of crystal tuff.

IRON KING VOLCANICS

Basalt of the Iron King volcanics crops out about 800 feet east of the Iron King deposit (pl. 11). It consists of grayish-green lava and minor amounts of intercalated sedimentary andesitic-basaltic tuffaceous rock. The lava is well foliated, but relict amygdules of quartz, carbonate, and epidote minerals are abundant. Chlorite and saussuritized plagioclase are the most abundant megascopic minerals.

TERTIARY ROCKS, HICKEY FORMATION

The Tertiary Hickey formation comprises the alluvial fill in the south end of Lonesome Valley. Near the Iron King deposit, it consists of boulders and cobbles crudely sorted in thick beds and intercalated with minor amounts of finer sediments. The Hickey formation, in large part, covers the alteration zone and veins of the Iron King deposit with as much as 500 feet of gravel. Underground openings have exposed the veins for more than 2,000 feet north of the point where they plunge beneath the gravel.

RHYOLITIC DIKE

A light-colored rhyolitic dike crops out about 700 feet east of the Iron King deposit. It is about 15 feet wide and extends from beneath the alluvial fill of Lonesome Valley south-southwestward for about 1,950 feet. A smaller rhyolitic dike of similar composition crops out about 300 feet south of the end of the larger one, and extends southward beyond the limit of the area shown on plate 11.

The dike rock is spherulitic. Thin sections reveal that the rhyolite consists of irregular-shaped, microscopic grains of quartz embedded in a cryptocrystalline base.

The rhyolitic dike is younger than the youngest dynamic metamorphism, which is Precambrian, but there is no way of dating it precisely. In the Bradshaw Mountains, rhyolitic porphyry dikes have been mineralized, and Lindgren (1926, p. 24) has tentatively dated them as Cretaceous or Tertiary. It appears likely that the dike in the Jerome area is one of these rhyolite porphyry dikes.

STRUCTURE

The Precambrian rocks near the Iron King deposit have been folded, faulted, and foliated, and transformed into mineral assemblages stable under low-grade metamorphism. At least two periods of deformation took place, resulting in complex structures.

An early period of deformation resulted in foliation trending northward in the breccia facies of Spud Mountain volcanics lying west of the Iron King deposit (pl. 6). The northeastward-trending folds and foliation, however, nearby and to the east may or may not have formed at the same time as those trending northward. The later deformation resulted in faults and foliation trending northeast and in lineation plunging steeply north. Foliation resulting from the later period or periods of deformation cuts the earlier northward-trending foliation in the breccia unit of the Spud Mountain volcanics (pl. 6, 1,270,500 N.; 390,000 W.), but parallels or accentuates the earlier foliation near the deposit. The shearing that localized the Iron King de-

posit was probably part of the later deformation. The early phase of mineralization occurred before the end of this deformation, for the minerals introduced by this phase were subsequently granulated and foliated.

The intense deformation of the rocks near the Iron King mine was regional in extent. The rocks are isoclinally folded, nearly all dips are steep, and many of the beds are overturned. Foliation formed concomitant with folds, resulting in approximate parallelism of axial planes of folds and foliation with the possible exception of the northward-trending foliation west of the Iron King mine (pl. 6). On the limbs of folds, bedding and foliation are parallel; but on crests and troughs, they are discordant. Within this disturbed terrane, the Iron King deposit lies in the middle limb of a fold between a syncline to the east and an anticline to the west (pl. 6). Here bedding and foliation are essentially parallel.

Faults undoubtedly are far more abundant near the deposit than is indicated (pls. 1 and 11). Faults are difficult to recognize partly because of lack of marker beds in the Iron King and Spud Mountain volcanics, and partly because of coincidence with foliation. Considerable movement along a relatively narrow zone in the foliated rocks might not be recognized, as deformation was so penetrative and intense that all units and structural elements tend to be parallel.

THE IRON KING DEPOSIT IN RELATION TO REGIONAL STRUCTURES

The Iron King deposit was localized in a sheared and deformed zone essentially parallel to the regional foliation. In this zone, cataclasis rendered much of the rock a fine- to very fine-grained fissile phyllite, but locally relicts of coarser grained andesitic tuff testify to the former similarity to the andesitic tuff adjacent to the deposit. Sporadically, the more fissile facies have been contorted, buckled, or gently flexed into wavelike structures during cataclasis, or by later deformation, or by both. After this deformation, early vein minerals were introduced, and the rocks hydrothermally altered; then renewed deformation once again granulated the early vein minerals. For example, vein quartz so mylonitized is now granoblastic except for a few relicts of bladed quartz with ragged or sutured boundaries (pl. 13*A*), and some quartz recrystallized in the pressure shadows of deformed or recrystallized grains of pyrite. A still later and apparently independent deformation superimposed a fracture cleavage on this intensely deformed zone.

This intense and recurrent deformation possibly was related to the same forces that produced the regional metamorphism, even though the later shearing took place after the rocks were foliated. Lineation is the common feature. Lineation (mineral streaking) in the walls of the veins and lineation that marks the northward plunge of the north ends of the veins both parallel lineation in the nearby rocks. This parallelism attests to the similarity of the stress fields during regional metamorphism and during the localized shearing that permitted access to the mineralizing solutions or fluids. It thus seems most probable that both are integral parts of one long and complex period of deformation. Post-mineral structural features are discussed on page 168.

DEPOSITIONAL HISTORY

Before deposition of early hydrothermal minerals and possibly toward the close of the foliation-producing deformation, shearing disrupted the continuity of the regionally foliated terrane and produced a local sheared zone of pronounced foliation. Tuffaceous beds of differing competence may have localized the shearing. In this sheared zone, pyrite, ankerite, quartz, and sericite were deposited. Along the east or footwall side of the sheared zone, quartz, pyrite, and ankerite were concentrated in echelon shears forming well-defined veins. West of the veins, these minerals and sericite were deposited as disseminations and as narrow veinlets alternating with schistose rock. These early veins probably were similar in form to the present ones; the quartzose north ends of the veins are relics of the early mineralized zone.

Later the early veins and the sporadically mineralized zone to the west were intensely sheared. There is no reason to suspect a time gap between deposition of early minerals and the second period of shearing, which may have interrupted deposition of the early minerals.

The second period of shearing mylonitized the northern parts of the veins, producing poorly defined planar structures that cut at an acute angle from the footwall to the hanging wall. Subsequently, sphalerite, arsenopyrite, galena, tennantite, sparse chalcopyrite, sericite, and probably pyrite, ankerite, and quartz were deposited in the fabric produced by the later period of shearing. The similarity of distribution of zinc plus lead and gold plus silver (figs. 31, 32, 33) indicates that all productive veins had identical histories of mineral deposition and probably all formed simultaneously.

North-end quartz largely escaped base-metal mineralization possibly because of its monomineralic and more massive character. Quartz "noses" were present before the ore-forming minerals, for the quartz is cut by veinlets containing ore minerals and the texture of the quartz is chiefly granoblastic-cataclastic, indicating strong deformation, plate 13*A*.

Locally high precious-metal content in quartz is difficult to interpret. Perhaps the location of greatest con-

centration of precious metals north of greatest concentration of base metals suggests that the peak of mineral deposition of the precious metals was offset from that of the base metals. There is no compelling evidence to place gold in the base-metal period—silver, providing it is chiefly in tennantite, is definitely associated with galena, hence, came in during the later period.

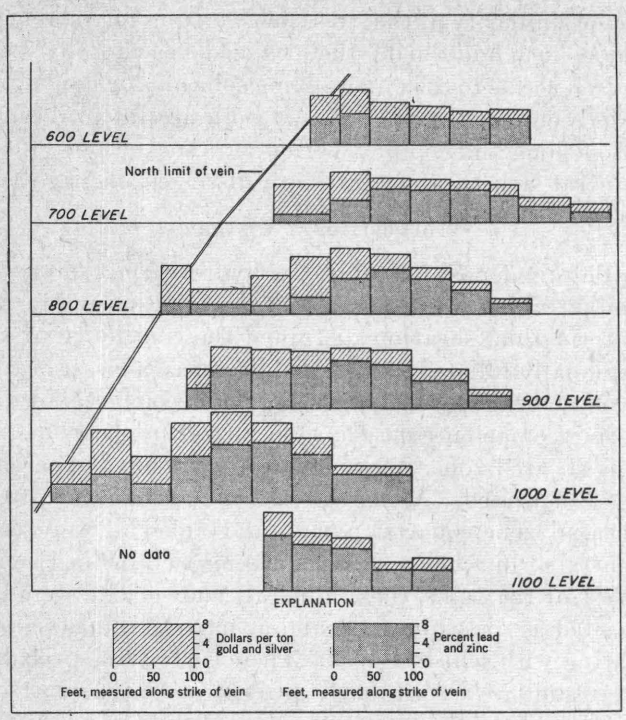

FIGURE 31.—Graph showing grade of ore, *G* vein, Iron King mine. Data furnished by Shattuck Denn Corp.

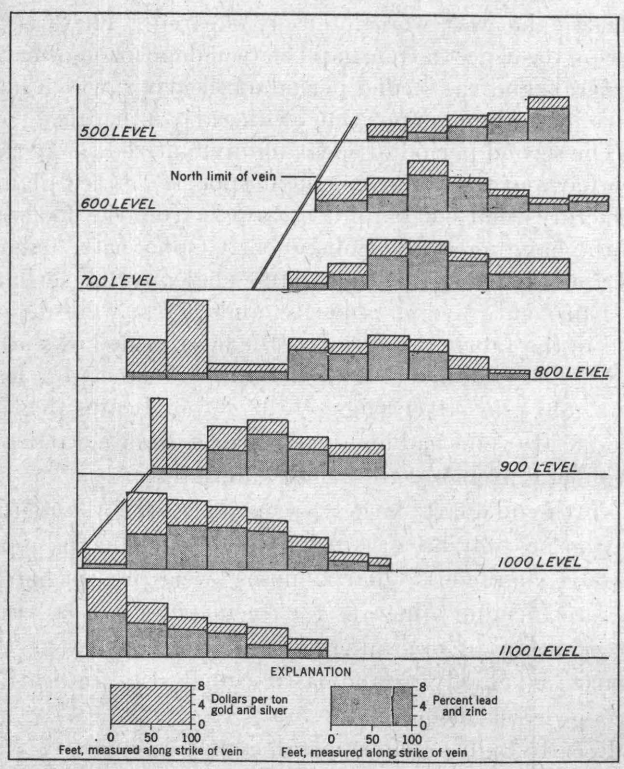

FIGURE 32.—Graph showing grade of ore, *F* vein, Iron King mine. Data furnished by Shattuck Denn Corp.

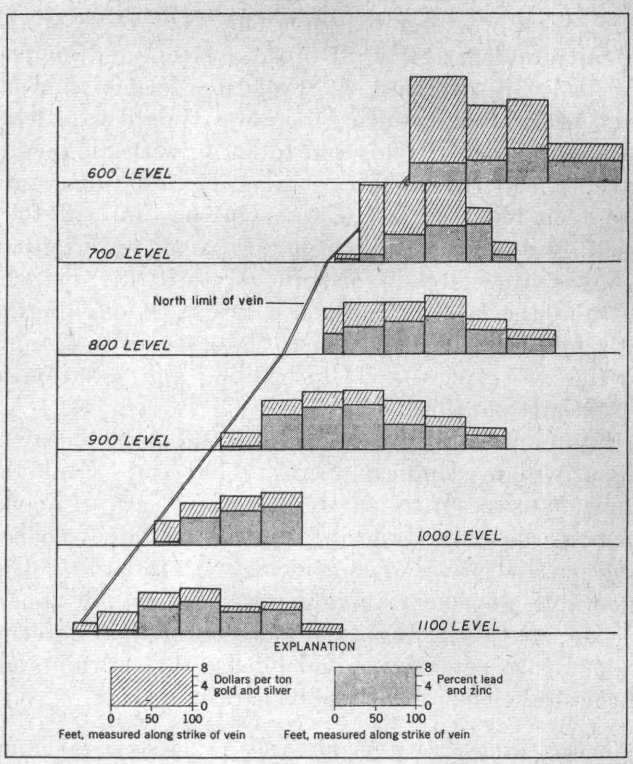

FIGURE 33.—Graph showing grade of ore, *D* vein, Iron King mine. Data furnished by Shattuck Denn Corp.

HYDROTHERMAL ALTERATION

Designation of the early phase of mineral deposition as hydrothermal alteration is arbitrary, but useful in distinguishing between (1) early widespread introduction of gangue minerals and (2) ore-forming minerals concentrated in the veins. Deformation, manifested by shearing, provides the basis for separation, although no break in mineral deposition is implied.

The minerals introduced during hydrothermal alteration are quartz, pyrite, ankerite, sericite, and probably apatite. A second generation of sericite and ankerite, later than the ore minerals, is described in this section for convenience. Quartz occurs in disseminated granoblastic grains as a general silicification and in veinlets associated with pyrite and ankerite. The strong shearing that followed the introduction of the hydrothermal minerals makes it impossible to distinguish introduced quartz, unless it is in veinlets from quartz originally present in the metamorphic rocks, even though quartz is more abundant in the altered rocks. The texture of the quartz in veinlets is largely granoblastic, but a few patches of relict bladed quartz remain.

Pyrite occurs as individual crystals disseminated in the altered rocks and as veinlets associated with other alteration minerals. A series of individual crystals of pyrite along a given "folium" is common in rocks in underground openings of the mine. Brown-stained cavities in outcrops mark the distribution of pyrite in altered rock. Ankerite is abundant, occurring as disseminated grains and as veinlets, commonly associated with pyrite and quartz. Ankerite apparently accompanies both early and late stages of mineral deposition, and it veins the last-stage sulfide minerals.

There are two generations of sericite. The early sericite is disseminated in the alteration zones, and the late sericite, with ankerite, is concentrated locally near the northern parts of the veins. The early sericite is in flakes and "ribbons" concordant with foliation. The "ribbons" anastomose through the rocks, in places dividing and rejoining. The sericite grew with its basal plane parallel to the shear planes that controlled its distribution.

Within the veins, late sericite in microscopic veinlets cuts quartz, pyrite, ankerite, and sphalerite, and individual flakes lie in the cleavage planes in ankerite and in pressure shadows of pyrite, transverse to planar structures in the vein material. Filmlike veinlets of sericite cut the quartz at the north ends of the veins and commonly separate this quartz from the remainder of the vein. Residual screens of wall rock in veins commonly are rich in sericite. Adjacent to the veins, sericite of this generation is restricted to narrow selvages that generally widen around the north ends of the veins to masses as much as 10–20 feet across.

In the wall rocks, spatial distribution is the only basis for distinction between early- and late-stage sericite. The sericite in the veins is clearly younger than the ore minerals, and it is assumed that the local concentrations of sericite adjacent to the veins are of the same age. The sericite disseminated throughout the alteration zone has the same distribution as the pyrite, ankerite, and quartz with which it is assumed to be contemporaneous. No reliable textural criteria were found for distinguishing early from late sericite in wall rocks, for here sericite lies with its basal section parallel to the foliation whether it formed before or after the intramineral deformation that is the basis for distinction between the hydrothermal alteration stage and the ore-forming stage.

Thin sections from outcrops of the hydrothermal alteration zone contain a small amount of a probable clay mineral that has a negative $2E$ of about 35° and a birefringence near 0.010. The indices of refraction could not be determined from thin sections and not enough was found for determination in oil mounts of powdered rocks. The clay(?) mineral occurs as irregular microscopic masses composed of fibers or narrow plates commonly oriented at right angles to the long dimensions of the masses and to the foliation in the rock. Growth of fibers or plates perpendicular to foliation indicates that the clay(?) mineral formed after the last period of deformation. This clay(?) mineral, which is probably from the kaolin group, was not recognized in rocks from underground and accordingly is believed to be a weathering product. The mineral probably formed in an acid environment produced by oxidization of pyrite. Bateman (1927, p. 585) has described kaolinization owing to oxidation of sulfide minerals in sericitic alteration zone near massive pyrite deposits of Rio Tinto, Spain.

Table 25 gives the results of chemical analyses of the fresh and altered tuffaceous rock from the Iron King deposit. Analysis 1 is the relatively unaltered mafic tuff, which is basaltic in composition; 2 is the analysis of the silicified, sericitized, and pyritized rock forming the hanging-wall alteration zone; and 3 is the sericitized and carbonatized rock that encloses the north ends

TABLE 25.—*Chemical analyses of unaltered and altered rocks, in percent*

	1	2	3	4	5
SiO_2	48. 38	59. 12	41. 88	48. 50	58. 70
Al_2O_3	12. 18	12. 13	11. 72	18. 10	20. 45
Fe_2O_3	2. 90	. 29	. 45	4. 66	. 11
FeO	[1] 10. 44	[1] 9. 91	[1] 8. 84	4. 86	2. 05
MnO	. 21	. 16	. 17	------	------
MgO	3. 67	5. 95	4. 33	5. 67	Tr.
CaO	7. 81	1. 11	9. 85	7. 10	. 50
BaO	. 01	. 02	. 09	------	------
K_2O	. 11	1. 45	2. 45	None	1. 88
Na_2O	2. 80	. 48	. 50	6. 64	3. 38
TiO_2	1. 76	. 56	. 51	------	------
H_2O-	. 14	. 12	. 20	------	------
H_2O+	3. 69	3. 70	2. 14	. 38	1. 05
S	. 36	4. 58	1. 07	------	------
P_2O_5	. 30	. 12	. 10	------	------
CO_2	5. 61	1. 51	16. 17	2. 46	. 02
N_2O	------	------	------	. 12	. 14
$BaSO_4$	------	------	------	Tr.	1. 50
FeS_2	------	------	------	. 12	9. 58
$CuFeS_2$	------	------	------	None	. 36
	100. 37	101. 21	100. 47	98. 61	99. 72
Less O for S[2]	. 09	1. 15	. 27		
Total	100. 28	100. 06	100. 20		
Bulk densities	2. 81	2. 78	2. 85		
Powder density	2. 86	2. 87	2. 90		

[1] In the usual method of determining ferrous iron, pyrite is not dissolved and the ferrous iron so present will be missed. In these analyses a correction was made by calculating the ferrous iron in the pyrite from the amount of sulfur present. It was assumed that all sulfur is present as pyrite and that the formula for pyrite is FeS_2.

[2] Based on the assumption that all sulfur is present as pyrite.

1. Unaltered mafic tuff, 900 level, Iron King mine.
2. Altered rock, quartz-sericite-pyrite facies, 900 level, Iron King mine.
3. Altered rock, sericite-carbonate facies, 900 level, Iron King mine. Analyses 1, 2, and 3 by Robert N. Eecher.
4. Fresh uralite-schist from a vertical depth of 420 feet and 100 feet from the Americansky ore body, Ural Mountains, Kyshtim, Russia.
5. Sericite-schist from a vertical depth of 520 feet and 2 feet from the Americansky ore body, Ural Mountains, Kyshtim, Russia. Analyses 4 and 5 after Stickney (1915, p. 630).

of the veins. Figure 34 graphically shows the losses and gains of the altered rocks in milligrams per cubic centimeter in comparison with the relatively unaltered rock.

In the hanging wall, the alteration resulted in addition of SiO_2, K_2O, and S and in leaching of Na_2O and CaO. Comparison of analyses of fresh and altered rock show a slight loss in iron, which probably is within the limits of variation of different samples. The significant point concerning iron is that about half the iron present in silicates and ankerite in the unaltered rock apparently united with sulfur in the altered rock to

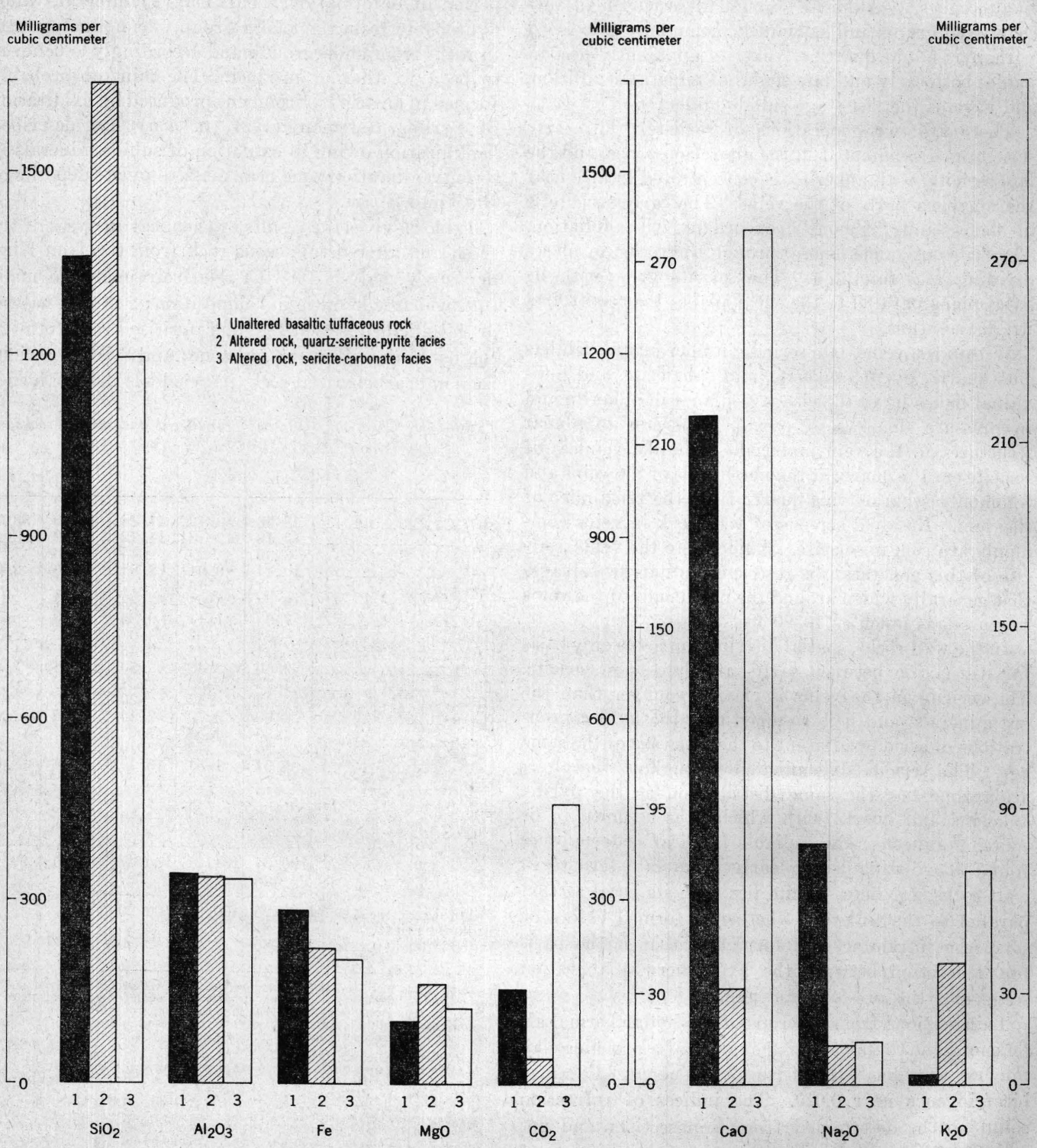

FIGURE 34.—Graph showing losses and gains of principal constituents in the alteration of the chlorite schist, Iron King mine.

form pyrite; the bulk of the remaining iron remained in the chlorite, for there is little carbonate left in the altered rock, as shown by the low CO_2 content. The introduced K_2O is in sericite. The introduced SiO_2, joined by SiO_2 released by breakdown of silicates, crystallized as free quartz, which occurs in veinlets and in disseminated silicification. The small increase in MgO is probably within the limits of variation from one sample of mafic tuff to another; MgO was probably neither introduced nor removed.

The altered rock surrounding the quartz noses consists chiefly of quartz, ankerite, sericite, and pyrite, locally somewhat stained by iron oxide. The analysis of this rock (table 25, analyses 3) reveals some interesting features. There is more Al_2O_3 than is needed to unite with the K_2O to form sericite. Small amounts of Al_2O_3 probably are combined with Na_2O in soda mica or in residual albite that was not entirely destroyed during the alteration. A small amount of clay minerals might be present, which would account for additional Al_2O_3, but none were recognized in section. The approximate composition of the ankerite was determined by its omega index (1.71–1.70), and all the CO_2 in the analysis was assumed to be in the ankerite. These two assumptions suggest that all the CaO and nearly all the MgO are in the ankerite. The iron content exceeds the amount necessary to fulfill the needs of the ankerite and to combine with the sulfur as pyrite. Some or all the iron above the needs for these two minerals probably occurs as iron oxide, which forms a secondary stain on the foliation planes. Despite the reduction in total SiO_2 from that of the unaltered rock, more than enough SiO_2 is present to supply the needs of the silicate minerals. The remaining SiO_2 forms free quartz, commonly in veinlets, which gives the rock a siliceous appearance.

Stickney (1915) has described a group of massive sulfide veins in the Ural Mountains, Kyshtim, Russia, that are similar in many ways to those in the Iron King deposit. One of these similarities is the character of the hydrothermally altered rocks adjacent to massive sulfide veins. The country rock near the deposits at Kyshtim is "uralite schist." It is composed of andesine usually altered to zoisite and albite, uralite, actinolite, diallage, and some chlorite, epidote, and quartz. Adjacent to the massive sulfide veins, however, this mafic rock is altered to a quartz-sericite-pyrite rock, similar to the hydrothermally altered rock in the Iron King deposit.

Table 25 gives the chemical composition of the unaltered and altered rocks in the Americansky deposit at Kyshtim. The similarity of the losses and gains to the altered rocks at the Iron King is most striking: in both there is a loss of CaO, Na_2O, and CO_2, and a gain of SiO_2 and K_2O. In the Americansky deposit, most of the iron in the original mafic minerals plus a little added iron has been combined with sulfur in pyrite, whereas in the Iron King deposits, iron was not added but the iron in the original mafic minerals is chiefly in pyrite in the altered rock. The most significant difference in the losses and gains is in the MgO. The quartz-sericite-pyrite rocks from the Americansky deposit show a decided leaching of MgO, whereas those from the Iron King do not. In the Americansky deposits, the mafic minerals appear to be more completely altered to sericite, and the carbonates appear to be leached to a greater extent. Loss of MgO would certainly occur in alteration of the mafic minerals to sericite, but would occur in the leaching of the carbonate minerals only if they contained magnesium.

The hydrothermally altered rocks associated with the Iron King deposit were divided into three intensity zones, whose contacts are generally arbitrary (pl. 11). The alteration in zone 1, adjacent to the veins, is most intense, and the rock contains a greater concentration of quartz, pyrite, and ankerite. Along the strike southward, zone 1 grades into 2 in which these minerals are less abundant but still common. Zone 2 grades into 3 southward beyond the limits shown on plate 11. In addition, zone 3 embodies separate belts of altered rocks lying east and west of zones 1 and 2. In zone 3, sericite appears to be the dominant platy mineral; pyrite is very rare; and secondary quartz and ankerite are not megascopically abundant except in local patches.

Zone 1 or the hanging-wall alteration zone is on the west or hanging-wall side of the Iron King deposit. It is the largest zone, ranging on the surface from 100 to 250 feet in width and cropping out for a length of about 2,500 feet. Zone 1 was studied in most detail, owing to more complete exposure by underground openings.

In zones 1 and 2, layers in which pyrite, quartz, and ankerite are more abundant alternate with layers in which these minerals are distributed sparsely and in which the original type of rock (mafic tuff) is apparent through partly altered relics. These alternating layers persist to the walls of the veins, where commonly both grade into sericitic schist. Chlorite is abundant in the layers that contain sparse hydrothermal minerals. In a few thin sections, crosscutting microscopic veinlets of chlorite attest to the mobility of chlorite during the alteration.

It appears that distinct veins formed along the east or footwall side of the sheared zone, while within it (zones 1 and 2), minerals were sporadically introduced along foliation planes and fractures. Quartz, pyrite, and ankerite compose the veins for a short distance

south of the ore shoots. Farther south, schistose partings become more abundant. About 400 to 500 feet south of the well-defined veins, the position of the veins can be recognized by layers in which there is a greater concentration of veinlets of quartz, pyrite, and ankerite alternating with wall rocks. Still farther south, these layers pass imperceptibly into altered rocks of zone 1.

Zone 3 is characterized by sericite in which the eastern unit appears to be at a slight angle to bedding, using the nearby rhyolitic tuffaceous bed as a reference (pl. 11). Rocks of zone 3 have a more pronounced foliation than adjacent andesitic tuffaceous rocks, probably because more intense shear localized the alteration and because shearing also took place later. The typical sericitic altered rocks grade into partly altered relics of andesitic tuffaceous rocks through rocks in which both sericite and chlorite are abundant. Foliation in the relic is less pronounced.

An unsolved problem is whether any of the original rocks in the alteration zones were rhyolitic. Most of the partly altered relics indicate an original andesitic or basaltic composition. Especially in rocks of zone 3, sporadic outcrops appear to be rhyolitic. Whether the rhyolitic composition is original or secondary is uncertain, but the possibility of some rhyolitic beds intercalated in the original rock certainly was not eliminated.

Outcrops of the alteration zones are generally white or cream colored and in zones 1 and 2 are stained by iron oxide derived from pyrite. At first, the white color was attributed to abundant sericite, but rocks below the water table are generally green, and thin sections from surface rocks contain abundant pale-green chlorite as well as sericite. Hence the light color probably is due in part to bleaching by sulfuric acid formed by oxidation of pyrite, like that described by Lasky (1936, p. 64) for many veins in the Bayard area, New Mexico. In outcrops of alteration zones 1 and 2, casts after pyrite are common, and sulfuric acid has dissolved most of the carbonate.

SULFIDE VEINS

Two groups of veins are in the Iron King deposit: the well-defined massive sulfide veins, which yielded all but a few tons of the production; and the poorly defined nonproductive ones lying to the west of the massive sulfide. The productive veins comprise the group along the footwall (east side) of the alteration zone. They crop out in an area about 2,500 by 100 feet (pl. 11). By 1948, subsurface openings on the productive veins extended to a depth of about 1,140 feet below the collar of shaft 6, and the vein system had been mined for about 3,400 feet along the strike (pl. 12).

The poorly defined veins, here called nonproductive to distinguish them from the ore veins, crop out erratically in the alteration zone to the west. Their strike lengths are short, and probably they are just as discontinuous with depth. They parallel the foliation, and commonly consist of pyrite, ankerite, and quartz. One of these veins—the copper vein—contains chalcopyrite also. It is the widest and most continuous of the nonproductive veins, but its continuity between its north and south segments is uncertain (pl. 11).

STRUCTURAL CONTROL

Fracturing and shearing controlled spatial relationship, width, length, and to some extent internal structure of the veins and size and shape of the alteration zones. A slight angular discordance between the strike of the veins and the strike of the alteration zones is apparent, as shown by the north ends of the veins terminating in footwall rocks (east of the deposit) and the south ends grading into alteration zone 1. Whether this discordance indicates a difference in age between the shearing that controlled the alteration zones and the shearing that controlled the veins is uncertain. However, early gangue minerals in the veins, although more concentrated, are the same as those sporadically distributed throughout the (hanging-wall) alteration zones, indicating that at the time of deposition of the earliest minerals both structures were present. This does not preclude the possibility of two periods of shearing before mineral deposition. Significance of the slight angular discordance between the strike of the veins and of the alteration zones is difficult to interpret, especially as the rock affected was strongly anisotropic owing to foliation and possibly to beds of different competency. Close spacing of the echelon fractures that control the veins and the relatively wide spacing of similar points of the veins on the levels such as the north ends of the veins suggest that fractures or faults resulted from a shear couple. Lineation (mineral streaking) in the walls of the veins and in the north ends of the veins plunges steeply northward. Judging from the interpretation of the regional geology, this lineation marks the *a* structural coordinate; if so, the direction of relative movement plunged steeply northward.

Strong fracturing and brecciation of early introduced minerals clearly indicate a second period of movement that controlled the localization of ore-forming minerals in the Iron King mine. Distribution of ore-minerals indicates that the later movement took place along fracture zones that extended from footwall to hanging wall of the barren quartz-pyrite-ankerite veins.

The ore deposit at the Eustis mine, Quebec, has a somewhat similar en echelon arrangement of massive sulfide lenses. Stevenson (1937, p. 348) attributes the localizing structure to action of two opposite and

tangential forces. In the Flin Flon mine, according to Koffman and others (1948, p. 295–301) massive sulfide ore in sheared limbs of drag folds has an echelon pattern and consistent plunge. Here, shearing occurred along the contact of competent lava flows and more easily sheared pyroclastic and flow breccia. Drag folds were not recognized at the Iron King, but variation in competency of adjacent tuffaceous beds may have localized shearing.

Massive sulfide bodies localized by fractured and brecciated zones have been described by Finlayson (1910, p. 406) for the Huelva deposit, Rio Tinto district, Spain; and by Capps and Johnson (1915, p. 92) for massive sulfide deposit in the Ellamar district, Alaska. Stickney (1915, p. 620) found that the amount of sulfide increased as the schist became more thinly foliated or sheeted. This relationship is also true of the Iron King deposit. Hanson (1920, p. 574–609) studied pyritic massive sulfide deposits in Canada to determine the relationship of form and structure of the ore bodies to enclosing rock. After considering such mines as the Mandy, Manitoba; the Northpines, Ontario; the Flin Flon, Manitoba-Saskatchewan; and the Eustis, Quebec, he concluded that the sulfide resulted from replacement in "zones of more intense shearing and brecciation." The Sullivan deposit, Canada, appears to be an exception, for, according to Swanson and Gunning (1945, p. 651), a stratigraphic zone controls the deposit. However, tourmaline and cassiterite in the deposit indicate a greater depth of origin than for many massive sulfide deposits.

DISTRIBUTION AND CHARACTER OF THE VEINS

The ore deposit consists of 12 veins arranged en echelon, plate 12; individually they strike about N. 22° E. and dip 71° NW. As shown on the map each vein extends farther to the north than the adjacent one on the east. In section, they maintain a similar en echelon arrangement; in any particular vertical section each one to the west extends to a higher altitude than the one on the east. All veins plunge northward, and individually the plunge is commonly constant, the angle ranging between 55° and 60°. Individually the width is commonly constant for short distances but ranges from 1 to 14 feet for different veins; the lengths are measurable in hundreds of feet. From southeast to northwest, the veins are designated as follows: *X*, *Y*, *P*, *A*, *B*, *C*, *D*, *E*, *F*, *G*, *H*, and *I*, plate 12 (*P* vein does not extend to the 900 level; the *B* vein is omitted on some levels).

I vein, although designated as one vein, actually consists of four closely spaced individual ones arranged en echelon, of which the fourth was not exposed completely by underground openings (pl. 12). It is longer than any of the others because of its composite character. Individual echelons of *I* vein have all characteristics of the other veins, such as quartz "noses," consistency of plunge, and characteristic mineral distribution.

Some of the veins deviate from the ideal echelon pattern. On 900 and 700 levels, *X* vein consists of two separated segments, but on 800 level only one vein exists; they must join and split between 700 and 900 levels. *B* and *C* veins, lying between *A* and *D*, are subsidiary structures, as they do not maintain the echelon pattern; *A* vein which probably should be considered as parallel to *B* and *C*, extends as far to the north as *C* and considerably farther than *B*.

P vein is anomalous, for it plunges less steeply than the others and is completely blocked out to its edges within the zone mined. The 25° northward plunge of *P* vein is at least 25° flatter than any other, so that it occupies an intermediate position between *Y* and *A* veins. *P* vein apexes above the 400 level and bottoms a short distance below the 700 level (fig. 35). On the 400 level it is almost a part of *Y* vein; however, in depth it diverges to the north and on 700 level, 250 feet separates *P* vein from the north end of *Y*, where it lies next to and about as far north as *A* vein.

The continuity (except for *P* vein) of veins and schist septa and their general widths are consistent from the surface to 1,100 level; and with rare exceptions, such as the local junction of the north ends of *E* and *F* veins, adjacent veins do not join. The septum between *E* and *F* is commonly not more than 2 feet wide, and the septum separating *H* vein from *G* and *I* is most commonly not more than 5 feet. In places, as on 1,100 level,

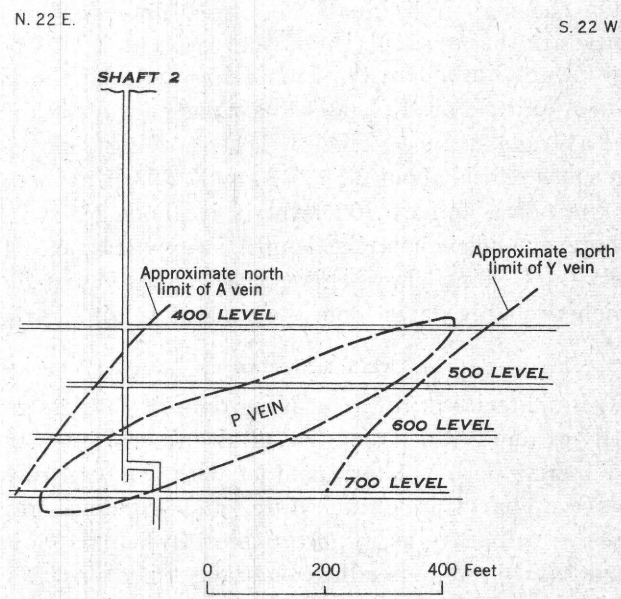

FIGURE 35.—Longitudinal projection, *P* vein, Iron King mine.

faulting after mineral deposition moved the veins together, but these effects are not considered here. Variations in widths of *C*, *H*, *F*, and *G* veins are known. Between 200 and 600 levels *C* ranges in width from 5 to 10 feet, but it is narrow and unproductive on lower levels. *H* vein ranges from 10 to 15 feet in width on 300 and 400 levels and is 10 feet wide on 500 level, but it narrows to about 2 feet between 500 and 700 levels. Below 600 level, *G* increases in width almost commensurate with the decrease in width of *H*. According to Mills (1941, p. 57), *F* changes into a wide low-grade extremely siliceous ore body between 400 and 500 levels, but it narrows to its former width before reaching 600 level.

Thin schist partings or "screens" commonly occur within the veins. The most common type enters from the hanging wall, generally south of ore shoots, and terminates after continuing northward for distances as much as several hundred feet. In places, the last few feet are thin chloritic or sericitic films only a few millimeters thick. Thick schist partings that cross *I* vein divide it into several parts. In a few places narrow partings enter the veins from the footwall, migrate well into them and then back into the footwall tuff. Some veins contain many narrow "screens," ranging in length from a few inches to as much as 100 feet, that on the map are shown entirely enclosed in sulfide. In three dimensions many of these "screens," especially the larger ones, undoubtedly connect to wall rocks.

One of the most striking features of the veins is the abrupt change from massive sulfide to wall rock. In places a knife blade will cover the contact; whereas in others a narrow septum of gouge separates vein from rock.

On the map individual veins are shown to have a gentle arc shape, which is convex eastward. This feature is not limited to the Iron King deposit; the Kit Carson and Silver Belt-McCabe veins curve similarly. The average surface strike of the Iron King deposit at the south end is about N. 25° E. to N. 30° E.; whereas on the north end of 700 level it is about N. 20° E. Wherever sufficient strike length is exposed (pl. 11), individual veins show this same general curvature, which probably dates from the time of origin.

VEIN MATERIAL

Vein material in the Iron King deposit consists generally of fine-grained massive sulfide minerals and massive quartz both in sharp contact with the wall rocks. Massive quartz is gray, white, and greenish gray. Massive sulfide veins range in color from pale yellow to nearly black, depending on the ratio of pyrite to sphalerite and carbonate. Fine banding produced by differences in relative amounts of pyrite, sphalerite, or gangue is almost universally present in the massive sulfide ore (pl. 13*C*, *D*); banding is less marked or absent in nonproductive parts of veins (pl. 13*B*).

The sulfide vein minerals are pyrite, arsenopyrite, sphalerite, chalcopyrite, galena, and tennantite; the gangue minerals are ankerite, quartz, sericite, and a little residual chlorite. Gold and silver constitute the rare metals. Gold is free and, according to Mills (1947, p. 4), is carried largely in pyrite; silver is probably in tennantite. Pyrite is the dominant sulfide and in places makes up about 75 percent of the vein. Quartz and ankerite are the dominant nonsulfide minerals, and locally either may be the more abundant. North ends of the veins are almost exclusively quartz; whereas in the central section, ankerite is commonly more abundant.

The vein material is very fine grained, with most grains between 100 and 300 microns in diameter but ranging from as much as 1.5 millimeters to less than 10 microns. Arsenopyrite grains attain a diameter of 1.5 millimeters, and intergrowths of galena and tennantite form grains 1 millimeter long. Some pyrite crystals, especially in southern sections of the veins, attain diameters of about 300 to 400 microns.

Crystal habits range from idiomorphic to xenomorphic granular. Pyrite commonly has idiomorphic outlines against all other minerals, but aggregates of pyrite etched with nitric acid, reveal xenomorphic outlines. Arsenopyrite shows crystal outline against galena, ankerite, and sphalerite, but not against pyrite or other arsenopyrite. Crystal outlines of sphalerite, galena, and tennantite were not observed; but in places they replace ankerite, retaining the rhombohedral cleavage angle in the outline of the grain. They commonly occur as smaller grains interstitial to pyrite and arsenopyrite.

Banding is common in most massive sulfide deposits. It has been attributed to pseudomorphism of various older structures, such as bedding, fracture, and foliation (pl. 13*C*, *D*). Massive sulfide deposits are characteristically fine grained, thereby favoring preservation of small features by pseudomorphism.

Banding in the massive sulfide from the Iron King deposit probably resulted (1) from deposition of ore-forming minerals (chiefly sphalerite) in fractures or microscopic shear planes in early vein filling and (2) from variation in relative rates of deposition of minerals. The vein material in general consists of bands ranging from 1 millimeter to 10 centimeters in width (figs. 36, 37*A*, *B*). Banding generally parallels vein walls, but in detail is commonly irregular. The irregularities consist of: (1) continuous, wavy bands, (2) discontinuous, nonparallel banding, and (3) irreg-

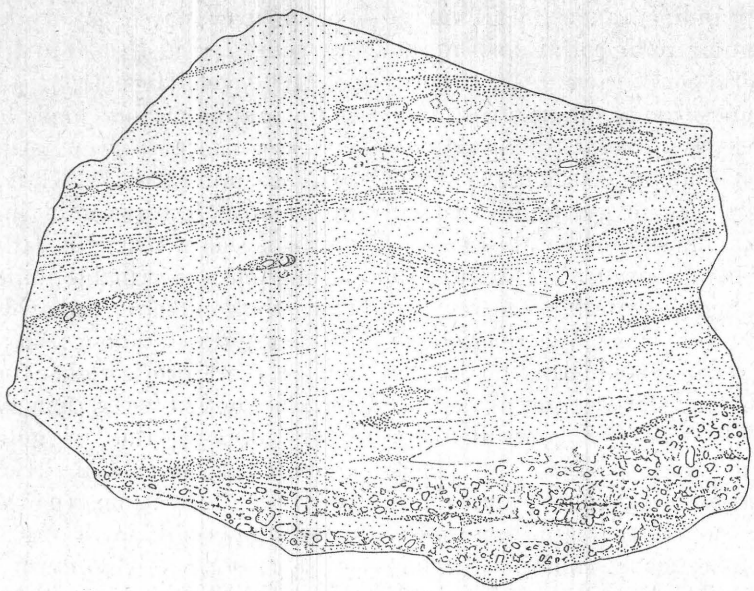

FIGURE 36.—Banded massive sulfide ore, Iron King mine. Dark bands are chiefly sphalerite; light bands, pyrite. Lenses are quartz-ankerite. Note angular discordance between banding in fine- and medium-grained massive sulfide (pyritic). Natural size.

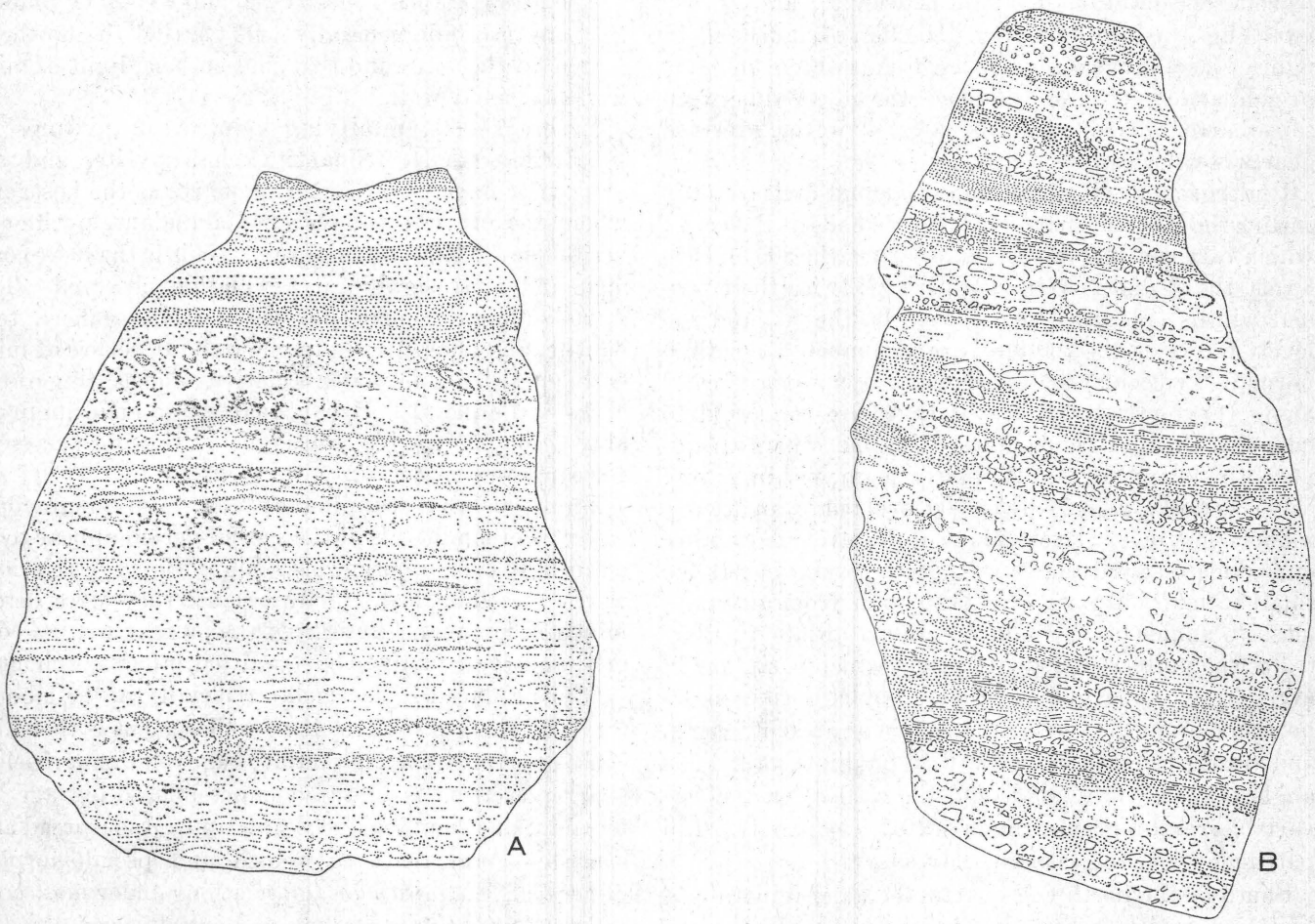

FIGURE 37.—Bands in massive sulfide ore, Iron King mine. *A*, Fine-grained well-banded massive sulfide ore. Dark bands are chiefly sphalerite; light bands, pyrite. Natural size. *B*, Low-grade banded massive sulfide, Iron King mine. Coarse texture is sulfide not intensely sheared, in contrast to sulfide ore in figure 36. Light bands and areas are chiefly pyrite. Dark bands are chiefly quartz-ankerite and some sphalerite. Natural size.

ular fracture networks. Banding ranges in all variations from regular to irregular, but generally not in the same vein or the same general zone in a vein. In the ore, banding is due chiefly to deposition of sphalerite in fractures and replacement of early minerals by sphalerite along microscopic fractures, but in vein gangue it results from bands relatively rich in pyrite alternating with bands rich in quartz, ankerite, or both. In some narrow veins, alternating bands conform in attitude to foliation in the wall rocks, and the banding is as regular and perfectly formed as the foliation. Where only one stage of mineral deposition occurred, banding in the veins may reflect variations in relative rates of deposition of different minerals, possibly controlled by foliation.

North ends of all the larger veins consist almost entirely of greenish to gray quartz that appears almost chalcedonic, plate 13*A*. The quartz contains disseminated idiomorphic pyrite crystals and is cut by ramifying sulfide veinlets and by massive white quartz. In places, massive quartz is rich in gold and silver, as in *G* vein, 800 level; *F*, 800 and 900 levels; and *D*, 600 level (figs. 31 and 32). Other bodies, as indicated in figure 33, contain average or less than average amounts of gold and silver. No difference was observed between some barren quartz and ore-quartz; however, greenish quartz is valueless.

Underground openings and diamond-drill records outline the size and shape of the north ends of the veins, which vary considerably in pattern. On the 800 and 900 levels, the *E* and *F* veins are perhaps twice their normal widths (pl. 12). The *D* vein on the 900 and 800 levels and *G* on the 700 level are also much wider than normal. Lenticular bodies of quartz occur sporadically along the footwall. Some of these are connected to quartz at the north ends of the veins, but others are apparently independent. A quartz mass within *I* vein connects to the quartz nose on 900 level but is apparently independent on 800 level. Maps of the levels show other quartz masses, and the veins contain several others too small to map on plate 12. Transition from quartz to massive sulfide is sharp and regular. Plate 12 illustrates the pattern of the transition, which is similar in every vein. The quartz-sulfide contact strikes more northerly than the vein, cuts the vein at an acute angle, and transgresses from footwall to hanging wall. The angle between the contact and the wall of the vein is variable, reaching a maximum of 30°. Commonly a thin parting of sericite lies along the contact.

South of the north-end quartz, the chief mineralized variation is an inverse relation between amounts of pyrite and of sphalerite, which partly replaces pyrite (pl. 13*D*). Where sphalerite is abundant, pyrite is proportionately sparse. Assay data illustrate these changes for veins *D* and *G*, 600 to 1,100 levels, and for *F*, 500 to 1,100 levels (figs. 31, 32, and 33). Variations shown in the graphs are due chiefly to sphalerite content, which increases sharply from the massive quartz southward to a maximum from which it decreases gradually. The information lost by combining lead and zinc in the diagrams is small, as lead usually is distributed uniformly throughout the minable limits of a vein. In *D* and *F* veins, the content of lead averages between 1.50 and 1.75 percent and in *G* about 2.50 percent. The ratio of lead to zinc, therefore, is higher for sections of the vein where the zinc content is less than average.

The dollar value of gold generally is consistent enough at twice that of silver to combine them for simplicity of illustration. In every vein the gold plus silver maximum occurs north of lead plus zinc, but not so far north that no overlap exists. The dollar value of the gold plus silver is high at the lead plus zinc maximum, but it is waning. This is a consistent feature of the deposit.

The assay graphs (figs. 31, 32, and 33) show plainly that any isograde generally will parallel the northern plunge of the veins and that the southern limit of mining is an "assay wall."

In ore shoots, ramifying veinlets of mixtures of sphalerite, galena, tennantite, chalcopyrite, and arsenopyrite have pyrite-ankerite-quartz as the host, and where the ore minerals are not abundant, pyrite-ankerite-quartz constitute the vein. This is the most common and most significant association observed. The introduction of sphalerite, arsenopyrite, galena, tennantite, and chalcopyrite in one general period of mineral deposition followed brecciation and shearing of the early vein minerals. Pyrite, quartz, and ankerite probably accompanied this later stage, as they also crosscut the early vein material.

From geologic mapping of veins and microscopic study of ore, minerals of the second period appear to be essentially contemporaneous. Significant age relations were not established, chiefly because they were determined in less than 1 percent of the pairs in mutual contact, and some of these contradicted others. Schouten (1934) produced replacement artificially in the laboratory and showed that selective replacement is the chief cause of error in determination of age relations; if selective replacement is admitted as possible, so few age determinations would not be significant. Widespread and consistent occurrence of veined and pseudomorphic minerals is diagnostic of age relations under most conditions, but such consistency was not observed.

The following more detailed descriptions of individual mineral species illustrate common associations and

occurrences of minerals. Inferred age relations are stated as a matter of record. Two types of sphalerite occur, rosin and brown. Brown sphalerite is earlier than the rosin, and in places the rosin variety cuts it. Brown sphalerite is most abundant, constituting more than 90 percent of the sphalerite. An analysis of the brown variety by the American Cyanamid Co. showed zinc 63.5 percent, iron 3.2 percent, and sulfur 33.3 percent, which corresponds to 95 percent zinc sulfide and 5 percent iron sulfide. Brown sphalerite occurs chiefly as streaks or bands of relatively pure mineral, which contributes greatly to banded structure of the veins (pl. 13*D*). In part, these bands represent replacement features, as residual pyrite grains are dispersed irregularly through them in various stages of replacement; whereas on either side of the veinlet, idiomorphic pyrite occurs (fig. 38*A*). Commonly brown sphalerite is interstitial to pyrite that shows little or no modification of crystal outlines (fig. 38*B*). Such a relationship could mean either (1) simultaneous deposition of pyrite and sphalerite, (2) selective replacement of early quartz-ankerite by sphalerite, or (3) replacement of sphalerite by pyrite. Sphalerite occurs as microscopic veinlets in quartz, ankerite, chalcopyrite, arsenopyrite, and galena. It is cut by veins of galena and is included as small grains in chalcopyrite and in pyrite. Idiomorphic arsenopyrite occurs in a matrix of sphalerite, but locally sphalerite embays the crystal outline of arsenopyrite. In places sphalerite replaces ankerite, retaining the rhombohedral angle in the outline of the grain. Parts of veins enriched in brown sphalerite are enriched also in tennantite, galena, and arsenopyrite.

Rosin sphalerite is in late crosscutting veinlets associated with clear quartz; to a lesser extent it occurs with ankerite and tennantite and as disseminated crystals whose age relations are obscure. Nests of rosin sphalerite, perched on comb quartz, occur rarely in late crosscutting fractures. Brown and rosin sphalerite occur together in a few crosscutting veinlets. Their age relations were not discerned in these veinlets, but rosin sphalerite only occurs in late crosscutting veinlets that are younger than the bulk of the brown sphalerite.

Galena rarely constitutes more than 4 percent of the vein, yet is in every polished specimen examined from productive veins. Galena and tennantite are associated closely and are generally in contact but no evidence of an age difference between them was seen. Galena, tennantite, and chalcopyrite also make a common associa-

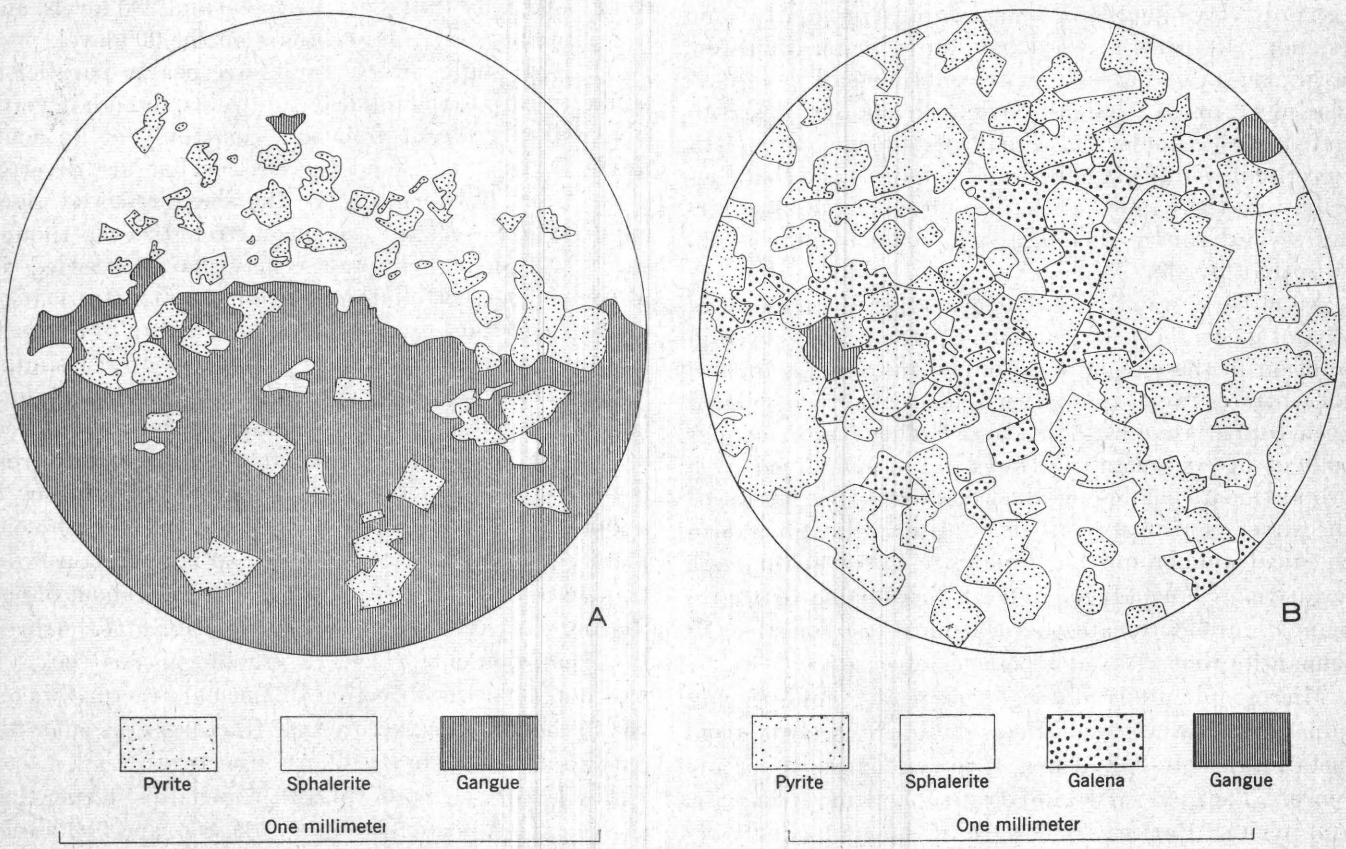

FIGURE 38.—Drawings of polished thin sections, Iron King mine. *A*, Illustrates character of pyrite on either side of contact of a band rich in sphalerite. *I* vein, 700 level. In upper part of field, sphalerite partly replaces pyrite; pyrite in gangue is unmodified. *B*, Sphalerite interstitial to pyrite that shows no modification of crystal form, *H* vein, 900 level.

tion. Galena is interstitial to pyrite and arsenopyrite; veins of galena cut sphalerite and arsenopyrite, and veins of sphalerite cut galena. In a few open fractures, galena, resin sphalerite, and pyrite are perched on comb quartz. Galena and associated tennantite are commonly in pyrite-free zones of quartz and ankerite that form part of the megascopic banded structure. They also occur as large grains at the ends of ankerite augen and, where more abundant, rim the augen completely. In a few places galena replaced pyrite, as indicated by small irregular enclosed residuals of pyrite. Had the pyrite been younger, it most certainly would be idiomorphic because the power of crystallization of pyrite is many times that of galena. The distribution and occurrence of tennantite is similar to that of galena.

Chalcopyrite, which occurs only in very small amounts in the productive veins is in small irregular independent grains; in small masses associated with tennantite; and in microscopic blebs in sphalerite. Microscopic veins of chalcopyrite cut pyrite, arsenopyrite, and quartz ankerite, and veins of sphalerite cut chalcopyrite.

Arsenopyrite is abundant in massive sulfide, particularly in the ore. It commonly has crystal outlines, chiefly diamond shaped in section, against all minerals except pyrite and other arsenopyrite. Apparently some arsenopyrite grew at the expense of pyrite, for a few larger crystals of arsenopyrite contain small irregular-shaped grains of residual pyrite. Arsenopyrite is associated with sphalerite, tennantite, galena, and chalcopyrite in veins that cut pyrite-rich veins and that contribute to the banded structure. In some areas covering several square inches, arsenopyrite is more abundant than pyrite.

No mineral was found that has silver as an essential part of its composition; the silver presumably is in solid solution in the galena and tennantite. Assays of mill concentrates reveal that silver is more closely associated with copper than with lead. Fluctuations in copper content of concentrates always are accompanied by a proportionate change in silver content. Hence most of the silver is probably in tennantite, although microchemical ammonium bichromate tests for silver were negative. Short (1940, p. 201) was unable to obtain ammonium bichromate microchemical test for silver in tennantite that assayed 1 percent silver.

Microscopic study did not reveal any gold-bearing minerals; however, sufficient data are known from metallurgical tests and assays to summarize its occurrence. The gold is free and occurs in galena, sphalerite, and pyrite. Pyrite carries most of the gold, chiefly because pyrite is more abundant. Figures 31, 32, and 33 indicate the distribution of gold in three veins. The ratio of precious metals to base metals is higher in the quartz at the north ends of veins than for the massive sulfide. The greatest concentrations of precious metals also occur in quartz "noses." Precious metals, like the base metals, gradually decrease in content southward.

Ankerite is the carbonate of the veins and hydrothermally altered rocks. It was identified through index of refraction, microchemical tests, and slow effervescence with dilute acid. Ankerite forming knots and vein-filling interstitial to sulfide minerals has an omega index of 1.710 indicating a composition of 65 percent dolomite ($CaMgC_2O_6$) and 35 percent ferrodolomite ($CaFeC_2O_6$) (Winchell, 1933, p. 74). Ankerite from a small quartz-ankerite-chlorite veinlet cutting andesitic tuffaceous rock south of the mine has an omega index of 1.725 indicating equal amounts of dolomite and ferrodolomite.

POST-MINERAL STRUCTURES

Younger structures consist of faults, joints, and probably fracture cleavage. Larger faults that offset veins are reverse strike faults. Those that dip steeper than veins produce an overlap; those that dip less steeply produce a gap. Other strike faults whose direction of relative displacement is not known also occur in the mine, especially in the northern 800 and 900 levels, and a nearly flat reverse fault occurs on the 900 level.

All high-angle reverse faults are nearly parallel in strike and dip to the foliation and veins, though in many places they crosscut foliation, commonly at an acute angle. The faults tend to crosscut but are diverted along folia, possibly because the shear resolved along the foliation exceeded the shear strength even though the stress on other planes may have exceeded it. Movement along foliation continued until stress at an angle to it became greater than the strength of the rock. Rupture then occurred across foliation. In a similar manner, faults were diverted for some distance along the wall of veins. In a scale-model experiment, Ekkernkamp (1939) found that planar structure at 20° from the direction that fractures would normally follow in a homogeneous material would control almost completely the direction of fracture. Where fractures did transect the planar structure, the direction about bisects the angle between the normal to foliation and the direction that would be taken in a homogeneous rock. It was not until the direction of normal fracture was at 75° to the planar structure that fractures crosscut without visible control by the planar structures.

The Iron King fault (pl. 12), which dips steeper than the veins, duplicated *I* vein on 700, 800, and 900 levels; on 1,000 level it is between *I* and *H* veins, which are separated by about 1 foot of gouge. This fault dies out

both northward and southward from a point located at 4,175 E.–4,040 N., 800 level, where the vertical separation is about 120 feet (pl. 12). Three reverse faults, having vertical separations of less than 50 feet, occur south of the Iron King fault. Two of these dip less steeply than the veins, producing small gaps in vein continuity, the other dips more steeply, overlapping the vein.

Faults on the north 800 and 900 levels, which have gouge zones about 1 foot wide, do not displace the *I* vein, although a small displacement would not have been noticeable at the stage of development in 1948. By analogy to the other faults of similar attitude, these probably are reverse faults.

Small faults that range in strike from N. 85° W. to N. 80° E. occur sporadically in some veins and dip southward at about 80°. These faults generally stop at the walls of veins, yet they contain gouge zones as much as 6 inches wide, composed of pulverized vein filling. Generally the faults occur as several closely spaced parallel breaks. In several places they extend into wall rock and have a small horizontal separation indicating that the north wall moved eastward relative to the south wall. Displacement on these faults is probably small, as they do not mark changes in grade or appearance of ore. Where the faults do not cut the wall rocks, they may represent minor adjustments to post-mineral deformation, wherein the vein moved slightly within the confines of its walls.

Flat joints, essentially at right angles to the dip of the veins, are abundant throughout all veins. They are spaced only a few feet apart vertically.

In the fine-grained well-foliated hydrothermal alteration zones in the Iron King area, fracture cleavage cuts the dominant foliation. The fracture cleavage has two distinct forms, (1) a series of V-shaped minor folds broken by a fracture that bisects the fold, and (2) sigmoid folds broken by fractures that bisect the folds.

As the attitude of the fracture cleavage is variable, 58 attitudes from outcrops scattered at random throughout the alteration zones were measured, plate 11. Only one of a particular direction of fracture was recorded at any one outcrop. Poles to planes of cleavage plotted on an equal area net revealed a 17 percent concentration at N. 50°–65° W., 80° SW., and a 12 percent concentration at N. 70°–80° E., 75° SE. The sigmoid folds indicate that the north sides of both the northwestward and northeastward-trending fracture cleavage moved southward relative to the south side. Thus the movement on the fracture cleavages is in the proper direction to be members of a conjugate set of shears.

Because the two directions of fracture cleavage are about symmetrical to the strike of veins, it might be assumed that the fracture cleavage was formed by the deformation that localized the fractures or shears now occupied by veins. But this is not correct. Two shear planes of different altitude, but related to the same stresses, must have the same deformation plane (ac plane of the structural coordinate system), and the deformation plane for the sheared zone localizing the veins does not coincide with that for the fracture cleavage. Structural coordinates for the sheared zones that localized the ore deposit are obtained by assuming that lineation (mineral streaking at 60° N.) in the walls of the veins and the parallel northward plunge of the veins is either *a* or *b*. Assuming the lineation is *a*, the deformation plane is nearly N. 58° W., 65° N.; assuming it is *b*, the deformation plane is about N. 80° E., 33° S.

The deformation plane for the fracture cleavage is based on the assumption that the intersection of the two slip planes delineated *b*, a valid assumption as indicated by Cloos (1946, p. 19), and others. Using the average attitude for the two directions of fracture cleavage, *b* plunges S. 20° E. at 75°, and the deformation plane, lying perpendicular to *b*, strikes N. 70° E. and dips 15° northward. The dissimilarity in attitude of deformation planes indicates that the fracture cleavage is probably postmineral in age, possibly owing to the deformation that produced postmineral faults in the ore deposit.

ORE DEPOSITS ALONG THE SILVER BELT–McCABE VEIN

The Silver Belt-McCabe vein was traced as a continuous zone for about 14,000 feet from a point about 1,500 feet west-northwest of the Iron King mine southwestward to the corner of the Jerome area (pl. 1). Northeastward, the structure disappears beneath the gravel of the Hickey formation in Lonesome Valley; and southwestward, the vein passes from the Jerome area in the quartz diorite. The Rebel mine, which is about a mile southwest of the McCabe mine, is reported to be on the structure. The strike ranges from N. 65° E. in the southern sector to N. 30° E. in the northern sector; the dip ranges from 70° NW. to 80° SE.; and the width ranges from 6 to 15 feet.

The vein is almost entirely within the breccia facies of the Spud Mountain volcanics. Chlorite and probably some sericite, lying with their basal sections essentially parallel to the strike of the vein, characterize the fissile, sheared zone comprising the vein; whereas the wall rocks, especially those on the hanging wall (west side), are foliated but not fissile, and are characterized by actinolitic hornblende.

The vein structure appears to have been mineralized only locally, and ore has been found only near the mines described in this section. The ore shoots occur as lenses

or pods that rarely, if at all, occupy the entire width of the vein.

The age of the deformation that produced the sheared zone cannot be definitely established. Because of the parallelism between the structures produced by the later periods of Precambrian deformation and the Silver Belt-McCabe vein, and because of the micaceous character of the sheared zone comprising the vein structure, a Precambrian age is most likely. The ore shoots, however, did not necessarily form penecontemporaneously with the deformation that produced the sheared zone.

The unoxidized sulfide ore minerals occur as coarsely crystalline, drusy masses that filled open spaces. This is in marked contrast to the massive sulfide veins in the Iron King deposit, which is believed to have formed nearly contemporaneously with intense deformation at considerable depth. The ore shoots along the Silver Belt-McCabe vein probably formed later than the last period of strong Precambrian deformation, and sufficiently later so that pressure conditions permitted open spaces, perhaps as much as several inches wide. This, however, does not necessarily preclude a Precambrian age for the mineralization.

The character of the ore along the strike of the Silver Belt-McCabe vein varies, suggesting lateral zoning. Silver and lead characterized the Silver Belt deposit, at the north end of the known zone. In the Arizona National mine, to the south of the Silver Belt, the content of lead and silver was lower; but the ore contained zinc and iron, chiefly as sphalerite and pyrite. The ore in the Lookout mine, to the south of the Arizona National, was complex and contained lead, zinc, iron, copper, silver, and gold. The silver content was lower than in the Arizona National, but the copper and gold content was higher. The McCabe-Gladstone mine is south of the Lookout; and like the Lookout, its ore was complex and contained copper, lead, zinc, iron mostly as sulfides, and gold and silver. In the McCabe-Gladstone the content of iron, copper, and gold was higher than that in the Lookout, but lower in lead and probably in silver.

SILVER BELT MINE

The old Silver Belt mine, which is about 5,500 feet southwest of the Iron King mine, has been inactive for many years, and the workings are all inaccessible. The following data are, in large part, abstracted from Lindgren (1926, p. 128–129). So far as is known, the Silver Belt is the oldest mine in the Jerome area. It was located about 1870, and worked from 1870 to 1880. The ore was freighted to Ehrenberg, and from there shipped down the Colorado River and to San Francisco for treatment. Later the ore was smelted at Humboldt. In 1906, the mine was reopened, and the shaft deepened by about 80 feet. But this later work did not produce much ore.

The property was explored by four shafts, one of which is reported to be 400 to 480 feet deep. The amount of lateral workings from the shafts is unknown.

Ore was mined from the surface to the 250 level, figure 39, where it occurred in ore shoots ranging from 3 inches to 3 feet in width, but the vein in this locality ranges from 8 to 12 feet in width. It is reported that the ore was chiefly oxidized, and contained much manganiferous ankerite and barite and minor amounts of galena and sphalerite. The oxidized material contained much silver and some lead. Elsing and Heineman (1936, p. 101) list the production from 1870 to 1880 as 1,000,000 pounds of lead, and silver valued at $300,000. Lindgren estimated the silver production at 300,000

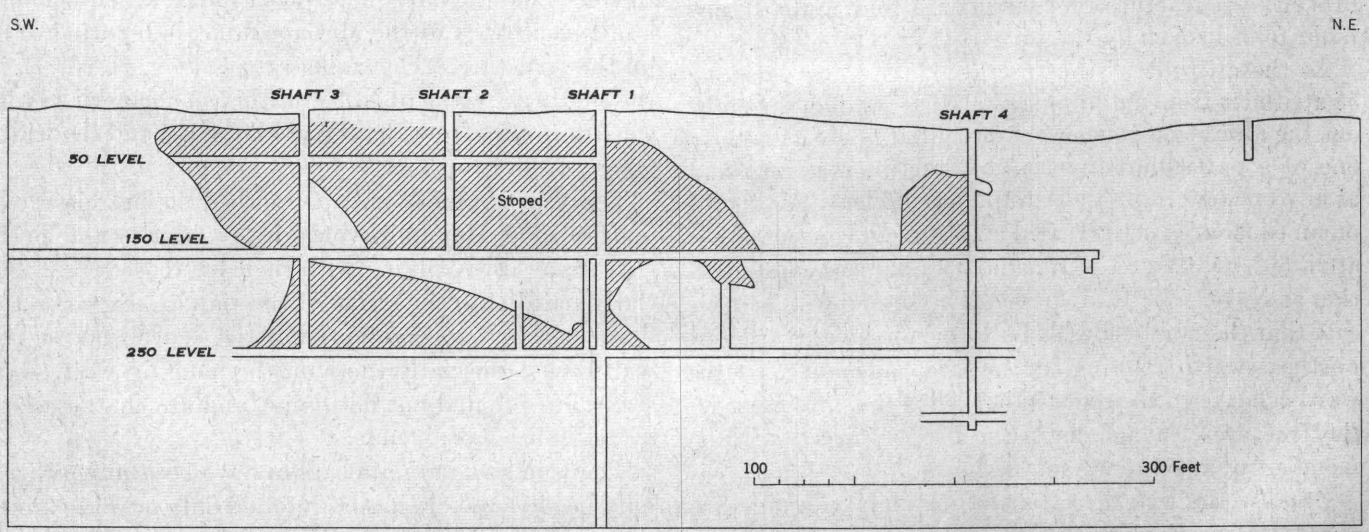

FIGURE 39.—Longitudinal projection of the Silver Belt mine showing stoped areas. Furnished by Fred Gibbs.

ounces, and local estimates were as high as 700,000 ounces.

ARIZONA NATIONAL MINE

The Arizona National mine is about half a mile southwest of the Silver Belt on the same vein. At the time of Lindgren's visit in the Jerome area in 1922, the Arizona National was operating, and had been since 1915. The property at that time was equipped with a 50-ton ball mill and concentrating tables. The mine at present is inaccessible.

The deposit was developed by two shafts, one about 520 feet deep, and by lateral workings at seven levels extending off the shaft (fig. 40). The lateral workings are separated vertically from 40 to 110 feet. Ore was mined between the surface and the level 7, which is about 500 feet below the surface (fig. 40).

The occurrence of the ore is essentially the same as at the Silver Belt mine. The vein strikes N. 20° E., and dips 70° W., and ranges in width from a few inches to several feet. Lindgren (1926, p. 129) reported, "The ore contains a little drusy milky-white quartz, some calcite, much pale brown ankeritic carbonate, with manganese and some barite." Galena and sphalerite comprise the megascopic ore minerals but include minor amounts of pyrite and chalcopyrite. Microscopic argentite and tetrahedrite were disseminated throughout the galena.

Data on production of the Arizona National mine were collected from several sources, which accounts for the overlap in time. Presumably, the data are accurate for the periods designated, but where there is an overlap, the data cannot be added to the total production of the deposit. Elsing and Heineman (1936, p. 101) report that between 1921 and 1926 the deposit yielded 1,000,000 pounds of lead, and silver valued at $60,000. The following production data were furnished by Fred Gibbs, Prescott, Ariz. From October 1915 to June 1923, F. M. Anderson recovered 2,068 tons of concentrate that averaged 103 ounces of silver per ton, 26.9 percent lead, 11.9 percent zinc, about 0.3 percent copper, and about 0.03–0.02 ounce of gold per ton. Production from October 1922 to January 1925 was about 3,600 tons of ore that yielded 70,850 ounces of silver and 228,713 pounds of lead. The grade of this ore was about 20 ounces per ton of silver and about 3 percent lead. Presumably some zinc and a small amount of copper were also present, although they were not mentioned. From October 1928 to February 1931 the Double O Metals Co. produced 377 tons of concentrate that averaged 114 ounces per ton of silver, 32 percent lead, 10 percent zinc, 0.5–1 percent copper, and about 0.03 ounce per ton of gold. So far as known, the Arizona National has been inactive since 1931.

428436—58——12

LOOKOUT MINE

This mine is on the Silver Belt-McCabe vein about 1,000 feet southwest of the Arizona National. Little is known of the history, development, and production of the Lookout. It was in operation in 1922 when Lindgren visited the mine, but presumably was closed shortly thereafter. The property was reopened for a short period during 1948 and 1949; and at that time a few tons of high-grade ore—consisting chiefly of galena, sphalerite, and pyrite—was piled on the dump. In 1952 there was no activity and it is not known if the underground openings were accessible.

The ore is a complex deposit of sulfide minerals, chiefly galena, sphalerite, pyrite, and chalcopyrite, and also contains a little gold—perhaps one-eighteenth to one-twelfth ounce per ton—and some silver. The silver content is reported to be less than that at the Arizona National. So far as known, there has been little production from the Lookout. Lindgren (1926, p. 130) reported that a mill test on 180 tons of ore yielded 8 tons of concentrate valued at $80 per ton at the current metal prices of that time.

McCABE-GLADSTONE MINE

The McCabe-Gladstone mine is in Galena Gulch about 1.2 miles southwest of the Lookout mine. It is one of the deepest mines and has produced more ore than any other mine on the vein system. The mine was operated from 1898 to 1913, but has not been active since.

The property consists of two mines, the McCabe on the east, and the Gladstone on the west. It was developed by two shafts 800 feet apart: the Gladstone, 1,100 feet deep; and the McCabe, 900 feet deep. The lateral workings extend along the vein from 200 to 600 feet on every 100-foot level, the total exploration aggregating several miles (fig. 41).

The vein strikes N. 55° E. and dips 80° SE.; the width averages 3½ feet. Lindgren (1926, p. 130–132) reported, "there are five shoots along the vein, each of which has a stope length of 200 to 500 feet. At least two of the shoots appear to reach the 1,100-foot level in the Gladstone mine. These shoots pitch steeply toward the west—and average less than a foot wide." Jaggar and Palache (1905, p. 10) say:

> "the vein is a series of lenses which are characterized by band and ribbon structure, the metallic contents being largely confined to the center of the vein. Open vugs lined with large crystals of quartz and arsenopyrite are common. Arsenopyrite with pyrite and chalcopyrite carry the values, which are largely gold with some silver. Galena is sparingly present."

The gold to silver ratio is inverse to that of the mines northward on this vein (table 26).

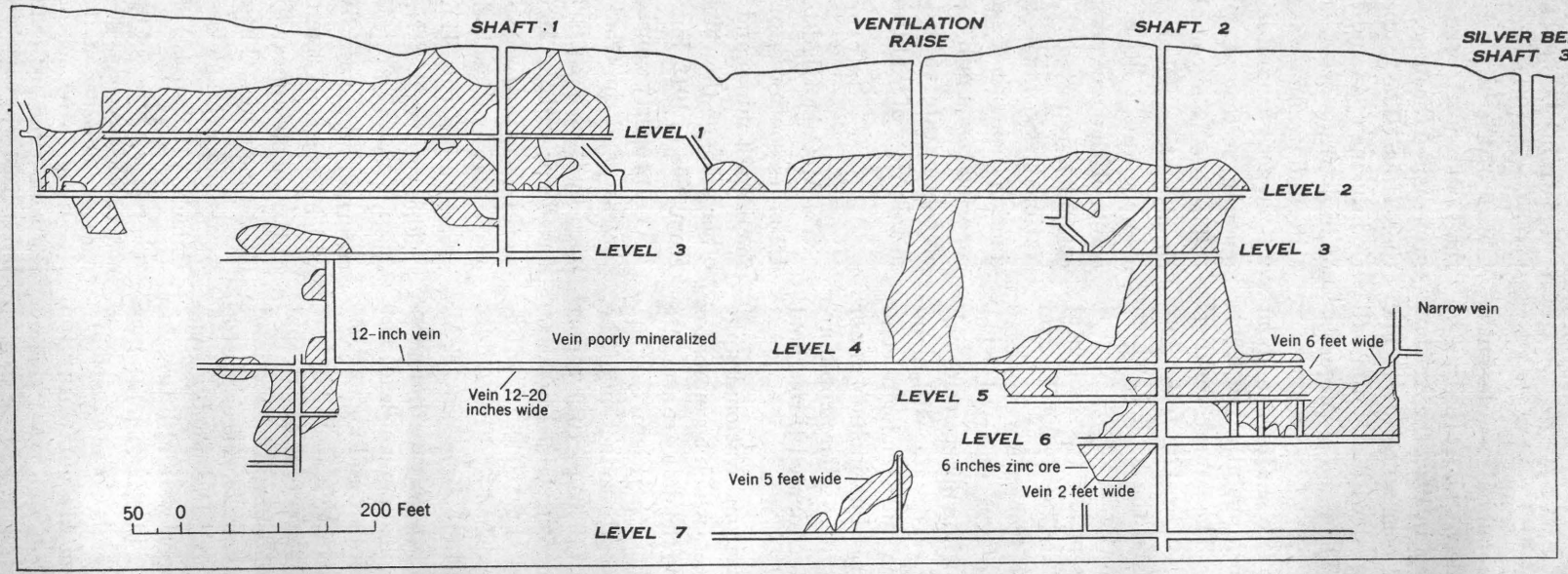

FIGURE 40.—Longitudinal projection of the Arizona National mine showing stoped areas. Furnished by Fred Gibbs.

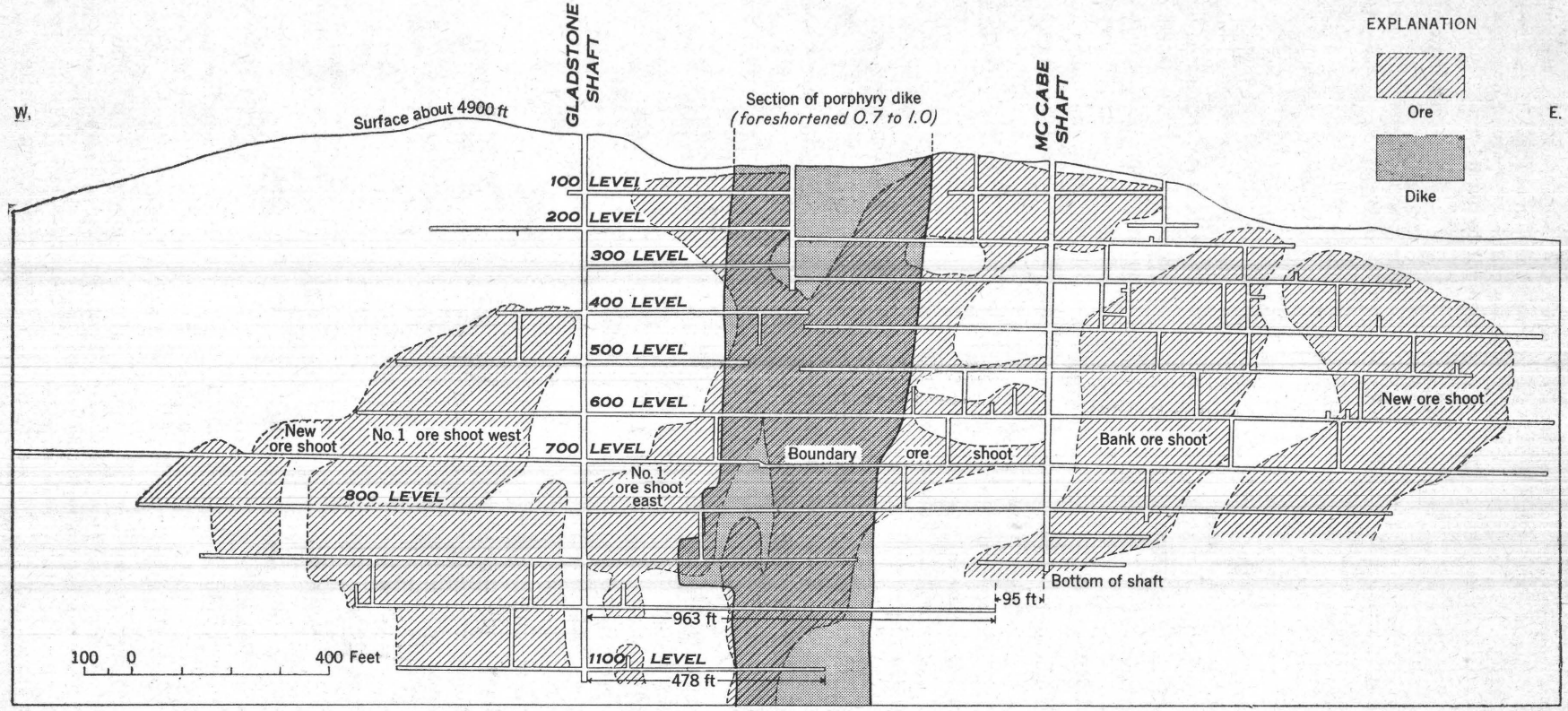

FIGURE 41.—Longitudinal projection of the McCabe-Gladstone mine showing stoped areas. After Lindgren (1926, p. 131).

TABLE 26.—*Average analysis of shipping ore and concentrates, McCabe-Gladstone mine. From Lindgren (1926, p. 132)*

Silica	percent	31.4
Copper	do	2.0
Lead	do	2.1
Zinc	do	4.7
Iron	do	24.6
Arsenic	do	3.9
Antimony	do	1.0
Sulfur	do	20.4
Gold	ounces per ton	1.6
Silver	do	10.2

The approximate production of the McCabe-Gladstone mine from 1880–1926 (Elsing and Heineman, 1936, p. 101) was 1,200,000 pounds of copper, 500,000 pounds of lead, $2,200,000 in gold, and $600,000 in silver. The total production was valued at $3,000,000.

KIT CARSON VEIN

The Kit Carson vein is about 1,500 feet west of the Silver Belt-McCabe vein. It strikes parallel to the Silver Belt-McCabe vein, but has an opposite dip to the east. The Kit Carson vein was traced on the surface for about 4,000 feet, but is somewhat longer, for its northern extent is overlain by the gravel of the Hickey formation in Lonesome Valley. A short segment near 1,268,000 N.–388,500 E. may be the southward continuation of the vein. Little is known about the Kit Carson vein. It appears to resemble the Silver Belt-McCabe vein in structure, alteration, and character of the mineralized zone.

The vein consists of a sheared zone as much as 5 feet wide, characterized by fissile, sericitic rock in which local stringers of comb quartz and boxwork, possibly after ankerite, were observed. The vein was prospected by eight shafts and prospect pits. The material on the dumps of these openings contains considerable ankerite, but no sulfide minerals were found. The size of some of the dumps indicates that these shafts are 100 or more feet deep, or that short lateral workings extend off shallow shafts. Production from the Kit Carson vein has not been recorded.

CHERRY CREEK MINING DISTRICT

By R. E. LEHNER

The Cherry Creek mining district occupies the basin drained by Cherry Creek in the southeast corner of the Jerome area (pl. 1). Mining of gold ore has been intermittent since about 1880. Lindgren (1926, p. 104) made the following statement about the district:

> The district contains many properties that have made some production, and some of them were in operation at the time of visit. Many of them appear to have a certain resemblance in their history. There was the discovery, the arrastre stage, the sinking of a shaft to a depth of 200 or 300 feet, followed by the erection of a small mill, and next usually a prolonged rest, with a watchman in charge. The pockety character of the ore is the cause of this stoppage of exploration. Whether any large ore shoots will be found is probably doubtful.

PRODUCTION

Economic conditions in large part have controlled activity in the district. From 1917 to 1920, a period of economic prosperity, no mines were active; but from 1930 to 1934, activity is related to the economic depression of that period.

Most of the mines in the district were visited briefly, but many were not accessible for examination. Hugh Allen of Cherry, Ariz., a life-long resident of the district, has generously supplied information on the past history, development, and production of the mines.

Total production of the district is not known, but all available information indicates that it is not large. Table 27 gives the best summary available on the number of producers and production from 1908 to 1933. The Cherry Creek district has yielded less than 1 percent of the lode-gold production of Arizona (Wilson, 1942).

TABLE 27.—*Production from Cherry Creek mining district, 1908–33*

[From Elsing and Heineman (1936, p. 83)]

Year	Producers	Tons	Gold (value)	Silver (ounces)	Copper (pounds)	Total
1908	6	464	$5,775	86		$5,821
1909	4	330	7,646	242		7,772
1910	6	1,332	6,352	93	394	6,452
1911	4	531				9,402
1912	3					
1913	2					
1914	4					2,866
1915	5	86				958
1916	2					
1917						
1918						
1919						
1920						
1921	1					
1922						
1923	1					
1924						
1925	2					
1926	2					
1927						
1928						
1929						
1930	3	201	1,897	74	4,155	2,465
1931	1	40				
1932	6	223	3,023	96	968	3,111
1933	14	327	9,214	423	1,969	9,488
Total		3,524	$33,907	1,014	7,484	$48,335

GEOLOGY

The country rock in the Cherry Creek mining district is Precambrian quartz diorite cut by a few granodiorite porphyry dikes (pl. 1). Tapeats sandstone(?) and Martin limestone cap some of the ridges and peaks. The Hickey formation, comprising lava and gravel, in part overlies the Paleozoic rocks and in part rests directly on the quartz diorite.

The veins are abundant but discontinuous, and generally are uniform in character. They occur as lenses or pods in sheared zones of quartz diorite and granodiorite porphyry. Their width ranges from 1 inch or less to 6 feet, but average about 1–2 feet. Where the strike changes or the dip flattens the thicknesses appear to be above average. The veins range in strike from north to N. 45° E., but most lie between N. 15° E. and N. 35° E.

The vein filling is chiefly milky-white quartz, but in a few places is stained on weathered surfaces by limonite. Locally it contains vugs, lined with quartz crystals. Minor constituents are gold, tourmaline, pyrite, and, according to Lindgren (1926, p. 103), small quantities of chalcopyrite, bornite, sphalerite, and galena. The tourmaline occurs as minute needles in the quartz, and the pyrite as irregular grains intergrown with quartz. Some gold is free and may be megascopic, but some is derived from oxidized sulfide minerals. Gold and sulfide minerals are associated in most of the ore shoots. This spatial coincidence suggests contemporaneous deposition.

Adjacent to the quartz veins, limonite has stained the quartz diorite and granodiorite porphyry dikes orange and rusty brown. Mafic minerals are absent, and the plagioclase is sericitized and altered to albite. Farther from the veins, chlorite has replaced biotite and hornblende, the plagioclase is saussuritized, and epidote veins are locally present.

The limitation of the quartz veins and alteration zones to the quartz diorite and granodiorite porphyry dikes and their absence in the overlying Paleozoic sedimentary rocks proves that the gold-bearing quartz veins in the Cherry Creek district are of Precambrian age.

LEGHORN MINE

One of the northernmost mines in the district, this mine is developed by an inclined shaft 600 feet deep, but was inaccessible in 1951. The vein cannot be traced at the surface, but at the portal of the adit, the vein strikes N. 65° E. and dips 25° W. According to reports most of the ore came from the hanging-wall side of the vein which is stoped from the 400 level to near the surface for several hundred feet. Production from the Leghorn is reported to be among the largest in the district.

SITTING BULL MINE

The Sitting Bull mine is about 300 feet east of the Leghorn mine at a higher altitude, and both mines may be on the same vein or on parallel veins of the same system. Three adits within a vertical range of about 60 feet crosscut the vein and some ore has been mined from underhand and shrinkage stopes. The underground workings are within the oxidized zone of the vein. The width of the vein ranges from small stringers to 2½ feet; the average strike is about N. 40° E. and the dip ranges from 25° to 45° W. The ore shoots occurred in the more gently dipping parts of the vein. About $11,000 of gold was produced in 1940, and an unknown amount was produced earlier.

FEDERAL MINE

The Federal mine is about a mile southeast of the Leghorn mine, and the vein, averaging 2 feet in width, strikes N. 45° E. and dips 50° W. Lindgren (1926, p. 107) states that the mine was active about 1907, and was developed by an inclined shaft, 260 feet deep. An exploratory adit 1,000 feet long was driven, and a mill costing $100,000 was built, but no ore was found.

GOLD BULLION MINE

The Gold Bullion mine is about a mile northwest of Cherry. Trenching and bulldozing have exposed about 350 feet of vein striking N. 15° E. and dipping 45° W. The underground workings, filled with water in 1951, consist of an inclined shaft 100 feet deep connecting with extensive drifts along the vein. Gold ore was shipped to the smelter at Clarkdale, but the amount of production is not known.

BUNKER MINE

Extensive underground workings were driven in an attempt to find ore shoots in a shear zone that is essentially barren. One small ore shoot, 3 feet wide, was found at the end of one of the exploratory drifts. The surface was scraped by mule teams in the early history of the district, and it is claimed that $100,000 worth of gold was produced about 1880. The mine, which is about 1½ miles north of Cherry, was reported to be one of the three largest producers in the district.

SUGAR BOWL MINE

The Sugar Bowl mine is about a mile north of Cherry, on the road leading to the Bunker mine. A shaft 30 feet deep connects with a drift along the vein, from which some ore has been stoped. An adit crosscuts the vein 30 feet from the portal, and some ore was mined from connecting drifts. The vein, ranging from 2 to 18 inches in width, strikes N. 25°–45° E. and dips from 25° to 45° W. It is reported that about 20 carloads of gold ore ranging in value from $66 to $88 per ton have been shipped.

GOLDEN IDOL MINE

The Golden Idol mine is more than a mile northeast of Cherry. An inclined shaft 400 feet deep was sunk on one of the veins on the property, but was inacces-

sible in 1951. The vein that was mined is about 18 inches wide, and strikes N. 35° E. and dips 45° W. Lindgren (1926, p. 106) states that a stamp mill and cyanide plant were on the property in 1922, but they were operated only from 1907 to 1910, on ore worth $7 to $12 to the ton. It is reported that the best ore was near the surface.

BLACK HAWK MINE

The old workings of the Black Hawk mine are 1 mile northeast of Cherry. An inclined shaft 200 feet deep was sunk on a vein striking N. 45° E. and dipping 55° W. The shaft is now filled with water. It has been claimed that 30 cars of gold ore averaging $25 per ton were shipped.

GOLD EAGLE MINE

The Gold Eagle mine is about 2 miles northeast of Cherry. An inclined shaft 100 feet deep connects with drifts along a vein striking N. 35° E. and dipping 5°–35° W. This mine has produced about 10 to 15 cars of ore according to unsubstantiated reports; some was reduced in arrastres and some was shipped.

WOMBACHER MINE

The Wombacher mine which is about 1½ miles east of Cherry, consists of a shaft 80 feet deep that connects with short drifts at the bottom. At a depth of 30 feet, a 10-foot crosscut connects to the vein, and below this level, the shaft follows the vein downward. The shaft cut through an ore shoot but the drifts were in barren rock. The mine is reported to have produced between $5,000 and $10,000 in gold ore; some ore was reduced in arrastres and some was shipped.

DOVE MINE

This mine which is about 100 feet west of the Wombacher mine, in 1951 consisted of a vertical shaft 70 feet deep, sunk in the vein. The operators planned to sink the shaft below the level of the Wombacher workings and run a drift eastward on the vein, which strikes N. 35° E. and dips 70° W. to vertical. This is one of the steepest veins in the district.

SUNNYBROOK MINE

The Sunnybrook mine, which is less than a mile west-southwest of Cherry, consists of an incline shaft sunk on the vein. The depth of the shaft is unknown. About 65 feet below the collar, short drifts connect to the shaft, and in 1951 ground water stood at the drift level. The vein strikes N. 15° W. and dips 55° W.; the average width is 18 inches. The vein wedges out at the face of the drift southwest of the shaft, and the gold content and the dip of the vein are reported to decrease with depth.

LOGAN MINE

This mine, 2 miles southeast of Cherry, is one of the deepest in the district. An inclined shaft on the vein is 600 feet deep and connects to three levels; at 60, 160, and 400 feet below the surface. The shaft was filled with water in 1951 to within 50 feet of the surface. Ore was not found below the 400 level. The vein ranges from 1 to 5 feet in width and strikes N. 40° W. and dips 35° W.; the northwest strike is unique in the district.

Gold worth about $14,000 was mined from the 160 level and smelter return sheets show that the ore averaged about $30 per ton. A northwest drift 350 feet long on the 400 level produced ore worth about $25 per ton 175 feet from the shaft and near the end of the drift. Lindgren (1926, p. 107) stated that a mill was on the property in 1922, but it was gone in 1951.

BLACK HILLS MINING DISTRICT

This mining district is on the western slope of the Black Hills and extends eastward along the south margin of Mingus Mountain to near Cherry. Many scattered prospects are in this district, but only one mine, the Yaeger, has had any appreciable production. Small showings of copper occur in many parts of the district, but most prospecting has been disappointing. The district includes much of the quartz diorite in the southern part of the Bingus Mountain quadrangle and much of the Ash Creek and Alder groups. The age of mineral deposition is presumably Precambrian, because of the general similarity to the veins in the Cherry district that are definitely of Precambrian age. Although mineral deposition in the Yaeger mine cannot be proven positively to be of Precambrian age, the absence of mineralized faults and fractures in the nearby Paleozoic rocks is strongly indicative.

YAEGER MINE

The first production from this mine which is located about a mile south of the Prescott-Jerome highway next to the Shylock fault apparently was in 1890. According to Elsing and Heineman (1936, p. 102), 10,000,000 pounds of copper, $52,000 of gold and $50 of silver, having a total value of $1,500,000, were produced from 1890 to 1922. Much of this production must have been before 1904, as subsequent production from the Black Hills district totals less than $300,000. The mine has been inactive since 1923, except for a few attempts to mill some of the dump ore.

The mine is developed by an inclined shaft to the 1,300 level, with drifts extending mostly to the east for a maximum of 750 feet. The mine is now inaccessible.

The Yaeger mine is in a lithic tuff member of the

Grapevine Gulch formation cut by a few narrow diorite porphyry dikes. The vein strikes east and dips 35° S. Apparently, it does not extend westward from the Shylock fault. According to Lindgren (1926, p. 98), the vein locally is 7 feet wide, containing 3 feet of solid bornite. The ore generally consists of calcite, quartz, bornite, and tennantite, and a little pyrite. Secondary chalcocite, azurite, and malachite are also present. Large aggregates of bornite and tennantite are common. Near the surface, oxide copper ore is present, but no well-defined chalcocite zone was found. The ore was reported to contain about 0.65 ounce of silver to the ton and 1 percent of copper, and was chiefly high grade. The ore shoot raked to the east, beginning near the collar of the shaft, and apparently ranges from 200 to 300 feet in length on the middle levels but becomes small on the 1,300 level.

SHYLOCK MINE

The Shylock mine is along the western front of the Black Hills about 3 miles south of the Yaeger mine. The Shylock fault is west of the shaft. It is not known whether shipments of ore have been made, but the mine is explored by an inclined shaft reported to be 1,053 feet deep with drifts totaling 2,000 feet. The country rock is the bedded tuffaceous member of the Grapevine Gulch formation. The shaft was sunk on a vein striking east-northeast and dipping 60° S. The vein was not found west of the Shylock fault.

The gangue in the vein is chiefly massive white quartz, but includes some ankerite, and tetrahedrite, galena, and sphalerite irregularly distributed through it. The ore is in part oxidized, and cinnabar has been recognized in the oxide zone. According to Lindgren (1926, p. 100), decomposition of mercurial tetrahedrite probably resulted in the deposition of the cinnabar, which occurs as coatings, and fracture fillings. On the two upper levels, the width of the quartz vein reached 5 to 6 feet. Benedict (quoted by Lindgren, 1926, p. 100) believes that the deposit was essentially a quartz lens that pinched out in both directions from the shaft.

BRINDLE PUP MINE

The Brindle Pup mine is about midway between Black Canyon and Ash Creek, south of Mingus Mountain. Two small prospect pits were sunk in Buzzard rhyolite beneath a capping of Tapeats sandstone(?). The main workings, however, are in the wide granodiorite porphyry dike that cuts the Deception and Buzzard rhyolite masses (1,327,500 N.; 441,800 E.). The shaft is caved and no information is available as to the extent of the underground workings. On the dump, quartz veins in granodiorite porphyry fragments contain streaks and aggregates of pyrite. Lindgren reports (1926, p. 100) that the ore is massive, and contains quartz, ankerite, pyrite, galena, and sphalerite.

COBALT PROSPECT

A small cobalt prospect is located about half a mile southeast from the Shylock mine in a tributary to Grapevine Gulch (1,315,000 N.; 422,000 E.). The country rock is a small patch of gabbro separating the granodiorite to the south from Grapevine Gulch formation to the north. An opencut 28 feet long leads to an 18-foot adit. A 20-foot winze was sunk at the portal which exposed a small vein striking N. 30° E. and dipping 60°–70° W. The vein ranges in width from a knife-edge to 14 inches, and is about 15 feet long. It pinches out before reaching the face of the adit, and is not exposed on the southwest side of the winze. The vein material is partly oxidized altered gabbro containing relics of sulfide, possibly cobaltiferous arsenopyrite. The vein in the adit is covered with erythrite (cobalt bloom).

PROSPECTING OPERATIONS IN THE BLACK HILLS DISTRICT

The Black Hills district may be divided into two areas. Many prospect pits, shafts, and cuts have been made in a belt along the Shylock fault, both within the Alder group and in the granodiorite, particularly where it is sheared. From the Shylock fault eastward, to the Cherry district, prospecting operations have been widely scattered in the granodiorite and Ash Creek group.

Along the Shylock fault zone, much of the prospecting has been along quartz veins ranging from 1 inch to 2 feet in width. The average width is about 3–4 inches. The trend of most of these veins ranges from N. 20° W. to north, but a few trend northeastward and eastward. Many of the veins contain some carbonate minerals (dolomite?, ankerite?), and surface croppings contain scattered films of malachite and azurite. Very little galena, pyrite, and chalcopyrite were observed on some dumps from the deeper shafts.

Where these narrow quartz veins cut the purple slate member of the Texas Gulch formation, the trend of the veins truncates the strike of the foliation, and bleached zones as much as 3 feet wide parallel the veins.

On the ridge east of the Yaeger mine in the Grapevine Gulch formation, some prospecting was done in 1946 along quartz veins striking north-south to N. 20° E. Galena, pyrite, bornite, anglesite, and azurite are exposed in shafts and on the dumps. An adit was driven 240 feet to the west without intersecting the vein

zone, and it is reported that a diamond-drill hole from the face of the adit cut through 40 feet of low-grade mineralized rock.

Silicified zones produce prominent ribs that stand out above the adjacent rocks in the Alder group west of the Shylock fault (pl. 1). Some of the silicified rock forms irregular blebs, but most of the zones are linear and are arranged en echelon or branch. They cut across lithologic contacts, thus proving that they are not quartzite interbeds. The width of the silicified ribs ranges from a few feet to more than 100 feet. Most of the silicified rocks are dark from a coating of iron and manganese oxides. Some probably contain magnetite as the compass needle is deflected near them. In places, the quartz is white and barren. Several prospect holes have been driven into these silicified zones with unsatisfactory results. A deep shaft was sunk on one of the silicified ribs, and a fair-sized dump indicates lateral workings (1,301,800 N.; 415,500 E.). Most of the rock on the dump is gray slate from the adjacent Texas Gulch formation, but some quartz fragments contain scattered pyrite coated with chalcocite.

The scattered prospect pits and shafts east of the Shylock fault zone are confined chiefly to northeastward-trending orange-brown shear zones in the granodiorite. The quartz veins in the shear zones generally parallel the shear zones, but locally some strike northwest. The dips of the veins are generally moderate, that is from 30° to 60°. Ankerite and (or) dolomite are common accessory minerals in the gangue. The veins are commonly stained red and brown from iron oxide, and malachite, azurite, and chrysocolla films are present. On some of the dumps, limonite pseudomorphs after pyrite were observed. The dump from a shaft (pl. 1), 1,293,400 N.; 432,000 E., has fragments of quartz-ankerite vein material containing bornite, as well as the carbonates of copper.

A few prospect shafts have been sunk along quartz veins in the Ash Creek group north of the granodiorite. Quartz, copper stains, and a little chalcopyrite are found on the dumps of some of these prospects.

Molybdenite was observed in a quartz vein in Burnt Canyon, half a mile northwest of the Brindle Pup mine (1,327,800 N.; 438,700 E.). The country rock is a small body of granodiorite porphyry, and the vein strikes N. 20° W. and dips 70° E. The width of the vein is about 1 foot. Scattered molybdenite crystals and coatings of ferrimolybdite, malachite, and limonite are present. A short adit was driven to explore the vein, but the grade of molybdenite is too low to encourage additional prospecting.

LITERATURE CITED

Anderson, C. A., 1927, Voltaite from Jerome, Ariz.: Am. Mineralogist, v. 12, p. 287–290.

——— 1933, The Tuscan formation of northern California, with a discussion concerning the origin of volcanic breccias: Calif. Univ., Dept. Geol. Sci. Bull., v. 23, p. 215–276.

——— 1951, Older Precambrian structure in Arizona: Geol. Soc. America Bull., v. 62, p. 1331–1346.

Babenroth, D. L., and Strahler, A. N., 1945, Geomorphology and structure of the east Kaibab monocline, Arizona and Utah: Geol. Soc. America Bull., v. 56, p. 107–150.

Barker, L. M., 1930, Concentrating plant of United Verde Copper Co.: Min. Cong. Jour., v. 16, p. 363–367.

Bartlett, Katharine, 1942, Notes upon the routes of Espejo and Farfan to the mines in the sixteenth century: N. Mex. Hist. Rev., p. 21–36.

Bateman, A. M., 1927, Ore deposits of the Rio Tinto (Huelva) district, Spain: Econ. Geology, v. 22, no. 6, p. 569–614.

Benson, Lyman, and Darrow, R. A., 1944, A manual of southwestern desert trees and shrubs: Ariz. Univ. Biol. Sci. Bull. no. 6, 411 p.

Billings, M. P., 1942, Structural geology: New York, Prentice-Hall, Inc., 473 p.

——— 1950, Stratigraphy and the study of metamorphic rocks: Geol. Soc. America Bull., v. 61, p. 436–448.

Blandy, J. F., 1883, The mining region around Prescott, Ariz.: Am. Inst. Min. Metall. Eng. Trans., v. 11, p. 286–291.

Bramlette, M. N., 1946, The Monterey formation of California and the origin of its siliceous rocks: U. S. Geol. Survey Prof. Paper 212, 57 p.

Brownell, G. M., and Kinkel, A. R., 1935, The Flin Flon mine, geology and paragenesis of the ore deposit: Canadian Inst. Min. Metall. Trans., v. 38, p. 261–286.

Butler, B. S., and Wilson, E. D., 1938, General features of some Arizona ore deposits: Ariz. Bur. Mines, Bull. 145, p. 9–25.

Campbell, Ian, and Maxson, J. H., 1938, Geological studies of the Archean rocks at Grand Canyon, Ariz.: Carnegie Inst. Washington, Year Book 37, p. 359–364.

Capps, S. R., and Johnson, B. L., 1915, The Ellamar district, Alaska: U. S. Geol. Survey Bull. 605, 125 p.

Childs, O. E., 1948, Geomorphology of the valley of the Little Colorado River, Ariz.: Geol. Soc. America Bull., v. 59, p. 353–388.

Clarke, F. W., 1924, Data of geochemistry: U. S. Geol. Survey Bull. 770, 841 p.

Clarke, F. W., and Washington, H. S., 1924, The composition of the earth's crust: U. S. Geol. Survey Prof. Paper 127, 117 p.

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

Cloos, H. 1930, Zur Experimentellen Tektonik: Die Naturwissenschaften, Jhg. 18, p. 714–747.

Cooke, H. C., 1947, The Canadian shield *in* Geology and economic minerals of Canada: 3d ed., Canada Geol. Survey Econ. Geology ser. 1, p. 11–97.

Cowley, J. F., and Quayle, T. W., 1930, Development and mining methods, United Verde Copper Co.: Min. Cong. Jour., v. 16, p. 313–325.

Creasey, S. C., 1950, Iron King mine, Yavapai County, Ariz., *in* Arizona zinc and lead deposits: Ariz. Bur. Mines Bull. 156, p. 112–122.

——— 1952, Geology of the Iron King mine, Yavapai County, Ariz.: Econ. Geology, v. 47, p. 24–56.

Daly, R. A., 1933, Igneous rocks and the depths of the earth: New York, McGraw-Hill, 598 p.

D'Arcy, R. L., 1930, Mining practice and methods at the United Verde Extension Mining Co., Jerome, Ariz.: U. S. Bur. Mines Inf. Circ. 6250, 12 p.

Darton, N. H., 1910, A reconnaissance of parts of northwestern New Mexico and northern Arizona: U. S. Geol. Survey Bull. 435, 88 p.

——— 1925, A résumé of Arizona geology: Ariz. Bur. Mines Bull. 119, 298 p.

Darton, N. H., and others, 1924, Geologic map of the State of Arizona: Prepared by the Arizona Bureau of Mines in cooperation with the U. S. Geological Survey.

Davis, E. F., 1918, The radiolarian cherts of the Franciscan group: Calif. Univ. Dept., Geol. Sci. Bull., v. 11, p. 235–432.

Denny, F. W., 1930, Cottrell plant of the United Verde Copper Co.: Min. Cong. Jour., v. 16, p. 370–375.

Dewey, H., and Flett, J. S., 1911, On some British pillow-lavas and the rocks associated with them: Geol. Mag., v. 8, p. 202–209.

Ekkernkamp, M., 1939, Zum Problem der altern Anlagen in Bruchgebieten: Geol. Rundschau 30, p. 713–764.

Elsing, M. J., and Heineman, R. E. S., 1936, Arizona metal production: Ariz. Bur. Mines Bull. 140, 112 p.

Fearing, J. L., Jr., 1926, Some notes on the geology of the Jerome district, Ariz.: Econ. Geology, v. 21, p. 757–773.

Fearing, J. L., Jr., and Benedict, P. C., 1925, Geology of the Verde Central mine: Eng. and Min. Jour., v. 119, p. 609–611.

Finlay, J. R., 1918, The Jerome district of Arizona: Eng. and Min. Jour., v. 106, p. 557–562; 605–610.

Finlayson, A. M., 1910, The pyrite deposits of Huelva, Spain: Econ. Geology, v. 5, no. 5, p. 357–372; 403–437.

Fuller, R. E., 1931, The aqueous chilling of basaltic lava on the Columbia River plateau: Am. Jour. Sci., v. 21, p. 281–300.

Gavelin, S., 1939, Geology and ore of the Malanas district, Vasterbotten, Sweden: Sveriges geol. undersökning, Årsbok v. 33, no. 4, 198 p.

Gidley, J. W., 1922, Preliminary report on fossil vertebrates of the San Pedro valley, Ariz., with descriptions of new species of Rodentia and Langomorphia: U. S. Geol. Survey Prof. Paper 131–E, p. 119–131.

——— 1926, Fossil Proboscidea and Edentata of the San Pedro valley, Ariz.: U. S. Geol. Survey Prof. Paper 140–B, p. 83–95.

Gilbert, G. K., 1875, Report on the geology of portions of New Mexico and Arizona: U. S. Geog. and Geol. Survey, W. 100th Mer. Rept., v. 3, p. 503–567.

Gill, J. E., 1948, The Canadian Precambrian shield, *in* Structural geology of Canadian ore deposits: Canadian Inst. Min. Metallurgy, Geology Div., p. 20–48.

Graton, L. C., 1909, The occurrence of copper in Shasta County, Calif.: U. S. Geol. Survey Bull. 430–B, p. 71–111.

Grubenmann, U., and Niggli, P., 1924, Die Gesteinsmetamorphose: Bornstraege, Berlin, 539 p.

Gunning, H. C., 1937, Cadillac area, Quebec: Canada Geol. Survey Mem. 206, 80 p.

Gunning, H. C., and Ambrose, J. W., 1940, Malartic area, Quebec: Canada Geol. Survey Mem. 222, 142 p.

Gutschick, R. C., 1943, The Redwall limestone (Mississippian) of Yavapai County, Ariz.: Plateau, v. 16, no. 1, p. 1–11.

Hamilton, Patrick, 1883, Resources of Arizona: 2d ed., San Francisco, A. L. Bancroft and Co., 275 p.

——— 1884, Resources of Arizona, 3d ed.: San Francisco, A. L. Bancroft and Co., 414 p.

Hansen, M. G., 1930, Geology and ore deposits of the United Verde Mine: Min. Cong. Jour., v. 16, p. 306–312.

Hanson, G., 1920, Some Canadian occurrences of pyrite deposits in metamorphic rocks: Econ. Geology, v. 15, p. 574–609.

Heineman, S. R. E., 1938, Summary of mining history in Arizona: Ariz. Bur. Mines Bull. 145, p. 26–31.

Hendericks, H. R., 1947, Milling lead-zinc ores at the Iron King mine, Prescott, Ariz.: Am. Inst. Min. Metall. Eng., Min. Tech., v. 11, no. 4, Tech. Paper 2191, 5 p.

Heyl, G. R., 1948, Foothill copper-zinc belt of the Sierra Nevada, Calif.: Calif. Div. Mines, Bull. 144, p. 11–29.

Hinds, N. E. A., 1936, Uncompahgran and Beltian deposits in western North America: Carnegie Inst. Washington Pub. 463, p. 53–136.

Huddle, J. W., and Dobrovolny, E., 1945, Late Paleozoic stratigraphy and oil and gas possibilities of central and northeastern Arizona: U. S. Geol. Survey Oil and Gas Investigations, Prelim. Chart 10.

——— 1952, Devonian and Mississippian rocks of Central Arizona: U. S. Geol. Survey Prof. Paper 233–D, p. 67–112.

Huff, L. C., 1955, A Paleozoic geochemical anolomy near Jerome, Arizona: U. S. Geol. Survey Bull. 1000–C, p. 105–188.

Hughes, P. W., 1949, History of the Supai formation in the Black Mesa, Yavapai County, Ariz.: Plateau, v. 22, no. 2, p. 32–36.

——— 1952, Stratigraphy of Supai formation, Chino Valley area, Yavapai County, Ariz.: Am. Assoc. Petroleum Geologists Bull., v. 36, p. 635–657.

Ingalls, W. R., 1931, World survey of the zinc industry: Min. Met. Soc. America, N. Y., 128 p.

Jaggar, T. A., and Palache, Charles, 1905, Description of Bradshaw Mountains quadrangle, Ariz.: U. S. Geol. Survey Geol. Atlas, folio 126, 11 p.

James, H. L., 1954, Sedimentary facies of iron-formation: Econ. Geology, v. 49, p. 236–293.

Jenkins, O. P., 1923, Verde River lake beds near Clarkdale, Ariz.: Am. Jour. Sci., 5th ser., v. 5, no. 25, p. 65–81.

Keefe, P. C., 1930, Smelter crushing plant: Min. Cong. Jour., v. 16, p. 361–362.

Knechtel, M. M., 1936, Geologic relations of the Gila conglomerate in southeastern Arizona: Am. Jour. Sci., v. 31, p. 81–92.

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

Koffman, A. A., Price, R. L., Cairns, R. B., Ogoyzl, L. W., and Stockwell, C. H., 1948, Flin Flon mine: Structural geology of Canadian ore deposits: 1st ed., p. 295–301.

Kumke, C. A., and Mills, H. F., 1950, Mining methods and practices at the Iron King mine, Shattuck-Denn Mining Corp., Yavapai County, Ariz.: U. S. Bur. Mines Inf. Circ. 7539, 17 p.

Kuzell, C. R., 1930, Blast furnace smelting: Min. Cong. Jour., v. 16, p. 374–378.

Lanning, J. F., 1930, Historical growth of the United Verde Smelting plant at Clarkdale, Ariz.: Min. Cong. Jour., v. 16, p. 354–360.

Laskey, S. G., 1936, Geology and ore deposits of the Bayard area, Central mining district, N. Mex.: U. S. Geol. Survey Bull. 870, 144 p.

Lausen, Carl, 1928, Hydrous sulphates formed under fumarolic conditions at the United Verde mine: Am. Mineralogist, v. 13, p. 203–225.

——— 1930, The pre-Cambrian greenstone complex of the Jerome quadrangle: Jour. Geology, v. 38, p. 174–183.

Leith, C. K., 1923, Structural geology: New York, Holt and Co., 390 p.

Leith, C. K., Lund, R. J., and Leith, Andrew, 1935, Pre-Cambrian rocks of the Lake Superior region: U. S. Geol. Survey Prof. Paper 184, 34 p.

Lindgren, Waldemar, 1905, The copper deposits of the Clifton-Morenci district, Ariz.: U. S. Geol. Survey Prof. Paper 43, 375 p.

——— 1926, Ore deposits of the Jerome and Bradshaw Mountains quadrangles, Ariz.: U. S. Geol. Survey Bull. 782, 192 p.

——— 1933, Mineral deposits, 4th ed.: New York, McGraw-Hill Book Co., 930 p.

Lindgren W., and Irving, J. D., 1911, The origin of the Rammelsberg ore deposit: Econ. Geology, v. 6, p. 303–313.

Little, W. W., 1950, Radial blast holes for drilling an irregular ore body: Min. Engineering, p. 463–465.

McKee, E. D., 1937, Triassic pebbles in northern Arizona containing invertebrate fossils: Am. Jour. Sci., v. 33, p. 260–263.

——— 1940, Three types of cross-lamination in Paleozoic rocks of northern Arizona: Am. Jour. Sci., v. 238, p. 811–824.

——— 1945, Cambrian history of the Grand Canyon region: Carnegie Inst. Washington Pub. 563, 232 p.

——— 1951, Sedimentary basins of Arizona and adjoining areas: Geol. Soc. America Bull., v. 62, p. 481–506.

McNair, A. H., 1951, Paleozoic stratigraphy of part of northwestern Arizona: Am. Assoc. Petroleum Geologists Bull., v. 35, no. 3, p. 503–541.

Mahard, R. H., 1949, Late Cenozoic chronology of the Upper Verde Valley, Ariz.: Denison Univ. Sci. Lab. Bull., v. 41, p. 97–127.

Merriam, C. W., 1940, Devonian stratigraphy and paleontology of the Roberts Mountains region, Nev.: Geol. Soc. America Spec. Paper no. 25, 114 p.

Mills, H. F., 1941, Ore occurrence at the Iron King mine: Eng. and Min. Jour., v. 142, no. 10, p. 56–57.

——— 1947, Occurrence of lead-zinc ore at the Iron King mine, Prescott, Ariz.: Am. Inst. Min. Metall. Eng., Min. Tech., v. 11, no. 4, Tech. Paper 2190, 4 p.

Mooney, F. X., 1930, Reverberatory smelting: Min. Cong. Jour., v. 16, p. 379–381.

Newhouse, W. H., and Flaherty, G. E., 1930, The texture and origin of some banded or schistose sulphide ores: Econ. Geology, v. 25, p. 600–620.

Nichol, A. A., 1937, The natural vegetation of Arizona: Ariz. Univ., College Agriculture Tech. Bull. no. 68, p. 181–222.

Noble, L. F., 1922, A section of the Paleozoic formations of the Grand Canyon at the Bass Trail: U. S. Geol. Survey Prof. Paper 131–B, p. 23–73.

Noble, L. F., and Hunter, J. F., 1917, A reconnaissance of the Archean complex of the Granite Gorge, Grand Canyon, Ariz.: U. S. Geol. Survey Prof. Paper 98–I, p. 95–113.

Palache, Charles, Berman, Harry, and Frondel, Clifford, 1944. The system of mineralogy, 7th ed., v. 1, New York, John Wiley & Sons, 834 p.

Parsons, F. H., 1930, The roaster plant: Min. Cong. Jour., v. 16, p. 369.

Peterson, N. P., Gilbert, C. M., and Quick, G. L., 1951, Geology and ore deposits of the Castle Dome area, Gila County, Ariz.: U. S. Geol. Survey Bull. 971, 134 p.

Price, Peter, 1934, The geology and ore deposits of the Horne mine, Noranda, Quebec: Canadian Inst. Min. Metall. Trans., v. 36, p. 108–140.

Price, W. E., Jr., 1950, Cenozoic gravels on the rim of Sycamore Canyon, Ariz.: Geol. Soc. America Bull., v. 61, no. 5, p. 501–508.

Provot, F. A., 1916, Jerome mining district geology: Eng. and Min. Jour., v. 102, p. 1028–1031.

Pullen, J. B., 1941, Modified mining methods in the United Verde mine: Am. Inst. Min. Metall. Eng., Tech. Paper 1273, 18 p.

Ralston, O. C., 1930, Possibilities of zinc production in Arizona: Min. Jour., v. 14, p. 11.

——— 1930a, Research at the United Verde: Min. Cong. Jour., v. 16, p. 384–385.

Ransome, F. L., 1903, Geology of the Globe copper district, Ariz.: U. S. Geol. Survey Prof. Paper 12, 168 p.

——— 1904, The geology and ore deposits of the Bisbee quadrangle, Ariz.: U. S. Geol. Survey Prof. Paper 21, 168 p.

——— 1916, Some Paleozoic sections in Arizona and their correlation: U. S. Geol. Survey Prof. Paper 98-K, p. 133–166.

——— 1919, The copper deposits of Ray and Miami, Ariz.: U. S. Geol. Survey Prof. Paper 115, 192 p.

——— 1932, General geology and summary of ore deposits, *in* Ore deposits of the Southwest: 16th Internat. Geol. Cong. Guidebook 14, Excursion C–1, p. 1–23.

Reber, L. E., Jr., 1922, Geology and ore deposits of Jerome district: Am. Inst. Min. Metall. Eng. Trans., v. 66, p. 3–26.

——— 1938, Jerome district: Ariz. Bur. Mines Bull. 145, p. 41–65.

Reid, J. A., 1906, Sketch of the geology and ore deposits of the Cherry Creek district, Ariz.: Econ. Geology, v. 1, p. 417–436.

Rice, Marion, 1920, Petrographic notes on the ore deposits of Jerome, Ariz.: Am. Inst. Min. Metall. Eng. Trans., v. 61, p. 60–65.

Rickard, T. A., 1918, The story of the U. V. X. bonanza: Min. Sci. Press, v. 116, p. 9–17; p. 47–52.

Robinson, H. H., 1913, The San Franciscan volcanic field, Arizona: U. S. Geol. Survey Prof. Paper 76, 213 p.

Rubey, W. W., 1929, Origin of the siliceous Mowry shale of the Black Hills region: U. S. Geol. Survey Prof. Paper 154–D, p. 153–170.

Sandell, E. B., and Goldich, S. S., 1943, The rarer metallic constituents of some American igneous rocks: Jour. Geology, v. 51, p. 167–189.

Satterly, J., 1941, Pillow lavas from the Dryden-Wabigoon area, Kenora district, Ontario: Toronto Univ. Studies, Geol. ser. no. 46, p. 119–136.

Schouten, C., 1934, Structures and textures of synthetic replacement in "open space": Econ. Geology, v. 29, no. 7, p. 611–658.

Schwartz, G. M., 1938, Oxidized copper ores of the United Verde Extension mine: Econ. Geology, v. 33, p. 21–33.

Short, M. N., 1940, Microscopic determination of the ore minerals: U. S. Geol. Survey Bull. 914, 314 p.

Stevenson, J. S., 1937, Mineralization and metamorphism at the Eustis mine, Quebec: Econ. Geology, v. 32, p. 335–363.

Stickney, A. W., 1915, The pyritic copper deposits of Kyshtim, Russia: Econ. Geology, v. 10, no. 7, p. 593–633.

Stoyanow, A. A., 1926, Notes on recent stratigraphic work in Arizona: Am. Jour. Sci., v. 12, p. 311–324.

——— 1936, Correlation of Arizona Paleozoic formations: Geol. Soc. America Bull., v. 47, p. 459–540.

——— 1942, Paleozoic paleogeography of Arizona: Geol. Soc. America Bull. 53, p. 1255–1282.

Swanson, C. O., and Gunning, H. C., 1945, Geology of the Sullivan mine: Canadian Inst. Min. Metallurgy Trans., v. 48, p. 645–667.

Taliaferro, N. L., 1933, The relation of volcanism to diatomaceous and associated siliceous sediments: Calif. Univ. Dept. Geol. Sci. Bull., v. 23, p. 1–55.

Tally, R. E., 1917, Mine-fire methods employed by the United Verde Copper Co.: Am. Inst. Min. Metall. Eng. Trans., v. 55, p. 186–202.

Tenney, J. B., 1935, The copper deposits of Arizona *in* Copper resources of the world: 16th Internat. Geol. Cong., v. 1, p. 167–235.

Tovote, W. L., 1918, Notes on certain ore deposits of the southwest: Am. Inst. Min. Metall. Eng., Bull. 142, p. 1599–1612.

Tyrrell, G. W., 1929, The principles of petrology: New York, Dutton and Co., 349 p.

——— 1931, Volcanoes: London, Thornton Butterwork, Ltd., 252 p.

Walcott, C. D., 1890, Study of a line of displacement in the Grand Canyon of the Colorado, in northern Arizona: Geol. Soc. America Bull., v. 1, p. 49–64.

Williams, David, 1934, The geology of the Rio Tinto mines, Spain: Inst. Min. Metallurgy Bull. 355, 48 p.

Williams, J. J., 1930, Converter department of the United Verde Copper Co.: Min. Cong. Jour., v. 16, p. 382–383.

Wilson, E. D., 1939, Pre-Cambrian Mazatzal revolution in central Arizona: Geol. Soc. America Bull., v. 50, p. 1113–1164.

——— 1942, Arizona lode gold deposits, *in* Ore deposits as related to structural features, edited by W. H. Newhouse: Princeton Univ. Press, p. 242–243.

Wilson, M. E., 1941, Noranda district, Quebec: Canada Geol. Survey Mem. 229, 162 p.

Winchell, A. N., 1933, Elements of optical mineralogy: 3d ed., 459 p.

——— 1936, A third study of chlorite: Am. Mineralogist, v. 21, p. 642–651.

Young, H. V., 1930, Historical sketch of the United Verde Copper Co.: Min. Cong. Jour., v. 16, p. 303–305.

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