Geology and Ore Deposits of the Castle Dome Area Gila County, Arizona

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By N. P. PETERSON, C. M. GILBERT, *and* G. L. QUICK

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CONTENTS

ILLUSTRATIONS

TABLES

GEOLOGY AND ORE DEPOSITS OF THE CASTLE DOME AREA, GILA COUNTY, ARIZONA

By N. P. Peterson, C. M. Gilbert, and G. L. Quick

ABSTRACT

The Castle Dome area is in the Globe-Miami mining district, Gila County, Ariz., 5 miles west of Miami. The map area includes about 6 square miles in the west-central part of the Globe quadrangle.

The Pinal schist, the oldest rock in the region, is composed of metamorphosed sedimentary and igneous rocks of pre-Cambrian age. Also of pre-Cambrian age but formed much later than the Pinal schist are rocks of the Apache group. A major unconformity separates the Pinal schist from the overlying Apache group.

The Cambrian Troy quartzite was deposited after a period of erosion during which all the Mescal limestone (uppermost formation of the Apache group) probably was removed. The deposition of the Devonian Martin limestone followed another period of erosion during which all previously deposited Paleozoic sediments except a few small remnants of the Troy quartzite were removed. The Martin limestone is conformably overlain by the Mississippian Escabrosa limestone, and it in turn is overlain by the Pennsylvanian Naco limestone. The absence of upper Mississippian beds, however, suggests a disconformity between the Escabrosa and the Naco.

The period after the deposition of the Naco limestone was marked by faulting and the intrusion of a variety of igneous rocks. The first intrusion of this period was a small mass of granodiorite, and it was followed by the larger quartz monzonite complex in which the Castle Dome ore body occurs. The quartz monzonite intruded the lower formations of the Apache group and possibly the lower part of the limestone of Paleozoic age.

Normal faulting, followed by the intrusion of great volumes of diabase, occurred after the emplacement of the quartz monzonite, probably during the Mesozoic era. The diabase forced its way between the beds of sedimentary rock and occupied many of the faults. The last intrusion of this period is represented by dikes and small masses of granite porphyry south and east of the mine. The copper mineralization followed the intrusion of granite porphyry.

Most, if not all, of the Whitetail conglomerate was deposited during the early part of the Tertiary period, after which a great sheet of dacite covered the entire region. The Gila conglomerate was deposited during late Pliocene and Pleistocene time. Normal faulting continued throughout the Mesozoic era, during the Tertiary period, and until after the deposition of the Gila conglomerate in the Quaternary; all except the most recent gravel and talus deposits have been displaced.

The copper mineralization of the Castle Dome deposit is confined mainly to the quartz monzonite, which occurs in an uplifted block about 1 by 1½ miles in

size. The hypogene minerals consist of pyrite, chalcopyrite, and a little molybdenite, sphalerite, and galena, together with products of the hydrothermal alteration of the host rock. Pyrite and chalcopyrite are disseminated or occur in small, closely spaced quartz veins striking east-northeast and dipping steeply to the south.

The distribution of the hypogene minerals shows a distinct zoning pattern; therefore the mineralized area can be divided into (1) a pyritic zone and (2) a chalcopyrite-rich zone that borders the south side of the pyritic zone. The hydrothermal alteration accompanying the mineralization consists of a clay phase that is general throughout the pyritic and chalcopyrite-rich zones and a quartz-sericite phase related to pyrite veins. The quartz-sericite alteration is intense in the pyritic zone, where the veins are largest and most numerous. North and south of the most intensely mineralized area, the clay and quartz-sericite alteration gives way to a weak propylitic-type alteration.

Several thin diabase sills crop out in the southern part of the quartz monzonite body and dip gently to the north through the chalcopyrite and pyritic zones. The copper content of the rock in and near these sills is substantially higher than average. Supergene enrichment, though not extensive, play an important part in the formation of the ore body, the chalcopyrite zone being the protore.

The localization of the ore was due to the combined effects of three geologic controls: (1) zoning in the hypogene mineralization, (2) the rich hypogene mineralization associated with the diabase sills, and (3) supergene enrichment.

The Castle Dome open-pit mine was developed in 1942 and 1943 by the Castle Dome Copper Co., Inc., as a war project. Slightly more than 177,000,000 pounds of copper from about 14,000,000 tons of ore was produced from June 1943 to the end of 1946.

INTRODUCTION

The Castle Dome area is in the Globe-Miami mining district, Gila County, Ariz., 5 miles west of Miami, the nearest railroad station. It is in T. 1 N., R. 14 E., and is a part of the Globe quadrangle as mapped by the Geological Survey in 1901 (figs. 1, 2). The Castle Dome open-pit copper mine is on the south flank of Porphyry Mountain, a prominent landmark of the region, and is accessible only by a branch road 4 miles long that connects, 5 miles west of Miami, with U. S. Highway 70.

Trucks are used to bring in supplies and to transport the copper concentrate to the International Smelting & Refining Co.'s smelter at Miami. Water for mine and mill use is pumped from the Old Dominion mine at Globe. Domestic water is supplied by a well in Pinto Creek about 3 miles south of the camp.

The mine is owned and operated by the Castle Dome Copper Co., Inc., a subsidiary of the Miami Copper Co. It was developed by the owners as a war project, and a concentrator was provided by the Defense Plant Corporation. The production of copper concentrate began in June 1943. The plant has a capacity of 12,000 tons of ore per day or about 4,000,000 pounds of copper per month.

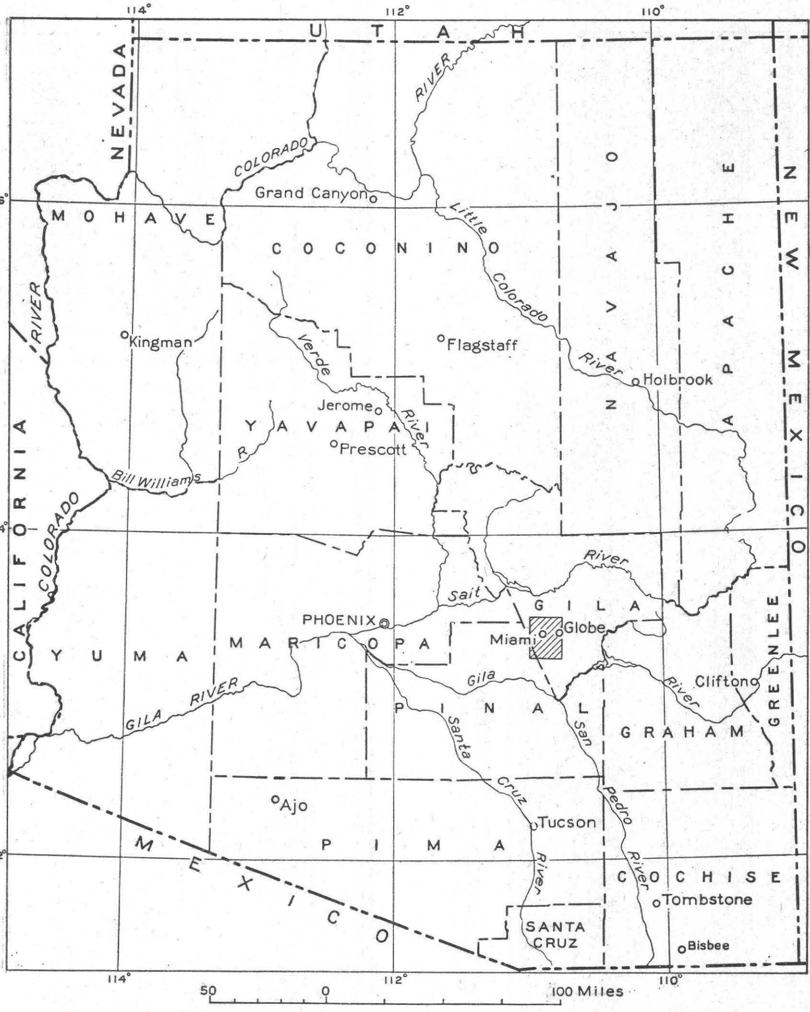

FIGURE 1.—Index map of Arizona showing location of Globe quadrangle.

The general geology of the Globe quadrangle, which includes the Castle Dome area, was studied by F. L. Ransome in 1901 and 1902. His findings were published by the Geological Survey as Professional Paper 12 (1903) and as folio 111 of the Geologic Atlas (1904). In 1911, after the importance of the low-grade disseminated copper deposit had become apparent, Ransome returned to the district for further studies, the results of which were included in Professional Paper 115 (1919). The present study is in much greater detail than Ransome's and is confined to the area surrounding the Castle Dome mine. The field work for the report was done during the period from July 1943 to September 1944. An area of about 6 square miles was mapped at

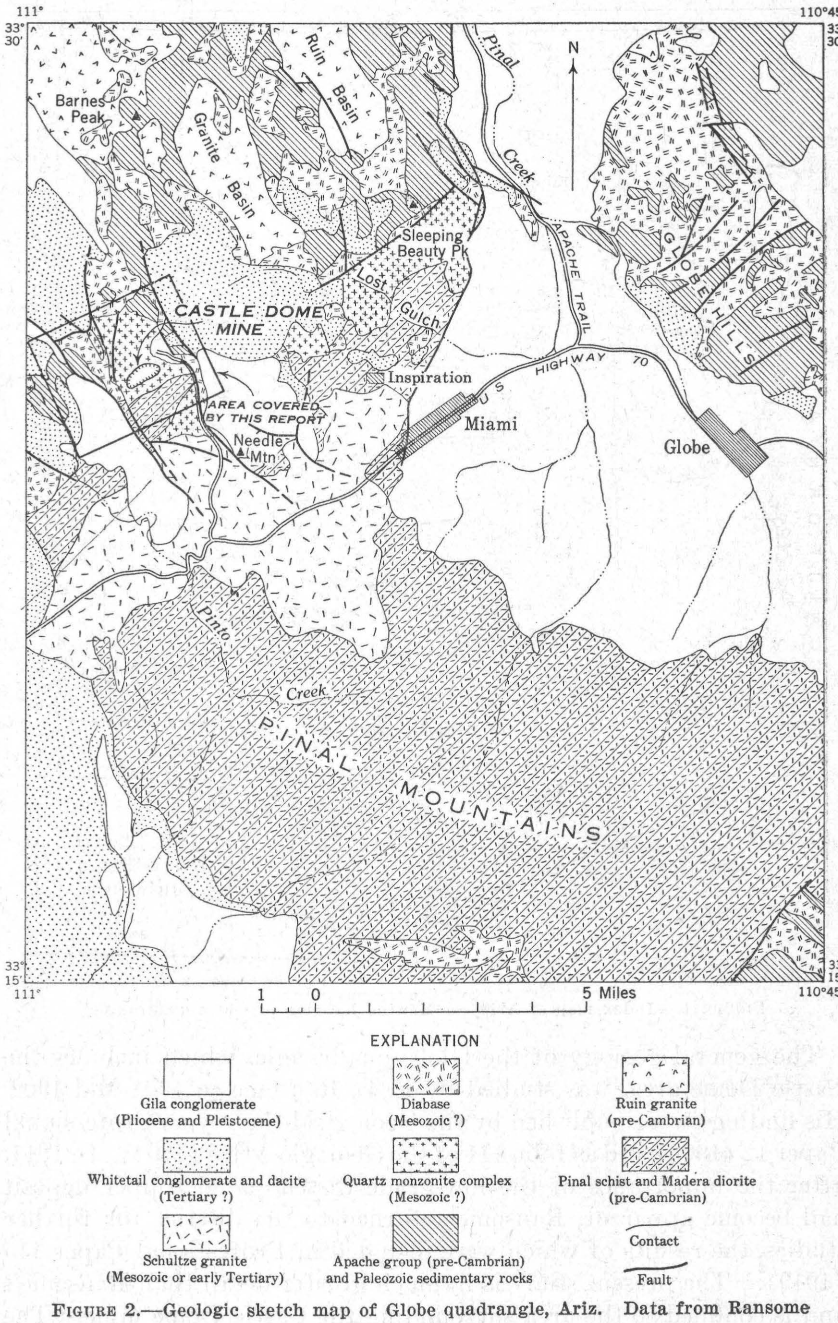

FIGURE 2.—Geologic sketch map of Globe quadrangle, Ariz. Data from Ransome with modifications.

a scale of 1:2,400 on a topographic base provided by the Castle Dome Copper Co. This map was reduced to a scale of 1:6,000 and transferred to a topographic base prepared by Survey staff by means of photogrammetric methods (pl. 1).

The writers wish to acknowledge the cordial cooperation of the staff of the Castle Dome Copper Co., especially R. W. Hughes, general manager, B. R. Coil, general superintendent, and J. C. Van De Water, mine superintendent. The company made available all records and engineering data pertaining to the exploration and development of the property, including the exploratory drill logs and sludges, and the 1:2,400 multiplex topographic map used as a base for areal geologic mapping.

GEOLOGIC HISTORY OF THE CASTLE DOME AREA

The generalized columnar section of sedimentary and volcanic rocks occurring within the map area is illustrated in figure 3. Within this area, the record of pre-Cambrian history is meager and obscure. What remains of that record is contained in the small exposure of Pinal schist south of the Castle Dome mine. A series of sandstones and shales, containing a few sills or flows of basic volcanic rocks, was deposited, and stresses in a general northwesterly direction produced metamorphism in the sedimentary and igneous rocks alike and developed the series of crystalline schists seen today. Thick sills of granite were intruded into the schist late in the period of metamorphism. As a result of this orogenic activity, the area was probably high and mountainous, and a new cycle of weathering and erosion was begun. Ultimately erosion reduced the ancient mountains to a nearly plane surface littered with the disintegrated debris of the ancient rocks. Over this surface the Apache sea advanced, and the weathered debris on the surface was reworked to form the thin layer of Scanlan conglomerate. Over this were deposited the sandstones and sandy shales that comprise the Pioneer formation. An abrupt change in the character of sedimentation is recorded by the Barnes conglomerate; this was then buried by the Dripping Spring quartzite, which in turn was covered by the cherty, dolomitic Mescal limestone.

After the deposition of the Mescal limestone the region was elevated and the Apache sea withdrew. The next recorded event in the history of the Castle Dome area was erosion. Although faulting may have accompanied the uplift, there was no folding, and the strata of the Apache group remained essentially horizontal. Erosion cut deeply into the exposed rocks and removed all the Mescal limestone and most of the Dripping Spring quartzite, leaving a surface that probably had low relief and was traversed by shallow channels strewn with

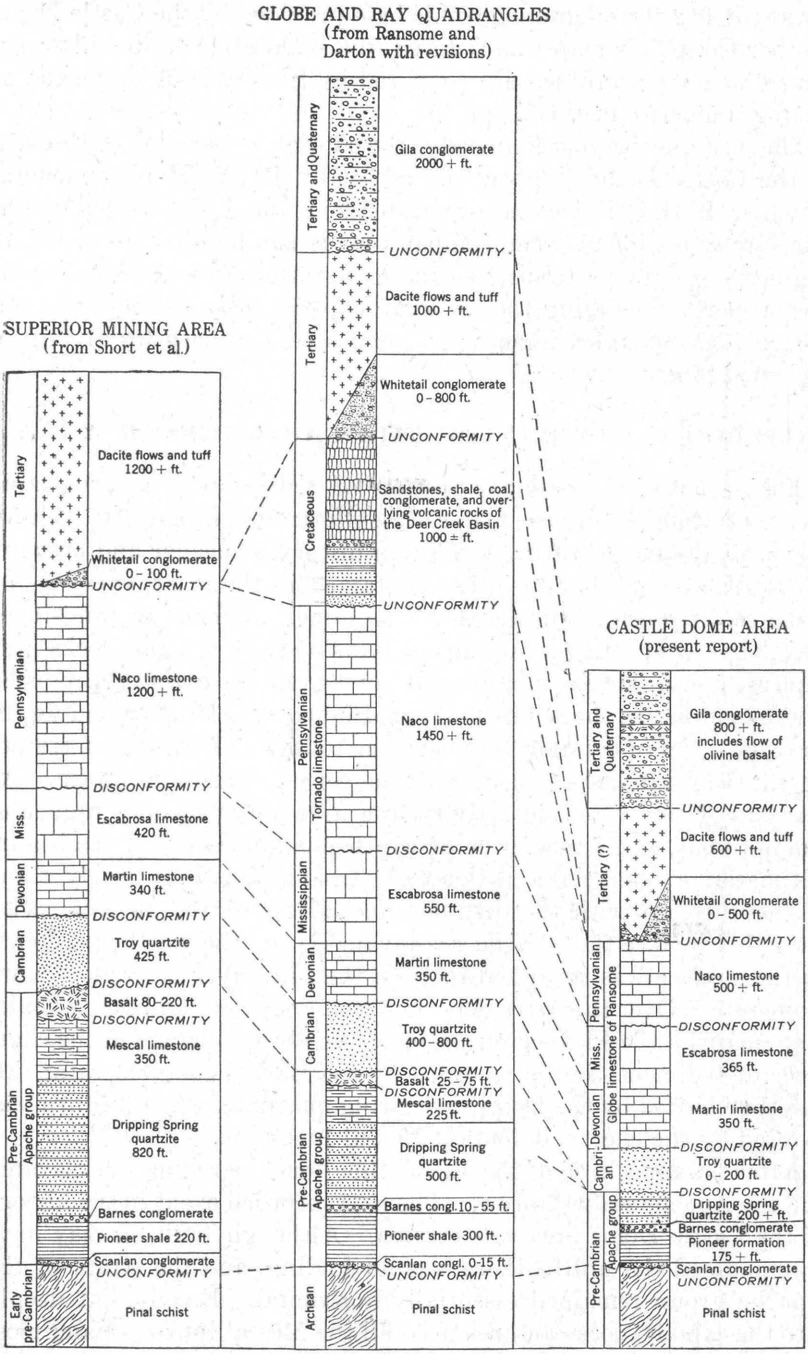

FIGURE 3.—Generalized columnar sections of sedimentary and volcanic rocks, Castle Dome area, Ariz.

blocks of Dripping Spring quartzite, Barnes conglomerate, and the shale and sandstone of the Pioneer formation.

During the Cambrian period the Troy quartzite was deposited on this pre-Cambrian erosion surface, probably in a shallow sea. The basal part of the Troy quartzite, particularly where it occurs in the channels on the old surface, is a coarse red conglomerate containing angular blocks of the underlying Apache rocks. The conglomerate grades upward into thinner-bedded, red, argillaceous sandstone. It is not known how long this deposition may have continued or how thick the deposit became in this area, for all that remains today is a few remnants of the lower beds of the Troy quartzite in the western part of the area. Again the region was elevated without noticeable tilting or folding, and erosion removed the Troy quartzite in most places and cut into the underlying rocks of the Apache group. The erosion surface had very little relief, and in Upper Devonian time the sea spread over it again.

In this sea the impure magnesian limestone, sandstones, shales, and marls of the Martin limestone were deposited. Deposition continued without a break at least through the early part of the Mississippian epoch, when the massive, pure Escabrosa limestone was laid down. A break in sedimentation probably occurred during the later part of the Mississippian epoch, and there may have been some erosion of the Escabrosa limestone before the deposition of the Naco limestone in the early part of the Pennsylvanian, but in the Castle Dome area the separation between the Escabrosa and Naco limestones is marked only by a red shale that locally contains lenses of chert breccia. If erosion occurred before the deposition of the Naco limestone, it was not sufficient to produce noticeable differences in the thickness of the Escabrosa limestone in this area.

There is no record of sedimentation between the Pennsylvanian epoch and the time when the Tertiary (?) Whitetail conglomerate was deposited. The history of that interval, comprising most or all of the Mesozoic era, is rather one of faulting and widespread intrusion by a variety of igneous rocks. The oldest of these intrusions is a small mass of biotite granodiorite south of Porphyry Mountain, a narrow body lying along the north edge of the Pinal schist. The granodiorite may be an early phase of the larger biotite-quartz monzonite intrusion that occupies the central part of the area.

The quartz monzonite intruded the lower part of the Apache group; although its relation to the Paleozoic limestones is unknown, it is tentatively regarded as post-Pennsylvanian and probably Mesozoic in age. North of Porphyry Mountain the intrusion has a concordant roof and may be a sill below the Scanlan conglomerate, but south of Porphyry Mountain it is probably a discordant body.

Normal faulting followed the emplacement of the quartz monzonite, producing a horst that in general trends north-northwestward through the central part of the area; within this horst the exposed quartz monzonite is now confined. These dislocations were followed by the intrusion of great volumes of diabase, probably during the Mesozoic era and particularly along the east, west, and north sides of the quartz monzonite block and into the surrounding sedimentary rocks. Earlier faults afforded the main channels along which the diabase invaded the older rocks, but the magma also forced its way as numerous sills between the beds of sedimentary rock. The largest masses were injected beneath the Martin limestone and produced faults in the overlying rocks. Little of the magma found its way into the quartz monzonite, but some diabase occurs there as dikes and small masses injected along faults and large joints. Small intrusions of diorite porphyry occur in the diabase and in the rocks of the Apache group, but they constitute only a minor phase of igneous activity that occurred prior to the deposition of the Whitetail conglomerate. Actually these diorite porphyry injections may be related to the great diabase intrusion.

The last intrusion of pre-Whitetail age consists of dikes and small masses of biotite granite porphyry injected into the quartz monzonite south of Porphyry Mountain and into the adjacent diabase. Most of the granite porphyry bodies are elongated and alined parallel to the granodiorite intrusion and the southern margin of the quartz monzonite as though they were controlled by an earlier structure. Though there is no definite evidence as yet of a relationship between them, a similarity exists between these granite porphyry injections and the porphyry along the border of the late Cretaceous or early Tertiary Schultze granite farther south. The two may be contemporaneous intrusions.

The hypogene mineralization in the Castle Dome ore deposit occurred between the intrusion of the granite porphyry and the deposition of the Whitetail conglomerate, most probably at the same time the hypogene mineralization of the Miami, Inspiration, and Old Dominion ore deposits took place. The copper-bearing solutions probably emanated from the same magma chamber that supplied the granite porphyry and the Schultze granite.

Whatever the details of the geologic history during the Mesozoic era may be, a great deal of erosion undoubtedly occurred at that time. Faulting and the intrusion of such great volumes of magma relatively close to the surface must have caused elevation and erosion of the area. By the time streams began to deposit the Whitetail conglomerate, probably during the early part of the Tertiary period, much of the Paleozoic limestones had been eroded away, and the rocks of the Apache group and the diabase were exposed in many places. Whether or not the quartz monzonite and granite porphyry intrusions had been largely uncovered is uncertain, but a few fragments of both rocks occur in the Whitetail conglomerate, and it is possible that such extensive erosion as preceded the deposition of the Whitetail had actually removed the roof from the quartz monzonite mass. The great variation in the thickness of the conglomerate shows that the surface on which it was deposited was of moderate relief, but the Castle Dome area as a whole must have been relatively low lying compared with the surrounding region, for within it the conglomerate is thickest and most widespread. Probably most if not all of this area was buried by the bouldery detritus that comprises the Whitetail conglomerate.

Ransome has shown that most of the supergene enrichment in the Miami and Inspiration ore bodies took place during the period of erosion preceding the deposition of the Whitetail conglomerate. In the Castle Dome ore deposit no evidence of enrichment prior to the deposition of the Whitetail has been recognized. The mineralized area was then probably below the water level and therefore was not affected by supergene agencies.

There followed a widespread outpouring of silicic magma, forming a great sheet of biotite dacite; at least 1,000 feet thick, it covered the entire region. The first of the volcanic outbursts produced ash, some of which is interbedded in the top of the Whitetail conglomerate, but thereafter fluvial deposition ceased and eruptions must have been almost continuous.

After the dacite eruption the area was again disturbed by faulting. The old horst block was elevated along normal faults on its east side and perhaps along faults to the west. A new cycle of erosion was initiated, and the pre-Tertiary rocks of the highest areas were laid bare, although in other places the dacite and Whitetail conglomerate were not entirely removed. In the valleys of this erosion surface, streams deposited the sands and coarse gravels of the Gila conglomerate, mostly in the form of alluvial fans. Some volcanic activity at this time is indicated by a flow of olivine basalt in the conglomerate southwest of the mine and by the basalt dike south of Jewel Hill. Whether or not the quartz monzonite on Porphyry Mountain was stripped of its cover by the pre-Gila erosion is doubtful, but if it was uncovered it was buried again by the Gila conglomerate. Post-Gila faults of considerable displacement formed west of Porphyry Mountain, elevating the old horst again, and subsequent erosion developed the present rugged topography. As the stream valleys approached their present form, deposits of talus covered their sides and merged with alluviated channels in their bottoms, but because of very recent rejuvenation the older talus and alluvium have been dissected, and the present streams flow in entrenched channels as much as 50 feet deep.

The supergene enrichment of the Castle Dome ore deposit is clearly related to the present topography; it took place while the present drainage system was being developed and while the older alluvium was accumulating in the channel of Gold Gulch. Probably it was well advanced before the rejuvenation of the drainage occurred.

ROCK DESCRIPTIONS

PRE-CAMBRIAN ROCKS

PINAL SCHIST AND ASSOCIATED ROCKS

The schist exposed in the southern part of the Castle Dome area is an outlier of the extensive area of Pinal schist that makes up the Pinal Mountains (fig. 2). It is a series of dynamically metamorphosed sedimentary and igneous rocks of early pre-Cambrian age. Although the attitude of the schistosity differs from place to place, the general strike is about N. 50° E., and the dip is approximately vertical. The schistosity is generally parallel to the bedding of the original sedimentary rocks.

Several types of schist, derived from a variety of original rocks, have been recognized but are not separately mapped. A common type is coarse-grained quartz-muscovite schist with or without feldspar and biotite and with relic sedimentary structures recognizable in some places. Probably most of it is metamorphosed feldspathic sandstone, although some may have been derived from fine-grained silicic igneous rock. A fine-grained quartz-sericite-chlorite schist that is gray in color and has a satiny luster was probably derived from shale. Several lenticular masses of fine-grained amphibole schist, surrounded by mica schist, indicate that basic igneous rocks were intruded into or interbedded with the sedimentary rocks before metamorphism. Generally all these types of schist are contorted and contain abundant quartz veins and a few small tourmaline-muscovite pegmatite veins. Quartz from these veins undoubtedly composes the quartz pebbles in the Scanlan conglomerate to the north. A fourth type of schist, probably derived from sandstones, occurs south of the map area. Unlike the schist to the north, it contains few veins and is not contorted. Its gneissic structure is due to the development of most of the mica in thin layers, separated by other layers composed mostly of equigranular quartz and feldspar.

Intruded into the Pinal schist is a distinctive light-gray, medium-grained granite containing orthoclase, plagioclase, quartz, muscovite, and a little biotite. As shown on the map (pl. 1), the several masses of this granite are elongated parallel to the general schistosity and are probably thick sills, although some of the smaller bodies are very irregular and cut across the schistosity in many places. Crude foliation has been developed in the granite by the crushing and elongation of the quartz and feldspar and by the orientation of mica. This foliation is clearly of metamorphic origin, as it is everywhere parallel to the regional schistosity and not related to discordant contacts of the granite bodies. Nevertheless, most of the metamorphism of the Pinal schist preceded the intrusion of the granite, for on Manitou Hill, about 2½ miles south of Porphyry Mountain, unoriented inclusions of Pinal schist occur in the granite.

Three northeastward-trending belts of schist can be distinguished in the area south of the Castle Dome mine. The northern belt, which crops out south of the concentrator, contains quartz-muscovite and quartz-sericite-chlorite schist and all the amphibole schist. The central belt contains the granite intruded into quartz-muscovite and quartz-sericite-chlorite schist. This belt of granite intrusions can be followed southwestward for at least 2 miles beyond the limits of the map. The southern belt, largely south of the map area, contains the uncontorted quartz-feldspar-mica schist with a gneissic structure.

APACHE GROUP

DISTRIBUTION, GENERAL CHARACTER, AND AGE

Rocks of the Apache group crop out through the northern half of the Globe quadrangle and are widespread in the Ray quadrangle to the south (Ransome, 1903, pl. 1; 1919, pl. 2). These two areas are separated by an older complex of schist and granitic rocks that composes the Pinal Range in the southern part of the Globe quadrangle.

The Apache group was named and first described by Ransome (1903, pp. 29–32) in his report on the Globe quadrangle. He chose Barnes Peak, in the northwestern part of the quadrangle 4½ miles north of the Castle Dome mine, as the most complete and representative section of the group; there he subdivided it into four distinct but conformable formations, which, in ascending order, are the Scanlan conglomerate, Pioneer shale, Barnes conglomerate, and Dripping Spring quartzite. The Scanlan conglomerate, which "at Barnes Peak" lies on the Ruin granite, was named from Scanlan Pass east of Barnes Peak. It ranges from 1 to 6 feet in thickness and is composed of abundant pebbles of vein quartz and a few pebbles of schist in a matrix of pink or reddish arkose. The Pioneer formation is about 200 feet thick at Barnes Peak and is composed largely of arenaceous shale containing a few beds of quartzite, except for about 25 feet of pink arkose at the base of the formation. It was named from the old mining camp of Pioneer just south of the Globe quadrangle. The Barnes conglomerate is a single bed, between 10 and 15 feet thick, composed of well-rounded pebbles and cobbles of quartzite and a few pebbles of red chert and vein quartz in a matrix of feldspathic quartzite. The Dripping Spring quartzite, named from its extensive exposures in the Dripping Spring Range south of the Globe quadrangle, is about 400 feet thick. The lower half is thick-bedded buff quartzite, whereas the upper half is thinner-bedded, laminated, rusty quartzite. The topmost formation on Barnes Peak is a gray limestone that conformably overlies the Dripping Spring quartzite and was included by Ransome in the Globe limestone.

Later, when he mapped the Ray quadrangle, Ransome (1919, pp. 39–45) discovered that the section of the Apache group at Barnes Peak is not complete. In the Ray quadrangle, the Dripping Spring quartzite is conformably overlain by thin-bedded, cherty limestone that Ransome named from the Mescal Range, where it is well exposed. At the top of this Mescal limestone is a flow of vesicular basalt. The combined thickness of the limestone and basalt is between 250 and 300 feet. Overlying the basalt is the Troy quartzite, a massive pebbly rock about 400 feet thick, and this in turn is overlain by the Devonian Martin limestone that was included in the Globe limestone in Ransome's report on the Globe quadrangle.

As a result of his work in the Ray quadrangle, Ransome (1919, pp. 49–50) regarded the Apache group as a conformable sequence of Cambrian and possibly Ordovician and Silurian strata, including the Scanlan conglomerate, Pioneer shale, Barnes conglomerate, Dripping Spring quartzite, Mescal limestone, and Troy quartzite. He also thought that the Apache group was conformably overlain by the Martin limestone of Devonian age. More recently, Darton (1925, pp. 34–37) described a major disconformity between the Cambrian Troy quartzite and the underlying Mescal limestone and basalt in the Mescal Range south of Globe. A disconformity between Troy quartzite and Mescal limestone near Superior, 11 miles southwest of Castle Dome, has also been described (Short and others, 1943, pp. 22–24). In the Mescal Range, Darton (1925, pp. 52–53) found a second disconformity at the base of the Martin limestone. As a result of Darton's work, the Apache group is now regarded as pre-Cambrian and includes the five formations—Scanlan conglomerate, Pioneer formation, Barnes conglomerate, Dripping Spring quartzite, and Mescal limestone—that disconformably underlie the Cambrian Troy quartzite in central Arizona.

Present mapping in the northwestern part of the Globe quadrangle has clearly revealed the two major disconformities described by Darton, one between Troy quartzite and the underlying Apache group and the other at the base of the Martin limestone. Both Mescal limestone and Troy quartzite crop out in this area, although neither was recognized by Ransome as a distinct formation. On the road between Ruin Basin and Granite Basin, 5 miles northeast of Castle Dome, Dripping Spring quartzite is overlain by thin-bedded, cherty Mescal limestone at least 300 feet thick, and as was suggested by Darton (1925, p. 246), the limestone overlying the Dripping Spring quartzite at Barnes Peak is probably the Mescal. These and other outcrops of Mescal limestone in the Globe quadrangle were originally included by Ransome in the Globe limestone. The Troy quartzite crops out along Pinto Creek near the western edge of the quadrangle, where it lies on the eroded surface of both Dripping Spring quartzite and Pioneer formation. This exposure of the Troy was well described by Ransome (1903, pp. 33–34), but he did not recognize it as a distinct formation and instead considered it a variation of the Dripping Spring quartzite. Generally, throughout the Castle Dome area, both Mescal limestone and Troy quartzite have been removed by pre-Devonian erosion, and the Martin limestone lies on the rocks of the Apache group.

DESCRIPTION

The Mescal limestone is absent in the Castle Dome area, but the lower four formations of the Apache group crop out along the east, north, and west sides of the quartz monzonite. These are among the southernmost exposures of Apache rocks north of the Pinal Range.

The Scanlan may be considered the basal conglomerate of the Pioneer formation. It ranges from a thin layer of scattered pebbles to a bed about 6 feet thick. Subangular pebbles of glassy vein quartz as much as 4 or 5 inches long in a reddish arkose matrix are its most characteristic features and distinguish it from other formations. It contains, also, angular fragments of aplite and small flat pebbles of Pinal schist, but the schist pebbles are abundant only where the conglomerate is thickest. At the top is a layer of hard, fine-grained black rock about 1 foot thick. Where the conglomerate is very thin, this black bed makes up the entire formation. Generally it contains angular fragments of vein quartz and pink orthoclase scattered through the dense black matrix. Microscopic examination shows that it is a silty shale containing angular pebbles of orthoclase and quartz.

The black color, which is its distinctive feature, is caused by tiny, scattered grains of specular hematite. These are so abundant in some places that a thin section is almost opaque.

A concordant intrusion of quartz monzonite underlies the Scanlan conglomerate in the Castle Dome area, but in most places the two are separated by a thin metamorphosed arkose (mapped as "granite") composed essentially of red orthoclase and quartz. This arkose is coarse-grained and has the appearance of a sheet of red granite between the Scanlan conglomerate and the quartz monzonite. Indeed, Ransome (1903, p. 33) regarded it as the surface of the underlying granitic rocks reddened by weathering on an old pre-Apache erosion surface. Actually it probably is deeply weathered pre-Apache granite, but it is intruded by, not derived from, the quartz monzonite now exposed beneath it. Blocks of it are included in the quartz monzonite, and thin sections show that it has been partly metamorphosed.

The best exposure of this intrusive relationship between the arkose and quartz monzonite is seen half a mile north of the Continental mine, where two large and several small blocks of arkose are included in the quartz monzonite (pl. 1). The two large inclusions are thin, tabular masses oriented parallel to the bedding in the Scanlan conglomerate, from which they are separated by a quartz monzonite sill about 10 feet thick.

Because, in many places, the arkose seems to grade into the overlying Scanlan conglomerate, it is regarded as a sedimentary rock and called an arkose, but neither in thin section nor in the field does it show any typical sedimentary structures. It may include remnants of an older granitic rock, but most of it is probably a residual deposit of disintegrated granite. In some places the arkose is absent, but generally it is a layer no more than 20 feet thick between the quartz monzonite and the Scanlan conglomerate. One block included in the quartz monzonite northwest of the Castle Dome mine is at least 50 feet thick and may be an inclusion of an old granite.

Although little has been learned as yet about the original extent of the arkose, its presence indicates that the Scanlan conglomerate within the map area was deposited on an erosion surface cut across disintegrating granite that is not now exposed. Such a relationship is visible today in Richmond Basin, about 10 miles north of Globe, where a thin layer of arkose lies on an older granite and is overlain by Scanlan conglomerate. The arkose and the conglomerate are gradational, and except for the numerous pebbles in the conglomerate the two could not be distinguished from each other. Here and there the conglomerate lies directly on the granite, and in several of these places aplite dikes in the granite are truncated and fragments of aplite included among the pebbles in the overlying conglomerate. The arkose is simply the weathered debris of the underlying granite accumulated in the lower parts of an old erosion surface. So little has the debris been transported that, in many places, one can scarcely decide where the granite leaves off and the arkose begins. A similar relationship probably existed between the Scanlan conglomerate and the older granite in the Castle Dome area before the younger quartz monzonite was intruded.

The Pioneer formation is composed mainly of hard, fine-grained, reddish-brown arkosic sandstone. No complete section of it is exposed in the Castle Dome area, but it is at least 165 feet thick. Both east and west of Sleeping Beauty Peak, about 6 miles east-northeast of Castle Dome, the Pioneer formation is at least 300 feet thick. The prominent sandstone beds in the formation are 3 to 18 inches thick, but many of them are laminated or cross-laminated and ripple-marked and are separated by films of shale. The basal 15 or 20 feet of the Pioneer formation is pebbly arkose that is similar to the matrix of the Scanlan conglomerate. It grades upward into reddish, medium-grained arkosic sandstone. The upper part of the formation is fine-grained sandstone and arenaceous shale characterized by abundant greenish spots. It is less feldspathic than the lower beds and generally is light brown or gray in contrast to the underlying dark reddish-brown sandstone. However, the gradation from coarse to finer grain toward the top is not without exception, although it is a general feature of the formation. In the northeastern part of the map area, coarse pebbly arkose like that at the base of the formation occurs about 75 feet above the Scanlan conglomerate, and in an outcrop 2,500 feet east of the Horrell Ranch road in the northwestern part of the map area, the arenaceous shale in the upper part of the formation is overlain by distinctly coarser arkosic sandstone.

The Barnes conglomerate averages about 15 feet in thickness and is composed of well-rounded pebbles and cobbles of quartzite in a hard matrix of arkosic sandstone. In addition, it contains a few pebbles of vein quartz and red chalcedonic chert. In some places it is a single bed of closely packed pebbles, but elsewhere the pebbles are fewer and commonly occur in thin lenses in the matrix. It is generally gray or buff and forms prominent outcrops.

Dripping Spring quartzite crops out on the south side of Jewel Hill, in a fault zone about 1,000 feet north of the Continental mine, and in the northwestern part of the map area. In each of these areas only thin remnants of it are exposed. It is a medium-grained, gray or brown quartzite occurring in laminated beds as much as a foot thick. Though it contains feldspar, no part of it is as arkosic as the Pioneer formation, but in many small outcrops where neither the Barnes nor the Scanlan conglomerate is present the distinction between the Pioneer formation and the Dripping Spring quartzite is not clear. Near Sleeping Beauty Peak, the Dripping Spring quartzite has a maximum thickness of 500 feet. Near the base, in that area, the formation contains buff, medium-bedded feldspathic sandstone and a few massive layers of white quartzite, whereas the upper part is thin-bedded red sandstone containing some thin layers of black shale.

PALEOZOIC SEDIMENTARY ROCKS

Like the Apache group, the Paleozoic rocks occur throughout the Ray quadrangle and in the northern half of the Globe quadrangle, the two areas being separated by the older pre-Cambrian complex in the Pinal Range. Paleozoic rocks in the Castle Dome area are represented by the Troy quartzite (Cambrian), the Martin limestone (Devonian), the Escabrosa limestone (Mississippian), and the Naco limestone (Pennsylvanian).

TROY QUARTZITE

The Troy quartzite was named by Ransome (1919, pp. 44–45) for its extensive exposures on Troy Mountain in the Dripping Spring Range. The lower and middle parts of the formation are composed mainly of thick, cross-bedded layers of pebbly quartzite and conglomerate, whereas the upper part is "invariably composed of thin, generally yellowish or rusty, worm-marked shaly quartzite." In contrast to the Dripping Spring quartzite, the Troy quartzite contains little feldspar. Much of it is light gray or white on fresh fractures, but weathered surfaces are brown or maroon. In the Ray quadrangle it averages about 400 feet in thickness, but along the Gila River, a short distance to the east, Darton (1925, p. 36) estimated that it was about 800 feet thick. Only a few outcrops of Troy quartzite are known in the Globe quadrangle. Thin remnants occur in the western part of the Castle Dome area and along Pinto Creek, and Darton (1925, pp. 34, 240, 246) mentioned its presence in the Globe Hills and along Pinal Creek several miles north of the quadrangle boundary.

Darton (1925, pp. 32–37; pl. 72) was the first to describe the major disconformity that separates the Troy quartzite from the underlying Apache group. According to him, the Troy quartzite in the Mescal Mountains south of Globe contains a basal conglomerate that includes debris from the underlying Mescal limestone and Dripping Spring quartzite; near the southern end of the range it completely overlaps the Apache group and lies directly on older granitic rocks. In the

Castle Dome area, this disconformity is easily recognized because the Mescal limestone is absent, and the few remnants of the Troy quartzite lie on an older erosion surface from which even the Dripping Spring quartzite was removed in some places. Furthermore, a conglomerate at the base of the Troy is composed mostly of angular blocks of quartzite and Barnes conglomerate of the Apache group.

Along the Gila River, south of Globe, Darton (1925, pp. 32–36) found Upper Cambrian *Lingulae* in sandstones overlying but gradational into the Troy quartzite. These fossiliferous beds he correlated with the Upper Cambrian Abrigo limestone in the Bisbee region of southern Arizona. On the basis of later work, however, Stoyanow (1936, pp. 474–480) regards the Troy quartzite as Middle Cambrian.

The Troy quartzite crops out in only one place within the map area, on the west side of Gold Gulch due west of Porphyry Mountain, where it underlies the Martin limestone and apparently was deposited on the Pioneer formation, although a diabase sill has since been intruded between them. The relationships are somewhat uncertain because of the presence of the sill, but this remnant of Troy quartzite probably represents a part of the formation that was deposited in a shallow channel on the old pre-Troy erosion surface and later was overlapped by the Martin limestone. Just west of the map area, Troy quartzite is exposed on both sides of Pinto Creek near the edge of the Globe quadrangle. There it is between 160 and 200 feet thick and lies on the eroded surface of Dripping Spring quartzite and Pioneer formation.

Along both Gold Gulch and Pinto Creek, the basal part of the Troy quartzite is a massive, coarse red conglomerate, as much as 40 feet thick, composed of a red sandstone matrix that contains large, angular blocks of quartzite and Barnes conglomerate of the Apache group, together with numerous smaller, rounded pebbles of quartzite derived from the Barnes conglomerate. The coarse basal conglomerate grades upward into pebbly red sandstone and slabby beds of red, worm-marked, argillaceous sandstone. Both the fresh and weathered surfaces of these beds are red or maroon in contrast to the light-colored quartzite composing the Troy in its type locality to the south. However, in a remnant of Troy quartzite exposed below the Martin limestone in Powers Gulch, just west of the Globe quadrangle, the slabby red sandstone grades upward into pebbly quartzite, which, though iron-stained, is commonly white on fresh fractures and is composed almost entirely of quartz, as it is in the type locality.

MARTIN LIMESTONE

The limestones that overlie the Troy quartzite in this region are in part of Devonian and in part of Carboniferous age. During his first work in the Globe quadrangle, Ransome (1903, pp. 39–46) included them all in the Globe limestone. Actually, as it was mapped by Ransome, the Globe limestone included all the limestones in the Globe quadrangle that are younger than the Dripping Spring quartzite; although he recognized that his Globe limestone included rocks of both Devonian and Carboniferous age, he did not realize until his later work (Ransome, 1919, p. 42) that he had also included the Mescal limestone of the Apache group. During this later work, Ransome (1919, pp. 45–48) divided the Paleozoic part of the Globe limestone into two formations, the Martin limestone of Devonian age and the Tornado limestone of Carboniferous age, thereby automatically discarding the name "Globe."

The Martin limestone was originally named in the Bisbee quadrangle (Ransome, 1904, p. 33). The Upper Devonian fauna collected by Ransome from the lower part of his Globe limestone justifies the extension of the name "Martin" from the Bisbee and Ray quadrangles to the Globe. As described by Ransome, the lower half of the Martin limestone in the Globe and Ray quadrangles contains beds of hard gray limestone, generally less than 2 feet thick, that include abundant small chert concretions. The basal bed is cross-bedded calcareous grit. A layer of sandy limestone, 15 feet thick, near the middle of the formation separates the unfossiliferous limestone in the lower part from the fossiliferous gray and yellow limestone beds that compose the upper part. The limestones are generally magnesian. Fissile shale, about 20 feet thick, is the uppermost member of the formation.

In the Ray quadrangle, south of the Castle Dome area, Ransome estimated that the total thickness of the Martin limestone was between 300 and 350 feet, which is approximately its thickness in the Castle Dome area. Near Sleeping Beauty Peak, 6 miles east-northeast of Castle Dome, its thickness differs greatly, and it thins from a maximum of 350 feet to a minimum of 150 feet within a distance of 2,000 feet.

As described by Darton (1925, pp. 52–53), an interval of erosion preceded the deposition of the Martin limestone in the Mescal Range. Near Superior, also, 11 miles southwest of Castle Dome, there is a disconformity between the Troy quartzite and Martin limestone (Short, 1943, p. 26). Moreover, in the Castle Dome area and near Sleeping Beauty Peak to the east, there is a great disconformity between the Martin limestone and older rocks. Two thin remnants west of the Castle Dome mine are all that remains of the Troy quartzite; generally, except where diabase is intruded beneath it, the Martin limestone lies on the eroded surface of the Dripping Spring quartzite. A conglomerate at the base of the Martin limestone contains angular blocks of quartzites and Barnes conglomerate of the Apache group; where the Martin overlies the Troy quartzite on the west side of Gold Gulch, the formation also contains a few blocks derived from the Troy. Near Sleeping Beauty Peak, the erosion surface on which the Martin was deposited had a relief of at least 200 feet. There the Dripping Spring quartzite under the limestone ranges from 200 to 500 feet in thickness. Where the quartzite is thin, the Martin limestone is thick and includes a basal conglomerate made up of angular fragments derived from the quartzite; where the quartzite is thick, the Martin limestone is thinner and lacks the basal conglomerate.

The most extensive exposures of Martin limestone and the only complete sections through it are along the west side of Gold Gulch. The Martin also crops out in small fault blocks surrounded by diabase and Tertiary sedimentary rocks in the northwestern part of the map area; east of Porphyry Mountain, it is exposed on the south side of Jewel Hill and north of the Continental mine.

Lithologically, the Martin limestone is uniform over wide areas; it can be subdivided into persistent lithologic units that have been recognized wherever the formation has been observed in the northwestern part of the Globe quadrangle. The following description indicates the units separable in the formation along the west side of Gold Gulch, where an unbroken section can be observed.

Section of Martin limestone on west side of Gold Gulch

Base of Mississippian Escabrosa limestone.

	Feet
Buff, medium-bedded, impure limestone and thin-bedded marl	45
Gray or buff paper shale	25–30
Thin- or medium-bedded, gray or brown limestone; partly sandy; fossiliferous throughout. At the top, 10 feet of massive, fine-grained buff limestone; at the bottom, 12 feet of very thin bedded gray limestone	70
Brown or gray, cross-bedded, coarse-grained quartz sandstone. Some pebbly lenses. Grains well sorted, rounded, and frosted	30–35
Medium- or thin-bedded, brownish, sandy magnesian limestone. Contains small chert lenses and nodules and a very few fossils	135–155
Brown basal conglomerate and sandstone. Lenticular beds. Pebbles and blocks of quartzite and Barnes conglomerate compose the conglomerate	5–25

Disconformity.

Apache group.

The character of the units described differs only a little throughout the map area. On the south side of Jewel Hill, the paper-shale unit is split by a thin limestone into a lower shale about 6 feet thick and an upper shale about 30 feet thick. In the horseshoe bend in Gold Gulch a second brown quartz sandstone 12 feet thick occurs 25 feet below the sandstone unit at the base of the fossiliferous limestones.

Near Sleeping Beauty Peak, about 6 miles east-northeast of Castle Dome, the same lithologic units can be distinguished, but there the shale is thicker and the marly beds above it are much thinner than in the Castle Dome area. The total thickness of the formation differs greatly from place to place, ranging from 150 to 350 feet. Where it is thin, the same lithologic units are recognizable, though thinner, and most of the thinning occurs in the lower sandy magnesian limestones. The basal conglomerate is present only where the formation is thickest.

ESCABROSA LIMESTONE

The Escabrosa limestone is the lower part of the formation, named by Ransome (1919, pp. 47–48) the Tornado limestone from Tornado Peak in the southeastern part of the Ray quadrangle. The Tornado limestone as described by Ransome conformably overlies the Martin limestone and is equivalent to the upper part of the sequence of rocks that he previously included in the Globe limestone. According to Ransome, the lower 75 feet of the Tornado consists of thin- to medium-bedded light- and dark-gray limestones, but this unit behaves as a single massive layer and forms cliffs in the present topography. Above this is a very massive, light-gray, cliff-forming limestone at least 100 feet thick. The remainder of the formation consists of thinner-bedded limestones with a few calcareous shale layers. Fossils indicate that the lower part of Ransome's Tornado limestone is lower Mississippian and equivalent to the beds of the Bisbee region that he named the Escabrosa limestone. The upper part of Ransome's Tornado limestone is Pennsylvanian and equivalent to the lower part of the Naco limestone. As stated by Darton (1925, p. 66), a disconformity at some horizon in Ransome's Tornado limestone is indicated by the absence of upper Mississippian strata.

Detailed measurements made by N. P. Peterson in 1942 show that the thickness of Ransome's Tornado limestone on Tornado Peak is about 1,000 feet and that the total thickness exposed in the general area is at least 2,000 feet. Approximately the lower 550 feet is Mississippian. A chert pebble conglomerate, 3 feet thick, was chosen as the probable division between strata of Mississippian and Pennsylvanian age, as fusulinids occur in beds about 100 feet above this conglomerate.

In the Superior area, Ransome's Tornado limestone has been subdivided into the Mississippian Escabrosa limestone, about 420 feet thick, and the Pennsylvanian Naco limestone, at least 1,200 feet thick (Short and others, 1943, pp. 29–33). At Superior the two formations are separated by a thin conglomerate containing pebbles of jasper, quartzite, limestone, and schist.

In the Castle Dome area, the Tornado limestone has likewise been subdivided into the Escabrosa limestone and the Naco limestone. Throughout the area the Escabrosa is about 365 feet thick, and no conclusive field evidence of a disconformity between it and the Naco has been found.

In the Castle Dome area the Escabrosa conformably overlies the Martin limestone. The distinction between them, as they have been mapped, is lithologic. The base of a massive, dark-gray limestone containing abundant small fossil fragments was chosen as the contact between them. This bed overlies the buff Devonian marl and is overlain by gray, massive, pure limestone. Ransome (1919, p. 45) placed the contact slightly lower, at the top of the shale member; near Sleeping Beauty Peak, however, well-preserved specimens of *Atrypa reticularis* occur in the buff limestone above the shale.

The distribution of Mississippian Escabrosa limestone in the Castle Dome area is the same as that of the underlying Martin limestone of Devonian age. The Escabrosa is composed of pure limestone containing many fossil fragments but only a few well-preserved corals and brachiopods. Throughout the map area, its thickness is about 365 feet, and its lithologic character is uniform. The lower two-thirds of the Escabrosa contains massive beds that commonly form cliffs, whereas the upper third is thinner-bedded and less resistant to erosion. The following section, measured on Jewel Hill, is representative of the formation throughout the area mapped.

Section of Escabrosa limestone on Jewel Hill

Base of Pennsylvanian Naco limestone (red shale containing lenticular beds of chert breccia).
Probable disconformity.

	Feet
Thin- to medium-bedded gray limestone containing abundant small fossil fragments. Lower 60–70 feet dark gray, weathering either dark gray or brown; upper part lighter gray, weathering nearly white and containing small, irregular masses of chert	100
Dark-gray, medium-bedded oolitic limestone that behaves as a massive member and generally forms a cliff. Weathers light gray. Oolites silicified along many small cracks	30
Thin-bedded limestone composed of interbedded dark-gray, oolitic limestone, white or buff crystalline limestone, and small chert lenses	25
Massive, cliff-forming, dark-gray limestone. Lower half oolitic, without distinct bedding; weathers medium gray. Upper half medium-bedded; weathers lighter gray than that below. Some beds contain abundant chert nodules; others, fragments of corals and crinoids	85
Dark-gray, medium-bedded limestone. Upper half contains interbedded layers of white crystalline limestone and weathers light gray. Lower half weathers dark gray and contains abundant fossil fragments and a little chert	95
Massive, medium-gray limestone, generally forming a cliff. Lower part bedded but behaves as a massive unit; contains fossil fragments and weathers brownish gray. Upper part a single bed; weathers light gray	35
Single massive bed of dark-gray limestone, slightly oolitic, containing very abundant fossil fragments	5–10

Top of Martin limestone (marl).

NACO LIMESTONE

In contrast to the underlying Escabrosa limestone, the Pennsylvanian Naco limestone is thin-bedded. Some of the beds are pure limestone as much as 3 feet thick, but most are thinner and are separated by thin layers of gray shale. Others contain abundant chert. Some distinctive beds are fragmental and consist of nodules of gray limestone surrounded by a matrix of shale. The base of the Naco limestone was mapped at the bottom of a red shale layer that is as much as 20 feet thick and in many places contains lenticular beds of chert breccia. Whether or not this bed separates rocks of Mississippian and Pennsylvanian age is uncertain, but it is the only lithologic division persistently recognized. Fusulinids occur in beds about 75 feet above the red shale. There is no direct field evidence of a disconformity between the Naco and Escabrosa limestones in the Castle Dome area; however, the basal red shale and chert breccia of the

Naco limestone may be the residual debris that accumulated on the weathered surface of the Escabrosa limestone before the Naco limestone was deposited.

MESOZOIC (?) AND EARLY TERTIARY (?) IGNEOUS ROCKS

GRANODIORITE

South of the Castle Dome mine a long narrow body of biotite granodiorite lies between the Pinal schist to the south and the quartz monzonite to the north. It tapers to a dike and ends east of the concentrator; its west end is an intrusive breccia consisting of numerous small blocks of granodiorite included in quartz monzonite porphyry. This breccia is well exposed in and near adit 2 (pl. 3) and affords the only evidence that the granodiorite is older than the quartz monzonite. How much older it is cannot be determined; it may be simply an early phase of the quartz monzonite intrusion. That the granodiorite has been intruded by the granite prophyry, however, is certain. Granite porphyry masses have been intruded along the contact between granodiorite and quartz monzonite and into the intrusive breccia exposed near adit 2.

In hand specimens the granodiorite is a light-colored, medium-grained rock that is generally equigranular but shows a porphyritic texture in some places near the margins of the body. The only minerals evident are plagioclase, quartz, and glistening black biotite. Thin sections show that orthoclase also is present but is subordinate to plagioclase and nowhere composes more than 20 percent of the rock. The plagioclase, which is generally zoned, has the composition of oligoclase, containing 70 to 80 percent albite molecule. Minor accessory minerals include zircon, apatite, magnetite, and a little sphene. The porphyritic texture seen along the marginal parts of the granodiorite mass is not pronounced, and thin sections show that this phase of the rock differs from the other only in the presence of a small amount of fine-grained groundmass composed largely of orthoclase and quartz.

QUARTZ MONZONITE

GENERAL CHARACTER AND RELATIONSHIPS

Two masses of biotite-quartz monzonite are exposed in the Castle Dome area. The larger of the two is in the central part of the map area, where quartz monzonite crops out for about a half-mile south, east, and west of the summit of Porphyry Mountain and for more than a mile to the north. This mass includes the entire Castle Dome ore body. The smaller mass, which is largely south of the map area, is separated from the quartz monzonite in the Porphyry Mountain area by a northeastward-trending belt of Pinal schist about half a mile wide and is intruded by the Schultze granite (Ransome, 1903, pp. 67–73), which forms the southern boundary of the quartz monzonite. The two quartz monzonite masses are undoubtedly related intrusions.

Petrologically, the quartz monzonite in the Castle Dome area is identical with the mass 5 miles to the northeast, which was named Lost Gulch monzonite by Ransome (1903, pp. 75–78). As mapped by him, the Lost Gulch monzonite includes a variety of intrusive rocks, each of which also occurs in the Castle Dome area. There is no visible connection between the quartz monzonite in these two areas, but they probably should be correlated.

In his first report on the Globe quadrangle, Ransome (1903, p. 72 and pl. 1) considered all the quartz monzonite in the Castle Dome area a part of the Schultze granite of Mesozoic or Tertiary age. Later he correlated the mass in the Porphyry Mountain area with the pre-Cambrian Ruin granite that crops out in Ruin Basin about 6 miles to the northeast, but he continued to regard the southern mass as a part of the Schultze granite (Ransome, 1919, p. 38 and pl. 2). He considered the Ruin granite to be pre-Cambrian in age. It is now clear that the quartz monzonite on Porphyry Mountain, which Ransome correlated with the Ruin granite, is post-Apache in age. In the Globe-Miami district the name "Ruin granite" signifies a granitic rock of pre-Apache age and should be restricted to the rock that crops out in the Ruin Basin and to other petrologically similar rocks of pre-Cambrian age that crop out in the northern part of the Globe quadrangle.

DESCRIPTION

The quartz monzonite has a generally uniform mineral composition, essentially quartz (30 to 35 percent), orthoclase (20 to 25 percent), and oligoclase (30 percent). Subhedral books of dark-brown biotite, commonly including oriented needles of rutile, are characteristic of the rock, but they are not abundant. Minor accessory minerals are zircon, sphene, apatite, ilmenite, or titanian magnetite. Zircon is decidedly rare and occurs in very small crystals, whereas the other accessory minerals are more common and are generally associated with biotite. Everywhere the quartz monzonite contains large phenocrysts of pink or reddish-brown microperthitic orthoclase that are almost equant in form and range from 1 to 3 inches in length. Most of the orthoclase occurs as large phenocrysts, many of which poikilitically enclose small crystals of oligoclase in a zonal arrangement or have rims of oligoclase (rapakivi structure). Without exception the phenocrysts rimmed by oligoclase have round or oval cross sections, whereas the unrimmed phenocrysts are euhedral.

Although the general composition of the quartz monzonite appears to be uniform throughout the map area, there are some local variations in the relative amounts of the constituent minerals. For example, in a few small areas in the Castle Dome mine, large orthoclase phenocrysts are concentrated and form 50 to 75 percent of the rock; in other areas, biotite and oligoclase are the most abundant minerals.

Primary linear or planar structures shown either by orientation of individual minerals or by variations in the distribution of the minerals have not been recognized anywhere.

Two widespread textural varieties of quartz monzonite have been distinguished and separately mapped, but although they differ in texture the mineral composition of the two is approximately the same. One is porphyritic quartz monzonite containing large phenocrysts of pink or reddish orthoclase in a coarse-grained groundmass, and the other is quartz monzonite porphyry in which large orthoclase phenocrysts, together with smaller phenocrysts of oligoclase, quartz, and biotite, are enclosed in a fine-grained groundmass. The contact between them is sharp in many places; elsewhere it is gradational but is generally definite outside the area of intense hydrothermal alteration. Quartz monzonite porphyry occurs in the central part of the Porphyry Mountain area and is nearly surrounded by the porphyritic quartz monzonite. Ransome (1903, p. 72) also noted a change from "granitite porphyry" to "porphyritic granitite" northward from Porphyry Mountain within what is now mapped as quartz monzonite.

The quartz monzonite porphyry itself shows considerable difference in texture. In the central part, on Porphyry Mountain and throughout a large part of the Castle Dome mine, phenocrysts compose no more than 25 percent of the rock, and the groundmass is very fine grained and very distinct, but nearer its margins phenocrysts predominate, and the groundmass is fine- or medium-grained and not readily apparent in every specimen. Thus the quartz monzonite porphyry mass is crudely zoned, having relatively few phenocrysts and a very fine groundmass near the center and many more phenocrysts and a slightly coarser groundmass nearer the margins.

Probably the quartz monzonite porphyry and the porphyritic quartz monzonite represent separate intrusive pulses of the same magma, the porphyry being slightly the younger. That the quartz monzonite porphyry intrudes the porphyritic quartz monzonite is suggested by its textural variation from the center outward and by the fact that it is nearly surrounded by the porphyritic quartz monzonite. Several small dikes of the porphyry occur in the porphyritic quartz monzonite, one near the northwest corner of the mass and another east of the mine at the southeast extremity of the mass, and what appear to be blocks of porphyritic quartz monzonite occur in several places in the quartz monzonite porphyry. In the western part of the mine, the contact between the two types is complicated by flat or low-dipping sill-like offshoots of the porphyry into the porphyritic quartz monzonite.

Dikes and small masses ranging from aplite to alaskite porphyry have been intruded into both types of quartz monzonite and are most abundant near the contacts between the two. The aplite is fine- to medium-grained and equigranular; it contains pink orthoclase, oligoclase, quartz, and in some places a little biotite. The alaskite porphyry has approximately the same mineral composition, but it contains a few phenocrysts. Most of the phenocrysts are oligoclase and quartz, but there are some large orthoclase phenocrysts and smaller ones of biotite. Except for the fewer phenocrysts and the less abundant biotite, the alaskite porphyry is not clearly distinct from the quartz monzonite porphyry, and in some places the two seem gradational. Probably they are very closely related. Neither occurs in the southern area of quartz monzonite.

No dikes nor veins of pegmatite related to the quartz monzonite have been recognized. Although pegmatites are numerous in the southern area of quartz monzonite, they are probably all related to the younger Schultze granite, which they also intrude.

RECRYSTALLIZATION OF BIOTITE

The original biotite that characterizes the quartz monzonite is present as dark-brown subhedral books, but within and south of the Castle Dome mine much of it is recrystallized to aggregates of small biotite plates, a few of which are intergrown with muscovite. The biotite aggregates are about the same size as the books (a tenth to a fourth of an inch in diameter), and they have about the same distribution in the rock. Invariably they are associated, as are the books, with small crystals of apatite, ilmenite, and sphene. In thin section, various stages in the change from book to aggregate biotite can be observed. Some aggregates have a core composed of a remnant of the original book surrounded by numerous small biotite crystals with various orientations; others are completely recrystallized. A few biotite aggregates form centers, from which small trains of tiny biotite crystals extend as much as a tenth of an inch into the surrounding quartz and feldspars. In the quartz monzonite, near some of its contacts with granite porphyry south of the mine, aggregates of biotite can be seen, but fine-grained biotite is scattered throughout the rock in veinlets that cut through all the other minerals.

The composition of the biotite in books and in aggregates may differ slightly, but the optical properties of the two are essentially the same.

Both show pleochroism in which X is straw-colored and Y and Z are greenish brown; the optic angle ($2V$) ranges from 5° to 20°, and the index of refraction for the slow ray (gamma) ranges between 1.615 and 1.630, though the lower values tend to be in the aggregate biotite. Most of the original books contain needlelike rutile inclusions oriented in three directions at 60° in the basal plane (001). These inclusions have not been noted in the aggregate biotite seen in thin section.

Aggregate biotite is not present in the rock north of the mine, but it increases in abundance toward the south. Near Gold Gulch and on South Hill, biotite occurs only in the form of aggregates, although the biotite in the granodiorite and granite porphyry in that area is in the form of thin plates, none of which appear to have been recrystallized.

The cause of the recrystallization is uncertain. At first it was thought that the recrystallization was an effect of hydrothermal alteration related to the copper mineralization, and a record of its distribution was made during the study of alteration in the mine. However, the distribution of aggregate biotite does not conform to the zoning pattern in the hypogene copper mineralization or the associated hydrothermal alteration, although it is confined entirely to the mineralized area. Furthermore, both types of biotite are similarly affected by hydrothermal alteration. Perhaps the recrystallization is a metamorphic effect produced by the granite porphyry intrusion near the south margin of the quartz monzonite, or if, as is likely (see p. 31), the source of the quartz monzonite was south of Porphyry Mountain, the recrystallization of biotite may be a deuteric change. S. G. Lasky, of the Geological Survey, reports in an unpublished manuscript that in the Little Hatchet Mountains in New Mexico he found such recrystallized biotite to be a metamorphic effect, both endomorphic and exomorphic.

METAMORPHISM PRODUCED BY THE QUARTZ MONZONITE

Formations are metamorphosed near their contacts with the porphyritic quartz monzonite north of Porphyry Mountain. The most widespread metamorphic change was the recrystallization of the matrix in the Scanlan conglomerate and the arkose of the Pioneer formation to microcrystalline or fine-grained orthoclase. Original sand grains composed of quartz and of weathered orthoclase generally remain in the matrix, but they are surrounded and corroded by a fine-grained aggregate of clear orthoclase. In parts of a few thin sections, no relic sand grains remain. The larger pebbles of orthoclase, quartz, and schist also are veined and replaced by the fine-grained, clear orthoclase, and many of the schist pebbles are so impregnated by it

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as to become scarcely recognizable. Probably some of the smaller schist pebbles have been completely destroyed. In most thin sections the recrystallized orthoclase contains sericite, which is distributed through the rock as small flakes and rosettes. Sericite veinlets also are common in the pebbles, and a few of them cut through the matrix of recrystallized orthoclase.

Clusters of 10 or more apatite crystals are numerous in all the metamorphosed rocks; in some, the clusters are grouped in veinlike zones. There is also some apatite in well-defined veinlets that cut quartz and orthoclase pebbles. It is commonly associated with aggregates of tiny crystals of rutile or sphene that are coated by leucoxene. Many of the apatite clusters contain slender prisms and radiating aggregates of pale-blue tourmaline, and a few crystals of apatite and tourmaline are intergrown. Tourmaline is present in pebbles of schist; in them it might be original, but it also occurs in quartz and orthoclase pebbles and in the recrystallized orthoclase matrix. All of it is pale blue in thin section.

Thin sections show that the coarse red arkose between the Scanlan conglomerate and quartz monzonite north of Porphyry Mountain has also been metamorphosed. The coarse-grained quartz and red orthoclase, which are the original constituents of the rock, are veined and partly replaced by fine-grained orthoclase, and apatite and tourmaline occur in the arkose as they do in the overlying Scanlan conglomerate. In several specimens, numerous veinlets of microcrystalline orthoclase containing large crystals of apatite and plates of specularite cut the original grains of quartz and orthoclase.

The thin black member of the Scanlan conglomerate contains scattered angular pebbles of vein quartz and orthoclase enclosed in a matrix that ranges from very fine sand or silt to shale. Its black color is caused by very abundant, finely disseminated specularite, most of which is in the shale matrix. The specularite is so abundant that parts of some thin sections are almost opaque. A few veinlets of specularite cut across the laminations of the rock, and these in turn are followed or cut by tiny stringers of quartz and chlorite. Such an abundance of specularite in one thin layer, particularly in the fine-grained matrix, suggests that it may have developed by the recrystallization of ferric oxide in the shale member of the Scanlan conglomerate.

About 2,000 feet north of the Continental mine, recrystallization of the lower arkosic part of the Pioneer formation has produced a micrographic intergrowth of orthoclase and quartz. Although the relationship of the recrystallized arkose to the surrounding rocks is obscured by talus, there can be little doubt that the quartz monzonite intrusion caused the recrystallization. Micrographic intergrowth is the dominant texture of the recrystallized rock, occurring throughout the smaller orthoclase grains and along the margins of, or as vein-like zones through, the large grains.

Biotite is scattered throughout the rock in broad but very thin plates and occurs also as rosettes and veinlets of small crystals replacing both orthoclase and quartz. Probably all of it is metamorphic, since it has not been seen in any of the unaltered sandstone of the Pioneer formation. Some of the biotite has altered to chlorite, but whether or not all the chlorite in the rock has been derived in this manner is uncertain. Scattered plates of specularite are numerous, and some are intergrown with biotite. Small apatite prisms, generally included in other minerals, and small clusters of rutile crystals are common.

Both megascopically and in thin section the recrystallized arkose distinctly shows a gradational decrease in grain size from its base near the quartz monzonite to the top, where typical fine-grained sandstone of the Pioneer formation is exposed. At the bottom, round or oval grains of quartz as much as half an inch long are probably relic quartz pebbles. All are quartz aggregates, and although many of them were probably pebbles of vein quartz, some appear to have relic clastic structure and may have been fine-grained quartzite. They are corroded and penetrated by orthoclase, and they grade into the surrounding matrix of intergrown orthoclase and quartz. A few of the pebbles are coated by thin films of biotite.

No relic pebbles can be seen more than a few feet above the base of the recrystallized rock, even though the degree of metamorphism decreases upward, which suggests that the rocks affected by metamorphism include the basal pebbly member of the Pioneer formation as well as some of the finer sandstones above. Likewise, the micrographic intergrowth that has been developed in the rock is much coarser grained at the bottom than at the top. The uppermost layers showing recrystallization are composed of small interlocking crystals of quartz and orthoclase, and only a little very fine micrographic intergrowth is evident.

INTRUSION OF THE QUARTZ MONZONITE

Little is certain about the manner of intrusion and the original shape of the quartz monzonite complex. In the Porphyry Mountain area, the present somewhat rectilinear outcrop of the quartz monzonite may not reflect the original outlines of the intrusion, as it is now confined to a north-northwestward-trending horst. Its east and west boundaries are also the boundary faults of the horst. Diabase was later intruded along these boundary faults and across the horst near the north edge of the mapped area, so that the quartz monzonite is now nearly surrounded by younger diabase. North of Porphyry Mountain part of the original roof of the complex remains and gently tilted beds of the Apache group cover the quartz monzonite. Along the northwest side of the complex, also, the original contact remains, but there—even though concordant—it is steep. In Gold Gulch, west of the mine, quartz monzonite is in discordant intrusive contact with a small block of sandstone of the Pioneer formation, and south of the mine it is intruded into granodiorite and Pinal schist along contacts dipping steeply to the south. This is meager evidence on which to base any assumption concerning the original form of the quartz monzonite complex. The steep contacts south of Porphyry Mountain and the concordant, gently dipping roof to the north may indicate that the quartz monzonite complex has its root near and south of Porphyry Mountain and spreads northward beneath the formations of the Apache group as a thick sill.

The intrusion of the quartz monzonite was multiple; the porphyritic quartz monzonite phase was followed by the quartz monzonite porphyry phase and later dikes of alaskite porphyry and aplite. The main mass of quartz monzonite porphyry is centered around Porphyry Mountain and, except where it has been intruded by younger diabase east of the mine, is surrounded by porphyritic quartz monzonite and older rocks. Certainly the original porphyry mass was little larger than that now exposed, and it may be described as a chonolith or irregular stock within the quartz monzonite complex. Its southern contact has a steep southerly dip outward from the center of intrusion. The relation to the topography of many other parts of the contact suggests that the contact generally dips outward from the center, but because of its irregularity and the complications introduced by later faulting, the actual attitude of the contact is uncertain.

A smaller mass of quartz monzonite porphyry underlies South Hill and is separated from the larger mass on Porphyry Mountain by porphyritic quartz monzonite. Remnants of porphyritic quartz monzonite on the north slope of South Hill have a relation to the topography that suggests that they are parts of a northward-inclined roof over the quartz monzonite porphyry. The porphyry masses on South Hill and Porphyry Mountain are undoubtedly connected below the present surface.

The locus of several intrusions was between Porphyry Mountain and the Pinal schist to the south. The first, a mass of granodiorite elongated in a general easterly direction, was intruded along what is now the north boundary of the schist. Later, both porphyritic quartz monzonite and quartz monzonite porphyry were intruded north and west of the granodiorite, and although both these rocks extend for more than a mile to the north, at least the quartz monzonite porphyry intrusion seems to have centered near the south end of the area. Still later, dikes and small masses of granite porphyry, elongated and alined in a general easterly direction, were intruded along the north boundary of the granodiorite and into the adjacent quartz monzonite and diabase. Because this has been a zone of repeated intrusive activity, it seems likely that the quartz monzonite was here connected with a deep-seated magma reservoir. If such an assumption is correct, the mass of porphyritic quartz monzonite north of Porphyry Mountain, which was intruded concordantly beneath the Scanlan conglomerate, is probably a sill at least 500 feet thick that was fed by a discordant root near or south of Porphyry Mountain.

The highest elevations in the central part of the quartz monzonite are several hundred feet above the gently tilted Scanlan conglomerate and Pioneer formation, which form the concordant roof of the intrusion north of Porphyry Mountain. The absence of this roof over the highest parts of the quartz monzonite may be accounted for in any or all of three ways: (1) faults in the quartz monzonite north of Porphyry Mountain along which the quartz monzonite to the south was elevated relative to that farther north beneath the formations of the Apache group, (2) doming of the overlying rocks by the intrusion in the central part of the area, and (3) a rise of the magma to a higher stratigraphic position in the central part of the area. The correct explanation for the absence of a sedimentary roof over Porphyry Mountain is uncertain. However, at least minor faulting in the quartz monzonite occurred north of Porphyry Mountain and has relatively elevated the area to the south. In Gold Gulch, west of the mine, the quartz monzonite has a discordant intrusive contact with the Pioneer formation, showing that there at least the magma penetrated the rocks above the Scanlan conglomerate.

The thickness of the rock that covered the quartz monzonite at the time of its intrusion can be roughly estimated because the quartz monzonite was intruded beneath and into sedimentary formations that were only slightly deformed. Therefore, the thickness of the stratigraphic column that may have existed above the quartz monzonite at the time of intrusion gives an approximate thickness for the roof. This estimate cannot be accurate because the age of the intrusion is uncertain and because the stratigraphic position to which the magma rose is known only along the north edge of the intrusion and may have been different elsewhere; moreover, erosion has removed some of the sedimentary rocks that once covered the region, making their original thickness uncertain. However, by assuming that the magma penetrated only to the base of the Apache group and using a reasonable maximum estimate for the thickness of the stratigraphic column, the approximate maximum thickness of rock over the intrusion is obtained. If the quartz monzonite was intruded during the Mesozoic era, as is now assumed, the overlying rocks consisted of the Apache group and the Paleozoic rocks now exposed, a total of about 1,700 feet plus the thickness of Pennsylvanian and possibly Permian and Mesozoic rocks since removed by erosion. The thickness of the cover above the intrusion may have been as much as 5,000 feet but was probably less. If the intrusion was pre-Devonian, the overlying rocks may have included all the formations of the Apache group except the Mescal limestone removed by pre-Troy erosion, plus the Troy quartzite. The total thickness of these formations was probably less than 2,000 feet.

AGE OF THE QUARTZ MONZONITE

The evidence is conclusive that the quartz monzonite is post-Apache in age. The rocks of the Apache group have been metamorphosed near their contacts with the quartz monzonite. Blocks of the arkose below the Scanlan conglomerate are included in the quartz monzonite, and one inclusion of Scanlan conglomerate occurs in it about 400 feet northeast of the Continental mine. A discordant intrusive contact between quartz monzonite and the Pioneer formation is exposed in Gold Gulch west of the Castle Dome mine. In the petrographically similar quartz monzonite south of Sleeping Beauty Peak, about 6 miles east-northeast of Castle Dome, inclusions that are probably Scanlan conglomerate have been found.

The quartz monzonite has been intruded by post-Pennsylvanian diabase, but nowhere is it in contact with the Paleozoic limestones present both east and west of it. Thus the age of the quartz monzonite cannot be determined accurately. Since it is post-Apache, and since it is unlikely that such an intrusion occurred while the Devonian and Carboniferous limestones were being deposited, it must be early Paleozoic, Mesozoic, or Tertiary in age. It has been assigned tentatively to the Cretaceous or early Tertiary orogeny because early Paleozoic intrusions have not as yet been recognized anywhere in Arizona. Ransome (1919, pp. 51–52) suggested a Mesozoic age for the Lost Gulch monzonite, which is probably to be correlated with the quartz monzonite in the Castle Dome area, but his evidence also is inconclusive.

DIABASE

MAIN DIABASE INTRUSION

Large masses of coarse- or medium-grained diabase crop out along the east, west, and north sides of the quartz monzonite. North of the map area diabase is widespread, but it does not crop out farther south. It has been deeply decomposed and is less resistant to erosion than any of the other formations. Generally it is found in the relatively low lying parts of the map area and crops out poorly, much of it being covered by talus from more resistant rocks. It is well exposed in most of the recently rejuvenated stream channels, but in many of these the only fresh diabase is in small spheroidal kernels surrounded by disintegrated rock. Where diabase crops out on gentle slopes, the ground is soft and is composed of sandy debris produced by disintegration, but the surface is commonly strewn with small blocks and cobbles of fresher rock that have weathered out and been concentrated there.

Ransome (1919, pp. 53–55) described the diabase composing the larger masses throughout both the Globe and Ray quadrangles as a typical olivine diabase. From a study of thin sections, he estimated the average composition to be 55 percent calcic labradorite or bytownite, 30 percent augite, 10 percent olivine, and 5 percent biotite, magnetite, apatite, and sphene. The texture is ophitic, and much of the augite is poikilitic. In some places the augite has been uralitized and the olivine converted to serpentine. Short (Short and others, 1943, pp. 35–37) considers the diabase in the Superior area west of the Globe quadrangle to be the same as that described by Ransome, but near Superior he found three types of diabase: quartz-orthoclase diabase, augite-hornblende diabase, and olivine augite diabase. He thinks all these types were derived by differentiation within a single magma chamber.

The petrography of the diabase throughout the Castle Dome area has not been studied in detail. Only a few specimens, all collected east of the mine, have been examined in thin section, and all of these contain about 60 percent labradorite (Ab_{40-50}), together with abundant green hornblende and a little quartz, biotite, apatite, and titanian magnetite. Two of the specimens contain orthoclase as well as quartz, and in them the two minerals together compose 10 to 15 percent of the rock. A little of the quartz and orthoclase are micrographically intergrown, but most occurs as discrete grains molded by the labradorite and hornblende. Needles of apatite occur in the quartz. In one specimen, the hornblende clearly replaces augite, which remains as unaltered cores in many of the hornblende crystals. The texture of all the specimens is ophitic, the hornblende filling the spaces between euhedral laths of labradorite. The rock is regarded as a quartz-bearing diabase in which original augite has been largely replaced by hornblende.

Although the specimens that were examined microscopically do not represent all the diabase in the Castle Dome area, they indicate clearly that Ransome's classification of the rock as typical olivine diabase requires modification. In the field, the diabase appears essentially uniform throughout the area, but careful examination with a hand lens and by immersion methods under a microscope shows that it ranges from hornblende diabase containing little or no augite to augite diabase containing no hornblende. The plagioclase ranges from calcic to sodic labradorite (Ab_{30-50}) and is most calcic in some of the augite diabase. The augite seen in thin section is either colorless or slightly purplish, suggesting that some of it is titanian. In hand specimens it is black and commonly appears glassy where it is fresh. It occurs as small grains between the euhedral labradorite crystals and as scattered large crystals that poikilitically enclose the labradorite. In contrast, the hornblende seen in hand specimens is greenish and shows glistening cleavage surfaces. Minor constituents visible everywhere include biotite, apatite, and magnetite. The apatite occurs as needles as much as half an inch long in some specimens, but in others the needles are smaller and are difficult to recognize. No olivine has been recognized, and quartz has not been seen in hand specimens. It is probable that the diabase in the Castle Dome area, as in the area near Superior, ranges from quartz-bearing to olivine-bearing types.

A coarse-grained syenitic rock (Ransome, 1919, pp. 54–55), generally reddish in color, which is probably a differentiation product of the diabasic magma, is present in a few places as small masses or narrow lenses in the typical diabase. Pink perthite composes as much as 90 percent of the rock in some masses; the other constituents are hornblende, a little quartz, and numerous large apatite needles.

Probably most of the diabase is one large intrusion, which has been intruded subsequently by small dikes and masses. One sill of very fine grained diabase, not over 2 feet thick, is well exposed in a cut along the access road east of the Castle Dome mine. Thin sections show that it is a quartz-hornblende diabase not noticeably different from the larger body into which it was injected. Probably there are many similar injections, but they are noticeable only where exposures are exceptionally good. What may be a larger late injection of diabase into the main mass crops out in Gold Gulch at the bend where it turns sharply westward about 1,000 feet west of the mine. It is a medium-grained rock containing very abundant large crystals of magnetite and is distinct from the surrounding diabase only because it is less decomposed and more resistant to erosion. It forms a mass about 50 feet wide by 500 feet long, trending in a north-northwesterly direction parallel to the trend of the main mass in that area.

The diabase magma was intruded along many faults, but much of it was also intruded as sills, both large and small, between the beds and along the contacts of the sedimentary formations. Most of the magma was intruded into rocks of the Apache group, but relatively small masses of it rose higher into the Paleozoic limestones and small bodies were injected into the quartz monzonite. It was intruded at shallow depths with probably no more than the thickness of the Paleozoic limestones separating the largest diabase masses from the surface. The force of its intrusion cracked and displaced the overlying limestones. West of Gold Gulch the diabase intrusion produced faults in the overlying formations, the largest of which has a displacement of about 700 feet. It is possible that some of the diabase magma reached the surface and poured out as basalt flows, but if it did, none of the lava is now evident.

In the northern part of the map area, where the largest masses of diabase crop out, the magma was intruded along many faults, joints, and bedding planes in the Pioneer formation, and it completely surrounded many large and small blocks. The general attitude of the stratification in these separated blocks is uniform, and the blocks have not been noticeably distorted or rotated by the intrusion. The structure of the area suggests that a great volume of diabase magma was injected into shallow levels of the crust and that the crust was dilated without much distortion. There is no evidence that the diabase magma caused fusion or metamorphism of the rocks it intruded.

The diabase is younger than both the Naco limestone of Pennsylvanian age and the quartz monzonite, but it is older than the granite porphyry. As suggested by Ransome (1903, p. 86), it was probably intruded during the Mesozoic era. Darton (1925, p. 254) disagreed with Ransome concerning the age of the main diabase intrusion, which, from studies in the Mescal Mountains south of Globe, he thought was pre-Devonian.

In the Castle Dome area, Martin limestone occurs in two large inclusions in the diabase southwest of the mine, showing conclusively that the diabase is post-Devonian in age. Intrusive contacts between diabase and Martin limestone occur elsewhere both west of Gold Gulch and north of the Continental mine (pl. 1). What are interpreted as intrusive contacts between diabase and Carboniferous limestones occur in both these areas, but they are not well exposed. The most convincing evidence that the diabase is post-Pennsylvanian in age is the fact that it was intruded along faults that displaced the Naco limestone and that it has caused other faults that displace the Naco.

South of Sleeping Beauty Peak, about 6 miles east-northeast of Castle Dome, diabase was intruded along a major fault between quartz monzonite on the south and sedimentary rocks of Paleozoic age and the pre-Cambrian Apache group on the north. In at least one place along this old fault, the diabase intruded the Naco limestone and caused recrystallization in it for a distance of a few feet from the contact.

DIABASE SILLS IN THE CASTLE DOME MINE

The diabase sills shown on the map and structure sections of the Castle Dome mine (pls. 3, 4) are thin and are well exposed only on the faces between mine levels. They are a minor type of rock but were an important control in the localization of mineral deposition and the formation of the ore body. None of them has been traced as far as the edge of the quartz monzonite, and their relations to the main diabase intrusion are uncertain. No contact between them and the granite porphyry has been seen, and comparison of the petrographic character of the sills with that of the main diabase is not possible because the sills are all greatly altered. The diabase in the sills contained tabular crystals of plagioclase feldspar, which are commonly visible in hand specimens as small white laths suggesting an ophitic texture, but thin sections show that all the feldspar has been replaced by clay or, along veins, by sericite. The other constituents of the rock include scattered flakes of green biotite, much of which has also been altered, abundant small grains of titanian magnetite, and a little quartz. It is probable that some quartz was present as a primary mineral in the rock, but much of it may have been introduced during the mineralization and alteration. None of the minerals or textures visible in the sills clearly distinguishes the rock from the main diabase, but neither does any one of them definitely indicate a relationship between the two rocks.

Half a mile north of Porphyry Mountain, a thin diabase dike has been intruded into the quartz monzonite. Because it is near the outer edge of the mineralized area, it is much less altered than the sills in the mine, and many of the original constituents remain. A thin section of the diabase from this dike shows that it has ophitic texture and contains sodic labradorite (Ab_{50}), green hornblende, biotite, and a little quartz, apatite, and titanian magnetite. Petrographically, it is similar to the quartz-hornblende diabase in the main intrusion east of the mine. The green hornblende in the dike is partly altered to brownish-green biotite, and although none of the hornblende is completely replaced, most of the crystals contain thin shreds or plates of biotite oriented parallel to the cleavage planes in the hornblende. Whether this dike is to be compared with the sills in the mine is doubtful, but like the sills it is small and far removed from the main diabase intrusion. The alteration of hornblende to biotite in the dike north of the mine suggests that the green biotite seen in the sills may also have been derived by the alteration of hornblende, and the similarity between the dike to the north and the main diabase east of the mine suggests that the sills, before they were altered, may also have been similar to the main diabase.

DIORITE PORPHYRY

A porphyry containing tabular phenocrysts of plagioclase in a microcrystalline groundmass occurs as dikes and small masses in the diabase and as sills near the top of the Pioneer formation. Within the Castle Dome area, it is most common in the diabase near Gold Gulch, but two small dikes of it crop out in the diabase 400 feet northeast of the Continental mine. It generally is light gray and so deeply decomposed that it crops out only in the most favorable places. Some of it contains both large and small phenocrysts of altered hornblende, together with the plagioclase phenocrysts, and some appears nonporphyritic. Ransome (1903, pp. 86–88) called the rock diorite porphyry; he found it throughout the northern part of the Globe quadrangle. According to him, it is much altered everywhere and includes both diorite and quartzdiorite porphyry types. Several fragments of it have been found in the Whitetail conglomerate northwest of Continental Spring. It may have been intruded shortly after the intrusion of the main diabase mass.

GRANITE PORPHYRY

With the possible exception of the diabase sills in the mine area, the last intrusion in the Castle Dome area prior to mineralization is represented by numerous small intrusive bodies of biotite granite porphyry exposed south and east of Porphyry Mountain. Three small masses intruded the diabase northwest of Jewel Hill, but all the others crop out in an east-northeastward-trending zone just south of the Castle Dome mine. The largest mass was intruded along the contact between quartz monzonite and granodiorite southeast of the mine. This is a narrow, steep-walled body elongated in a general east-west direction. It extends eastward into the diabase as a dike. About 500 feet farther west, another small body was intruded along the contact between quartz monzonite and granodiorite. The westward extension of this zone of granite porphyry intrusions is indicated by numerous small dikes and irregular masses in the quartz monzonite and by two small plugs in the diabase near Gold Gulch. Several other masses in this zone are not exposed at the surface. One was intersected near the bottom of an exploratory drill hole at mine coordinate position W. 2900, A, and several small dikes are cut by adit 2 approximately at coordinate position W. 2400, B (pl. 4 *E*). It is probable that this zone of small, apparently disconnected bodies of granite porphyry represents the top of a larger mass that has not yet been exposed by erosion.

Granite porphyry occurs, also, about 6 miles east-northeast of Castle Dome, where it is intruded into quartz monzonite, the Lost Gulch monzonite of Ransome, south of Sleeping Beauty Peak. In this area, as at Castle Dome, it occurs as a series of small intrusive masses alined in a general east-west zone through the quartz monzonite. Though granite porphyry has only a small areal extent both south of Sleeping Beauty Peak and in the Castle Dome area, its occurrence in two places so widely separated suggests that it may have a wider distribution than has yet been recognized.

The granite porphyry can be recognized most readily in the field by the numerous euhedral quartz phenocrysts scattered through the microcrystalline groundmass. Especially is this true where the rock has been hydrothermally altered, because hydrothermal alteration obscures albite phenocrysts, which are more abundant in the granite porphyry than those of quartz. Phenocrysts of biotite and pink orthoclase are fewer than those of albite and quartz, and orthoclase phenocrysts are generally much larger than the others. The groundmass is microcrystalline and ranges from a granular type to a granophyric type composed of graphically intergrown quartz and orthoclase. It is everywhere composed mostly of quartz and orthoclase but also contains a little albite in some specimens. Thin sections of the granite porphyry in the large mass northwest of Jewel Hill, where the rock is little altered, show that the groundmass close to the contact with the diabase is cryptocrystalline and distinctly finer than the groundmass in specimens farther from the contact. Minor accessory minerals present in most thin sections include zircon, apatite, and a very little sphene. A minor phase of the granite porphyry contains phenocrysts of albite only and is much more difficult to recognize than the more common type containing quartz phenocrysts.

The granite porphyry is clearly younger than both the quartz monzonite and the granodiorite. It intrudes the large mass of diabase northwest and south of Jewel Hill, as is shown by the chilled borders along the contact with the diabase and by tiny inclusions of altered diabase close to the margins of the porphyry. However, it is nowhere found in contact with the fine-grained diabase sills exposed in the mine area, and because these cannot certainly be related to the large diabase intrusion, the relation between them and the granite porphyry is doubtful.

Though there is no direct evidence, the granite porphyry may be related to the Schultze granite, which is exposed south of the Castle

Dome area. Not only does the granite porphyry appear similar to the porphyry that occurs in some places along the margins of the Schultze granite, but the two intrusions are at least approximately of the same age. The Schultze granite was named by Ransome (1903, pp. 67–92) from the Schultze ranch west of Miami. It extends 8 miles southwestward from the town of Miami, covering an area of about 20 square miles (fig. 2). The north boundary of the main mass is about 1½ miles south of the Castle Dome mine, but a single small mass of it crops out 3,000 feet south of South Hill. Generally the Schultze granite is a porphyritic biotite granite consisting of scattered large orthoclase phenocrysts in a medium- or coarse-grained groundmass containing quartz, orthoclase, oligoclase (Ab_{75-80}), and biotite, but in many places along the margins of the mass there is a finer-grained granite porphyry with a microcrystalline groundmass. Oligoclase is the dominant mineral, composing about half the rock; however, the granite was classified by Ransome as a soda granite on the basis of its chemical composition.

Mineralogically, the granite porphyry in the Castle Dome area differs only slightly from the Schultze granite; it probably contains more orthoclase feldspar, and its plagioclase is a little more sodic. The two are similar in their apparent association with the copper mineralization throughout the district.

TERTIARY AND QUATERNARY SEDIMENTARY AND VOLCANIC ROCKS

WHITETAIL CONGLOMERATE

The Castle Dome area includes the type locality of the Whitetail conglomerate. Though Ransome (1903, pp. 46–47) named the formation from Whitetail Gulch, which lies just north of the map area, he selected the section exposed 4,000 feet north of Continental Spring for detailed description and illustration. There, according to Ransome, the base of the deposit is coarse, unstratified diabase debris including blocks as much as 3 feet long. Above this the deposit is better stratified and includes beds of pebbly sandstone. It is largely composed of diabase debris and is overlain by dacite tuff. Recent detailed mapping shows that the formation in this locality is faulted against the older rocks; its base is not exposed. Furthermore, although the conglomerate was buried by dacite, layers of stratified tuff are interbedded in the upper 50 feet of the conglomerate. The maximum thickness of Whitetail conglomerate measured north of Continental Spring is about 200 feet.

Wherever it has been exposed in the map area, the Whitetail conglomerate is composed of coarse, bouldery debris derived largely from diabase. Abundant ferric oxide, probably derived from the oxidation of this diabase, gives the deposit a general reddish-brown color. Commonly the deposit is not stratified; where stratification is visible, it is generally so channeled and cross-bedded that it does not represent the attitude of the deposit as a whole. Better-sorted and better-stratified sandy layers crop out in a few places; more of them can be seen in the type section north of Continental Spring than in any other section mapped.

In addition to diabase detritus, the Whitetail conglomerate contains numerous fragments, both small and very large, derived from Paleozoic limestones, rocks of the Apache group, Pinal schist, and pre-Apache granite, but in the Castle Dome area such fragments are subordinate to those of diabase. No debris from the Schultze granite has been seen in the formation, and only a very few fragments of quartz monzonite and granite porphyry have been found in it. Altogether, perhaps five or six fragments of unaltered quartz monzonite have been collected from it north of Continental Spring and west of Gold Gulch, and the sludge from a drill hole in the Whitetail conglomerate north of Jewel Hill contained several fragments of mineralized granite porphyry. The scarcity of debris derived from the quartz monzonite suggests either that the Porphyry Mountain area, including all the mineralized area, was relatively low lying and largely buried by Whitetail conglomerate or that the quartz monzonite in that area was not yet exposed when the Whitetail conglomerate was deposited. Likewise, the absence of fragments of the Schultze granite in the conglomerate may mean either that the Schultze granite, now exposed south of the Castle Dome area, had not been unroofed when the Whitetail was deposited or that drainage from the Schultze granite did not flow northward into the area where the Whitetail is now exposed. The latter possibility seems less likely, since debris from the Pinal schist is common in the Whitetail and must also have been derived largely from the general area in which the Schultze granite is now exposed.

Probably almost all the Castle Dome area was once covered by Whitetail conglomerate. This is indicated by the present extent and thickness of the conglomerate, which are greater in this area than in any other part of the Globe quadrangle. Wherever the base of the dacite is exposed in the Castle Dome area, Whitetail conglomerate crops out beneath it, with a single exception near the west edge of the map area; beyond this area, the dacite generally rests directly on older rocks, and the Whitetail has been seen only in the headwaters of Lost Gulch a few miles to the east. The thickness of the Whitetail differs greatly within the Castle Dome area. The drill hole north of Jewel Hill (pl. 2, section *H–H′*) passes through Whitetail conglomerate for nearly 500 feet, and both north and west of Porphyry Mountain many sections measure more than 200 feet in thickness.

Ransome (1919, pp. 67–68) regarded the Whitetail conglomerate as a deposit formed partly from the "stony litter of an arid surface" locally reworked by transient streams and partly as fans accumulated at the mouth of shallow gulches. He states that "the material is generally of local origin and varies with the underlying bedrock" and that it accumulated particularly on areas of diabase. The bouldery character of the deposit and its crude stratification and great range in thickness all indicate that it is an accumulation laid down by streams on an irregular surface. However, to give a more definite description of its origin, a much larger area than is included within the limits of the present map would have to be studied. Within the Castle Dome area and also in the headwaters of Lost Gulch, the Whitetail conglomerate commonly rests upon Paleozoic limestone; diabase is the underlying bedrock in only a few places. Nevertheless, the conglomerate is everywhere dominantly composed of diabase debris regardless of what rocks immediately underlie it. This is particularly striking where the Whitetail conglomerate lies on limestone, as it does southwest of Gold Gulch. Only a few beds at the base of the deposit were certainly derived from the underlying rock.

Ransome also compares the formation of the Whitetail conglomerate with the formation of the younger Gila conglomerate. The two formations are alike in their very coarse detritus and crude stratification, and undoubtedly both are stream deposits, but the composition of the Whitetail conglomerate is comparatively uniform. The Whitetail is composed largely of diabase debris throughout the map area, whereas the composition of the Gila conglomerate differs greatly and, even within small areas, can usually be subdivided into members composed of distinctly different detritus derived from different local sources.

No new information regarding the age of the Whitetail conglomerate is available. Fossils have not been found in it, but it is thought to be Tertiary or possibly in part Mesozoic. The conglomerate appears to be conformable with the overlying massive dacite, because, in its upper part, there are interbedded layers of dacite tuff. In Pinto Creek, however, about a quarter of a mile west of the southwest corner of the mapped area, Pinal schist has been thrust over Whitetail conglomerate along a fault dipping 25° to 30° W. The dacite overlaps the fault without offset and lies on the schist in the hanging wall on the west side of the fault and on Whitetail conglomerate in the footwall on the east side. In this area, there was clearly important faulting between the deposition of the Whitetail and the outpouring of the dacite. Whether this disturbance was local or widespread is uncertain, but no evidence of it has been recognized elsewhere.

DACITE

Throughout the Globe quadrangle, north and west of the Pinal Mountains, there are remnants of a silicic volcanic rock that Ransome (1903, pp. 88–94) called dacite. It occurs in massive outcrops cut by widely spaced joints, and because it resists erosion more than most of the other rocks in the area, it generally underlies the highest parts of the area in which it occurs. It makes steep slopes, and the less resistant rocks below it are commonly strewn with dacite talus. Generally it is light brown with a decidedly pinkish tinge. On fresh fractures it is rough and shows an aphanitic groundmass including numerous small phenocrysts of quartz, plagioclase, sanidine, and glistening black biotite. It would probably be called rhyolite by most geologists in the field, but Ransome found that the plagioclase phenocrysts are andesine and are much more numerous than the sanidine phenocrysts, and a chemical analysis published by him shows that the dacite contains more iron and lime than is common in rhyolites. Here and there faint flow banding can be seen, but generally the rock is massive and appears uniform throughout. In some very clean outcrops, what appear to be the shadowy outlines of dacite fragments are visible, suggesting that part of the rock is a firmly welded agglomerate. It also commonly includes a few small fragments of the older rocks of the area. A thin layer, averaging about 10 feet in thickness and composed of black porphyritic obsidian, occurs nearly everywhere at the base of the massive brown dacite. This basal layer has a flow structure that is clearly shown by numerous small lenses of shiny, black nonporphyritic glass oriented parallel to the surface on which the dacite lies, and even where the dacite covered an irregular surface, a uniformly thin layer of the black vitrophyre is usually present as the basal member.

Below the massive brown dacite and the black vitrophyre, in the Castle Dome area, is a layer 10 to 100 feet thick composed of massive, white vitric-crystal tuff. North of Continental Spring, where this tuff is thickest and best exposed, the lower part is yellowish and contains many small fragments of diabase and pumice. It is commonly interbedded with the underlying Whitetail conglomerate. The upper part is unstratified and appears to grade into the overlying dacite, so that the contact between the two is doubtful. The black vitrophyre is not present between the white tuff and brown dacite north of Continental Spring, but it is found everywhere else. Although the basal tuff occurs throughout the Castle Dome area, it is generally absent where the base of the dacite is exposed in other parts of the Globe quadrangle. It overlies the Whitetail conglomerate near the head of Lost Gulch east of the mapped area and in Pinto Creek to the west, and Ransome (1903, p. 90) reported seeing it beneath the dacite near the old Black Warrior mine north of Miami. Like the greater extent and thickness of the Whitetail conglomerate in the Castle Dome area, the presence of the basal tuff throughout the map area also suggests that this area was relatively low lying and probably was covered by Whitetail conglomerate when the dacite eruptions began.

Remnants of the dacite scattered throughout the Globe quadrangle show that a sheet of volcanic rock at least 1,000 feet thick once covered the region. Because of this thickness, it probably covered the high as well as the low parts of the area, and it undoubtedly covered what is now Porphyry Mountain. The eruptions that produced this great volcanic blanket are thought to have occurred during the Tertiary period. After the eruptions, major faulting and enough erosion to remove the dacite completely from some places occurred before the deposition of the younger Gila conglomerate. Ransome (1919, pp. 169–173) has shown that the hypogene copper mineralization near Miami was probably early Tertiary in age and that the mineralized area had been exposed and a chalcocite ore body produced by supergene enrichment before the dacite covered the Miami area. In view of these relations, the dacite may be regarded as possibly middle Tertiary in age.

GILA CONGLOMERATE

The name "Gila conglomerate" is used to designate all conglomerate and sandstone deposited after the dacite eruptions and before the last period of deformation in the region. It probably includes deposits of Pleistocene as well as Pliocene age. The largest area covered by Gila conglomerate is the valley of Pinal Creek and includes the townsites of Globe and Miami, but as shown by Ransome's map of the Globe quadrangle (1903, p. 1), remnants of the formation are present throughout the northern half of the quadrangle north of the Pinal Mountains. Two separate areas of Gila conglomerate occur in the Castle Dome area. One of these is east and southeast of the mine and extends beyond the map area for about 2 miles to the south and east. The other is in the western part of the map area, southwest of Gold Gulch, and extends beyond the limits of the map area (pl. 1) to the south and northwest.

Most of the Gila conglomerate is a valley fill laid down by streams, but great changes in the regional topography have taken place since it was deposited. Many remnants of the formation, like the one southeast of the Castle Dome mine, are now in the higher parts of the Globe region, and the formation has been cut by a number of large faults. Northward for at least 6 miles from the town of Miami, the western limit of the Gila conglomerate underlying Pinal Creek is a major fault zone (Miami fault) with a displacement of more than 1,000 feet. Between Needle Mountain and Inspiration, southeast of Castle Dome, the Gila conglomerate has been displaced by several faults that indicate 500 to 1,000 feet of movement, and the Gila west of the mine has been depressed relative to the quartz monzonite along a normal fault with a displacement of at least several hundred feet.

Within the Castle Dome area, the formation is a buff, bouldery deposit composed of debris from all the older formations in the area. Its constituents are angular or subrounded fragments ranging from pebbles to huge blocks as much as 15 feet long set in a sandy matrix lightly cemented by calcium carbonate. In some places, such as that east of the mine where the formation rests on dacite, it is well stratified and composed of sandstone and pebbly conglomerate, but generally the bedding is lenticular and obscure and the structure of the deposit is not clear. Along some of the faults cutting the formation, steep dips have been observed, but generally the deposit is subhorizontal.

Detailed mapping of the Gila conglomerate shows that the formation is not uniform and that members distinguished by different types of detritus can be separately mapped. These members interfinger and overlap and clearly indicate that the deposit is composed of coalescing alluvial fans washed into large valleys from several local sources along the margins of the valleys. The base of the conglomerate for about a mile north of Needle Mountain, 3 miles southeast of the mine, is a breccia composed entirely of angular fragments of Pinal schist; northward this member interfingers with another composed of Pinal schist, pre-Cambrian granite, and dacite; and still farther north, where it lies on dacite, the base of the conglomerate contains only dacite. These members are overlapped from the south by one that is well exposed on Needle Mountain and is composed mostly of large blocks of Schultze granite. To the northwest they interfinger with and are overlapped by a member, in the southeast part of the area, containing some fragments of Pinal schist together with dacite, diabase, quartzite and Barnes conglomerate of the Apache group, and Paleozoic limestone. The diabase, limestone, and rocks of the Apache found in this last member must have been derived from the northwest, for only in that direction are such rocks exposed. In detritus derived from the northwest one would also expect to find debris of the quartz monzonite and granite porphyry, yet none has been seen. Likewise, quartz monzonite debris is lacking in the Gila conglomerate southwest of the mine. The absence of debris from the quartz monzonite in the Gila near Porphyry Mountain indicates either that this area was part of a basin of deposition and was covered by the conglomerate or that the quartz monzonite was buried under older rocks, perhaps under dacite or Whitetail conglomerate, when the Gila was deposited.

There is evidence to suggest that Porphyry Mountain was covered by Gila conglomerate. The summit of this mountain, before it was destroyed by mining operations, was a nearly flat surface, about 200 by 300 feet in size, that may have been the remnant of an old erosion surface. Today a few blocks of dacite, diabase, quartzite of the Apache group, and Pinal schist are scattered near the few remnants of the old surface on the uppermost bench of the mine.

The irregularity of the surface on which the Gila conglomerate was deposited is well illustrated by section *F–F′* on plate 2. In the churn-drill hole near the south end of the section, the basalt flow interbedded in the Gila formation is underlain by about 700 feet of conglomerate, whereas about 3,500 feet to the northeast it is underlain by only a few feet of conglomerate and to the northwest it rests directly on dacite. Mapping along the unfaulted margins of the Gila conglomerate clearly shows that the deposit has covered an old surface of moderate relief. Each of the two large areas now covered by the Gila, the one west and the other southeast of the mine, was probably part of a large basin. The deposits have been preserved in these two localities partly because they were relatively low areas, where the thickest accumulation of conglomerate was deposited, and partly because each has been relatively depressed by faults of post-Gila age. The lithology of the deposits shows that the debris was washed into these two areas from nearly every side; if the two were joined, the connection was probably across the area where Porphyry Mountain now stands. The original thickness of the Gila in these areas was at least 1,000 feet. Undoubtedly most of the valleys on the old surface were filled and the gravels spread more thinly over most of the higher areas.

Local volcanic activity occurred during Pliocene or Pleistocene time, as indicated by a flow of vesicular olivine basalt in the Gila conglomerate southwest of the mine. Except where scoriaceous crusts have been reddened by oxidation, the rock is dark gray or black. It has an aphanitic or microcrystalline groundmass and contains phenocrysts of both olivine and plagioclase. Much of the olivine is fresh, but some of it is converted to iddingsite. The flow is about 100 feet thick throughout most of its extent. Along its eastern margin it is underlain by only a few feet of conglomerate or rests directly on dacite, but farther southwest the conglomerate beneath it is much thicker. It probably accumulated in a large valley as it poured over the conglomerate and did not originally extend much farther than it does today. One possible source of this flow is a basalt plug on Manitou Hill, 2½ miles south of Porphyry Mountain.

East of the mine, a dike of vesicular olivine basalt is intruded along a fault between diabase and dacite south of Jewel Hill. It is probably a very shallow intrusion of Gila age, but whether or not it was the feeder of a surface flow is unknown.

STRUCTURE

REGIONAL SETTING

The Castle Dome area lies on the boundary between two larger areas of contrasting geology that are clearly shown on Ransome's map of the Globe quadrangle (1903, pl. 1) and on figure 2 of the present report. To the south the rocks consist of Pinal schist into which were intruded several pre-Cambrian granitic masses and the Schultze granite of Tertiary or Cretaceous age. To the north there are no outcrops of Pinal schist. Instead, rocks of the pre-Cambrian Apache group and of Paleozoic age are widespread, and wherever the underlying rocks are exposed, they consist of granitic rocks that Ransome (1903, pp. 73–74) considered to be pre-Cambrian. In the northern area, great volumes of diabase were intruded into both the granitic rocks and the sedimentary rocks of the overlying Apache group and of Paleozoic age.

The exact location and character of the boundary between these two areas is uncertain because parts of it are covered by Tertiary formations and also because both diabase and the two Mesozoic (?) quartz monzonite masses, the one on Porphyry Mountain and the other south of Sleeping Beauty Peak, have been intruded across it. West of Pinal Creek, the boundary crosses the Globe quadrangle in an east-northeasterly direction, passing just south of Sleeping Beauty Peak and Porphyry Mountain. Southwest of the Castle Dome mine, the boundary may be a fault, as shown by section *F–F′* on plate 2, but there it is concealed by Tertiary formations. North of the Tertiary cover in this area, the entire sequence of the Apache group and overlying Paleozoic formations crops out, dipping in a general southerly direction toward the Pinal schist that crops out to the south, and it is reasonable to suppose that, somewhere beneath the Tertiary formations, there is a fault between the schist and the Apache and later sedimentary rocks. If this fault is correctly inferred, it trends in a general northeasterly direction. Its extension northeastward would be expected to pass south of Porphyry Mountain, either through the quartz monzonite or along the northern edge of the Pinal schist, but no such fault has been recognized there. Although this may mean that the structure has not been correctly interpreted, it also suggests that the inferred fault is a structure that predates the quartz monzonite and has been obscured by the quartz monzonite intrusion as well as by the later diabase intrusion. If the fault predated the quartz monzonite, it might explain the locus of intrusions along the north margin of the Pinal schist south of Porphyry Mountain and also south of Sleeping Beauty Peak.

Throughout the northwestern part of the Globe quadrangle, the formations of the Apache group and of Paleozoic age have a gentle to moderate southwesterly dip (Ransome, 1903, pp. 97–100). A few open folds occur, but the regional dip is characteristic. The dominant structural feature is a complex of normal faults. Although there is a great variety of strike and dip among the faults, the general effect is to repeat the southwestward-dipping strata; as a result, all the older rocks are widespread throughout the northwestern part of the quadrangle. So numerous are these faults that Ransome described the structure as "regional brecciation." Most of them have small or moderate displacements, and their impressiveness is the result of their great number rather than the magnitude of individual faults.

GENERAL STRUCTURE OF THE CASTLE DOME AREA

The dominant structure of the Castle Dome area is a horst trending in a north-northwesterly direction through the central part of the area. The quartz monzonite is confined within this horst except for the large block west of the Gold Gulch fault that is separated from the main mass of quartz monzonite by a diabase intrusion. The horst is bounded on the east and west sides by steeply dipping normal fault systems along which the quartz monzonite was brought into fault contact with the sedimentary rocks of the Apache group and of Paleozoic age that probably formed the roof of the intrusion. Diabase was later intruded along these faults and now generally separates the quartz monzonite block from the relatively depressed blocks where the sedimentary formations are exposed. A second and smaller horst trending nearly north within the larger one is bounded on the west by the east branch of the Gold Gulch fault, which dips to the west, and on the east by the eastward-dipping Dome fault system and the group of eastward-dipping faults in the quartz monzonite north of the mine. A transverse fault, occupied along part of its extent by a diabase intrusion, cuts across the main horst near the north edge of the quartz monzonite, and along it the block to the south was relatively elevated. This transverse fault is probably a branch from the main fault system bounding the horst on the east; it joins the main fault about 1,000 feet north of the Continental mine. Faults of Tertiary age have also developed both east and west of the quartz monzonite, and along them the old horst structure has been renewed.

North of Porphyry Mountain the Scanlan conglomerate and the Pioneer formation overlying the quartz monzonite dip very gently toward the southwest, whereas the sedimentary rocks west of the quartz monzonite have much steeper southwesterly dips. Along the northwest margin of the quartz monzonite, the Scanlan conglomerate and the Pioneer formation are vertical or dip very steeply and, in a few places, are slightly overturned; along the west side of Gold Gulch, the dip of the Paleozoic limestones is generally between 45° and 50°. The steeper dips in this area probably result from tilting or doming caused by the intrusion of the quartz monzonite or by the diabase intrusion, as it seems unlikely that vertical dips could be caused by simple tilting of blocks along normal faults. In the extreme western part of the map area, the dip of the limestones is toward the south rather than the southwest. East of the quartz monzonite the Apache group and Paleozoic formations form a syncline whose axis trends east-northeastward (pl. 2, section *H–H′*). The south flank of the syncline is on Jewel Hill, and its center is buried by Whitetail conglomerate. Only a narrow segment of the fold is exposed between the two faults east and west of Jewel Hill, and its origin is doubtful, but the fact that its axis trends almost at right angles to the regional strike of the strata suggests that it was not produced by the same stress that caused the regional tilt of beds elsewhere.

The structural lines in the Pinal schist south of the mine and elsewhere beyond the map area trend in a northeasterly rather than a northwesterly direction as in the younger rocks. This is shown by the schistosity, which generally strikes northeast and dips very steeply either toward the northwest or the southeast, and by the alinement of granite bodies, older than the rocks of the Apache group, that occur in a northeastward-trending belt through the center of the schist area.

In general, the Tertiary formations have low dips. West and northwest of the quartz monzonite area, they dip toward the southwest like the older rocks, but more gently. The Gila conglomerate southwest of the mine has been warped into a shallow syncline the axis of which strikes northwest, as shown by the basalt flow interbedded in the formation in that area. East of the mine, where it lies on the dacite, the Gila conglomerate dips toward the south, but near the southeast edge of the map area it is subhorizontal. Whatever tilting or warping has affected the Tertiary formations appears to have been caused by displacements along Tertiary faults, and steep dips have been observed only locally along these faults. Where the formations have been tilted, they dip to the southwest more commonly than in any other direction.

PERIODS OF FAULTING

There have been at least four distinct periods of major faulting in the Castle Dome area. The oldest recognizable faults are post-Carboniferous. They were formed after the quartz monzonite intrusion but before the diabase intrusion. The next younger faults are contemporaneous with, and apparently were caused by, the diabase intrusion. The two youngest sets of faults are Tertiary in age; one set was formed after the dacite but before the deposition of the Gila conglomerate, whereas the other set followed the deposition of the Gila. A thrust fault that is younger than the Whitetail conglomerate but older than the dacite crops out near Pinto Creek west of the map area (p. 41). It is likely that faulting also occurred after the intrusion of the diabase and before the deposition of the Whitetail conglomerate, although no faults of that age have been recognized in the field.

Some of the largest faults were formed before the diabase intrusions. These faults are fundamental structures; they played an important part in controlling the diabase intrusions, and their pattern is partly reflected by the younger faults. Diabase intrusions and later displacements occurred along many of these older faults and masked them in numerous places. Many more faults are undoubtedly buried by Tertiary deposits. Nevertheless, the general pattern of the faults formed prior to the diabase intrusion is clear and has been illustrated on the sketch map (fig. 4). On the geologic map (pl. 1), many of the oldest faults are not readily recognizable; where diabase was intruded along them, intrusive contacts rather than faults are shown on the map unless some movement following the intrusion of the diabase has occurred along the same line. Actually there probably has been a little later movement along most of the old faults, but the later displacements are not indicated where they seem structurally insignificant.

It is clear that many faults are older than the diabase. For example, in two inclusions of sedimentary rocks in diabase southwest of the mine (pl. 1), faults between the Martin limestone and Pioneer formation obviously predate the diabase, as they are confined to the inclusions and do not extend into the younger diabase. Likewise, the fault between the Martin limestone and Pioneer formation just west of Gold Gulch (crossed by section line *E-E′* of pl. 1) cannot be traced either north or south into the diabase. The normal fault near the north edge of the quartz monzonite, 1,700 feet north of Continental Spring, has dropped the Pioneer formation and the underlying quartz monzonite on the north side with respect to the main mass of quartz monzonite on the south, and for more than 1,500 feet, a narrow diabase dike has

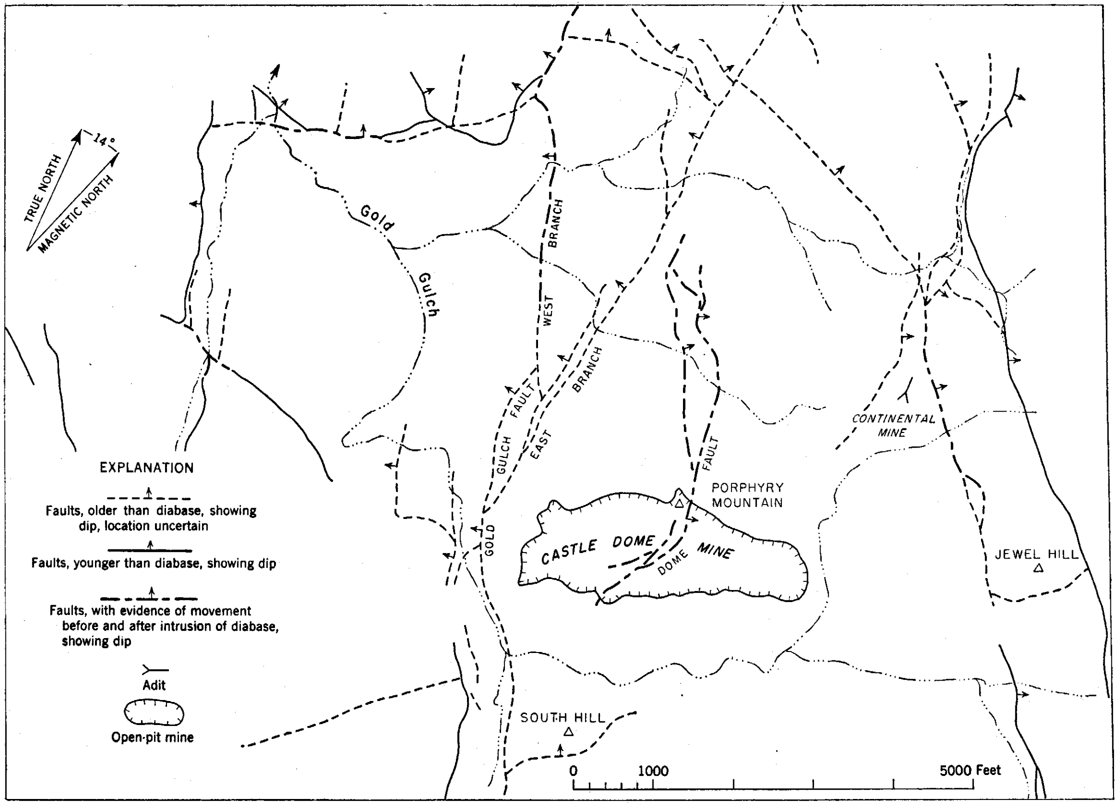


Figure 4.—Age interpretation of major faults, Castle Dome area, Ariz.

followed the fault. In many places, the brecciated quartz monzonite and the original fault surface can be seen along the south edge of the dike, but the diabase itself is not brecciated and here and there chilled selvages are exposed along the edges of the dike.

Another fault of large displacement that was formed prior to the diabase intrusion extended east-northeastward through the northwest corner of the map area. Today it is masked both by the later diabase intrusion and the Tertiary formations and by later Tertiary faults that limit the Whitetail conglomerate on the south and follow approximately the line of the old fault. The old fault cut directly across the strike of the Paleozoic rocks and the formations of the Apache group and displaced them so that the Paleozoic limestones on the north side were in contact with the Pioneer formation on the south. Movement along this fault prior to the diabase intrusion is clear only in one place where the diabase intrusion has left a remnant of the old fault, showing Barnes conglomerate and the Pioneer formation in fault contact with Naco limestone. Apparently the diabase, in the form of a thick sill, intruded the Pioneer formation south of the fault and stopped on the north against the limestone along the old fault.

North of the quartz monzonite, the diabase is in undisturbed intrusive contact with the Pioneer formation in many places along regular surfaces of dislocation that cut across the bedding of the sedimentary rocks. These surfaces were undoubtedly faults in the Pioneer formation that have been followed by the diabase intrusion.

Both the Gold Gulch fault system and the fault that bounded the main horst and the quartz monzonite on the east side are doubtless older than the diabase although the available evidence does not lead directly to this conclusion. Large, irregular diabase intrusions have followed approximately along both of them, and both have had later movement, but the later faulting seems inadequate to account for all the structural features on the two sides of the horst. Along Gold Gulch, west of the mine, the contacts of the diabase are irregular and little, if at all, brecciated, and no continuous fault can be followed either through the diabase or along its contacts. Nevertheless, a major fault in this area has dropped the Paleozoic limestone alongside quartz monzonite. The interpretation given is that this major fault predates the diabase and has been followed by the later diabase intrusion. Where the Gold Gulch fault cuts through the quartz monzonite, at mine coordinate position W. 4050, I (pl. 1), a small, relatively unbrecciated diabase dike has been intruded into the fault zone. Likewise, along the edge of the quartz monzonite north of the Continental mine, the areal relations of the formations can best be explained by assuming that most of the faulting that has brought the quartz monzonite and Paleozoic limestones into proximity occurred before the diabase was intruded between them.

Later displacements along the early faults have been recognized in many places, particularly where Tertiary formations have been displaced, as in the northwestern part of the map area and northwest of Jewel Hill, but in many other places, even where diabase has been intruded along an old fault, it is doubtful whether later displacements have occurred. Commonly the diabase contacts are poorly exposed. The west branch of the Gold Gulch fault is one that has been mapped as a fault younger than the diabase although it was undoubtedly an earlier fault that guided the diabase intrusion into the quartz monzonite. The later displacement is indicated because, where diabase in contact with quartz monzonite is exposed, it is both brecciated and silicified and also because the fault seems to connect with a fault farther north that is younger than the Whitetail conglomerate. However, along the same fault but farther south, where no diabase occurs, movement that occurred after the diabase intrusion cannot be distinguished from the earlier dislocation. The contact between limestone and diabase along the west side of Jewel Hill is the most difficult of all to interpret. It is well exposed in a few places, and nothing indicating much movement after the diabase intrusion has been seen along it, but along the projection of this contact both to the north and to the south, there are Tertiary faults that one might expect to connect along the west side of Jewel Hill.

The age of faults that are entirely within the quartz monzonite or older rocks is uncertain. In this class are those north of the Castle Dome mine. Where the small diabase sills are numerous, as they are in the mine, the evidence shows clearly that displacements took place both before and after the sills were intruded, but it is not known whether these sills are of the same age as the main diabase intrusion around the quartz monzonite. The interpretation given is that the larger faults within the mine, notably the Dome fault, have had movement both prior to and after the diabase intrusion, and this interpretation is also applied to the faults in the quartz monzonite north of the mine.

Faults that were probably caused by the diabase intrusion occur in the Paleozoic limestones along the west side of Gold Gulch west of the mine. None of these faults can be traced into the large body of diabase underlying the Martin limestone. The largest of them has an apparent vertical component of movement of at least 700 feet, but most of them had displacements of less than 100 feet. The faults differ greatly in both strike and dip and include both normal and reverse types, but most of them cross the strike of the limestone strata at large angles. Nothing clearly distinguishes these faults from those older than the diabase, although several lines of reasoning suggest that they actually were caused by the diabase intrusion. Including, as they do, both normal and reverse faults in a great variety of attitudes, they do not seem to fit into what is known of the fault pattern that preceded the diabase intrusion. Furthermore, they occur in the rocks overlying a large diabase mass, and one might expect such a great volume of magma intruded at a relatively shallow depth beneath massive and competent rocks to break and jostle the overlying formations. Probably many other faults caused by the intrusion are present in the Castle Dome area but have not been recognized as distinct from earlier faults formed prior to the intrusion. Ransome (1903, pp. 104–105) recognized that many of the early faults in the Globe quadrangle were caused by the diabase intrusion, but he was not sure how much of the early faulting was caused by stresses that preceded and were independent of the diabase intrusion.

Important faulting occurred during the Tertiary period, but faults of this age can be recognized only where they displace one of the Tertiary formations. Therefore, the known Tertiary faults are all near the east, west, and northwest edges of the map area where the Tertiary formations crop out. Other faults that displace the diabase may also be, and are generally assumed to be, Tertiary. Some of the Tertiary faults are younger than the dacite but older than the Gila conglomerate, whereas others are younger than the Gila. Faults belonging to each of these periods have been definitely recognized, but where the Gila conglomerate is absent, as northwest of Continental Spring, the Tertiary faults cannot be subdivided.

The only fault in the mapped area that is certainly younger than the dacite and older than the Gila conglomerate forms the west boundary of the dacite north and east of Jewel Hill. South of Jewel Hill, this fault has been intruded by a dike of vesicular olivine basalt and is overlapped by the Gila conglomerate. The displacement along it cannot be determined, but it was at least 500 feet and probably much more. Ransome (1903, p. 104) recognized this period of faulting as one of the most important in the Globe-Miami region, as indeed it must have been to account for the major unconformity between the dacite and the Gila conglomerate.

The only large fault in the Castle Dome area that is known to have followed the deposition of the Gila is southwest of the mine where Gila conglomerate and basalt have been brought into contact with Pinal schist and diabase along a normal fault that extends for several miles south of the map area and for an unknown distance northward.

This fault forms the east boundary of the Tertiary rocks west and southwest of the mine and is the probable explanation for the lower elevation and westward tilt of these formations in that part of the area. Faulting that occurred after the Gila conglomerate was deposited was not recognized by Ransome as very important. Although large faults younger than the Gila may not be numerous, several are known, and this period of faulting is certainly more significant than Ransome assumed. The Miami fault, which Ransome later (1919, pp. 112–113) recognized, bounds the Gila conglomerate west of Pinal Creek. It has a displacement of at least 1,000 feet and extends for at least 6 miles north of Miami. Farther west, two other faults cutting Gila conglomerate are known to have displacements between 500 and 1,000 feet (fig. 2). When the Gila conglomerate has been mapped over a much larger area than has yet been studied, it will probably be apparent that later faulting has been very important in shaping the present topography and the present distribution of the Gila conglomerate.

STRUCTURES IN THE CASTLE DOME MINE

The most prominent single structure within the Castle Dome mine is the Dome fault, which passes through the summit of Porphyry Mountain and nearly bisects the mine. It is a branching fault system that diverges in the southern part of the mine along a west branch, striking about N. 55° E. and dipping 60° SE., and along another branch farther east that strikes N. 30°–35° E. and dips 55° SE. Neither of these branches can be traced southward beyond the mine limit, and although both might be expected to cut adit 2 (pl. 3), neither can be recognized in it. Northward the east branch of the fault, striking N. 15°–20° E. and dipping 65° E., is dominant and can be traced to a point where it disappears beneath the dump. It may connect with the fault in the quartz monzonite northwest of the Continental mine or with the eastward-dipping faults north of Porphyry Mountain, all of which disappear southward under the dump.

The Dome fault clearly predates the period of mineralization. The east branch of the fault is occupied by a shoot of massive sphalerite, about 5 feet wide, that is exposed on the 4,590-foot level. That it is also older than the pyrite-chalcopyrite mineralization is certain because, in the southern part of the mine, relatively intense sericitization occurs along it, and to the north, beyond the dump, the faults that probably connect with it have controlled the distribution of the most intense mineralization in that area. The earliest movement along the Dome fault occurred even before the diabase sills were intruded. Although the sills east and west of the fault appear to be parts of the same sill offset by the fault, several steeply dipping dikes of diabase with no counterparts either east or west of the fault have been intruded into the fault zone. The offset of the most extensive sills on the two sides of the fault zone may be apparent only, or it may be due to an offset of some older fracture that was later intruded by the diabase. This doubtful offset affords the only suggestion of the amount or direction of displacement along the Dome fault. Judging by it, the displacement has been normal, amounting to about 300 feet. Some movement following mineralization has also occurred along the fault. In a few exposures on the lower levels of the mine, the fault gouge contains fragments of brecciated vein filling. However, the displacements following mineralization have not been large, for the massive sphalerite shoot exposed along the east branch of the fault has not been brecciated.

Many small faults, in addition to the Dome fault, occur in the mine area and are indicated on plate 3. Most of them strike between N. 45° W. and N. 65° W. and dip moderately to steeply toward the northeast. Although several have been followed for short distances by the diabase sills, later movements along many of them have offset the sills. Many also predate mineralization and have controlled intense local alteration of the wall rock. West of the Dome fault, the minor faults strike in a more northerly direction than those farther east, and they dip more gently to the east. Along several of them the veins are slightly offset. Some small faults, occurring east of the Dome fault and mostly on the lower levels of the mine, strike east-northeast parallel to the main set of veins and, like the veins, dip steeply to the south. With a few exceptions, all the faults in the mine dip in easterly directions, and wherever the direction of displacement is indicated, the faults appear to be normal, the eastern block having been displaced relatively downward.

Joints in the quartz monzonite exposed in the mine fall into two general sets. The more prominent and uniformly developed set strikes east-northeast, dips steeply to the south, and is now represented by closely spaced, generally parallel, thin quartz-pyrite veins. The second and less prominent set of joints is generally unmineralized. It strikes north to northwest and dips moderately to the northeast. The attitude of this set differs a little on the two sides of the Dome fault. Along some of the bench faces the rock is sheeted by closely spaced joints belonging to this second set, whereas along other faces the set is obscure. Generally these joints are more numerous on the lower, or southern, benches of the mine and east of the Dome fault.

The attitudes of about 1,300 veins and 2,000 joints were recorded from four levels, at elevations of 4,350, 4,390, 4,430, and 4,470 feet, respectively, in the part of the mine between coordinate lines W. 0 E. and W. 3000 and between lines *G* and *L*, an area about 3,000 feet long by 500 feet wide. The readings were made in many unit areas, each comprising 100 feet of the exposed face, scattered as uniformly as possible throughout the total area studied. Within each unit area an attempt was made to record every vein and joint that could be reached and was well enough exposed to permit reliable determination of its strike and dip. Probably the readings in some unit areas are not as reliable statistically as those in others, either because fewer readings could be made or because the attitude of the face was more favorable for the exposure of one set of veins or joints than another. Probably some fractures produced by blasting have been included among the joints, although an attempt was made to exclude these. However, the recordings made within the individual unit areas all combine to show a generally uniform fracture pattern that is somewhat different on the two sides of the Dome fault. This is illustrated by two pairs of diagrams (fig. 5), one pair showing the unmineralized and mineralized fracture sets east of the fault and the other pair showing the same sets on the west side.

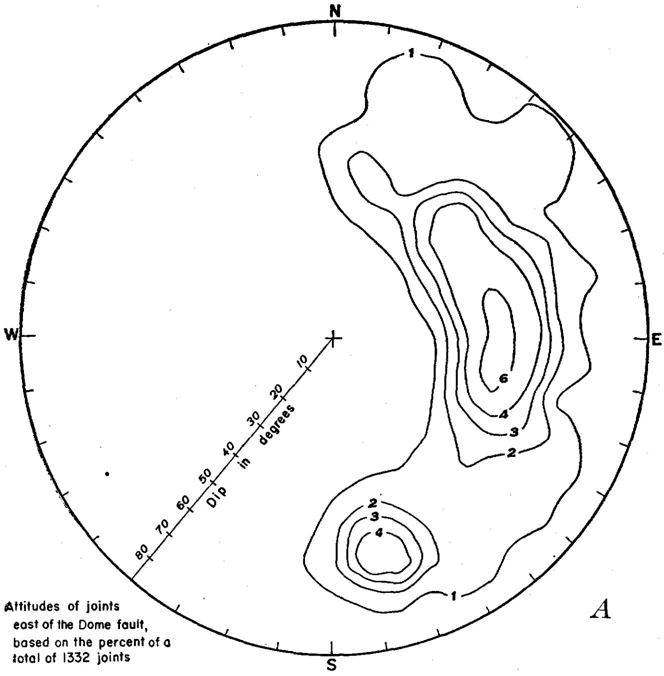
These vein and joint diagrams were compiled by the method generally used in structural petrology (petrofabric analysis) to show with statistical accuracy on a single diagram both the strike and the dip of all fractures. The method has been described by several authors, but in a very convenient form by Billings (1942, pp. 115–122). The poles of perpendiculars to the fractures are plotted on an equal-area projection net, resulting in a "point diagram." The concentration of points in various parts of the point diagram is determined by counting the number of points within uniformly spaced unit areas, each equaling 1 percent of the total area of the diagram, that cover every part of it. The number of points in each 1-percent area is expressed as a percentage of the total number plotted, and these percentages are then contoured to make the final diagrams. The diagrams, therefore, indicate the relative abundance of fractures with various attitudes rather than the exact number or percentage of fractures with a particular attitude. For example, in the vein diagram of figure 5*B* about a third of all the recorded veins are within the 10-percent contour—that is, about a third of all the veins strike between N. 65° E. and N. 85° E. and dip between 50° and 75° S. Thus the diagrams are chiefly useful for indicating the average strike and dip of each set of fractures and the variation of strike and dip within each set.

The diagrams here given show that east of the Dome fault the large majority of veins belong to a set that has an average strike of

N. 75° E. and an average dip of 60° S., but there is also a very minor set of veins that strikes about N. 30° W. and dips about 45° NE. The unmineralized joints east of the fault likewise fall into two sets. The lesser of the two corresponds exactly to the dominant set of veins and probably represents the original vein fractures that were not mineralized or later fractures that have opened parallel to the veins. The dominant set of joints ranges in strike from N. 20° E. to N. 40° W. and dips about 45° E. This set includes joints that have the same attitude as the minor set of veins, but it differs in the inclusion of many joints that strike more nearly north than the veins. Since this statistical study was completed, several new levels have been developed, and on these the dominant unmineralized set of joints east of the Dome fault is more prominent than on the higher levels where the readings were made.

West of the Dome fault the veins fall into a single very pronounced set with an average strike of N. 75° E. and an average dip of 72° S., but the set includes veins with a considerable variation in strike and dip. There is also a faint concentration (shown near the center of the diagram) of veins that in general dip gently to the southeast. Unmineralized joints west of the Dome fault fall into one set that corresponds to the prominent set of veins and another set that averages about N. 60° W. in strike and about 30° N. in dip. This second set is distinct from any other set of veins or joints either east or west of the Dome fault, but it includes fractures the strike of which varies greatly. A third, very minor set of joints on the west side strikes N. 10°–30° W. and dips very steeply to the southwest. Clearly shown by all the diagrams and likewise apparent in the field is the scarcity of veins and joints that dip toward the west.

The general fracture system apparent in the mine is widespread and has been recognized in all the massive rocks in the surrounding region. North of the mine, the same system can be seen in the quartz monzonite, and throughout the mineralized area, veins are common only along fractures striking east-northeast and dipping steeply to the south. This same pattern of veins occurs in both the granite porphyry and the diabase east of the quartz monzonite. In the granodiorite south of the mine, the dominant set of fractures strikes east-northeast to east and dips steeply to the south; although no true veins are present in the granodiorite, some of these fractures contain films of sericite. Near the southeast corner of the map area, thin veinlets in the quartz monzonite also have this attitude, and throughout the Schultze granite exposed south of the map area, a similar vein and joint system is well developed. The dominant set of fractures in the Schultze granite has an attitude similar to that of the main set of veins in the mine,



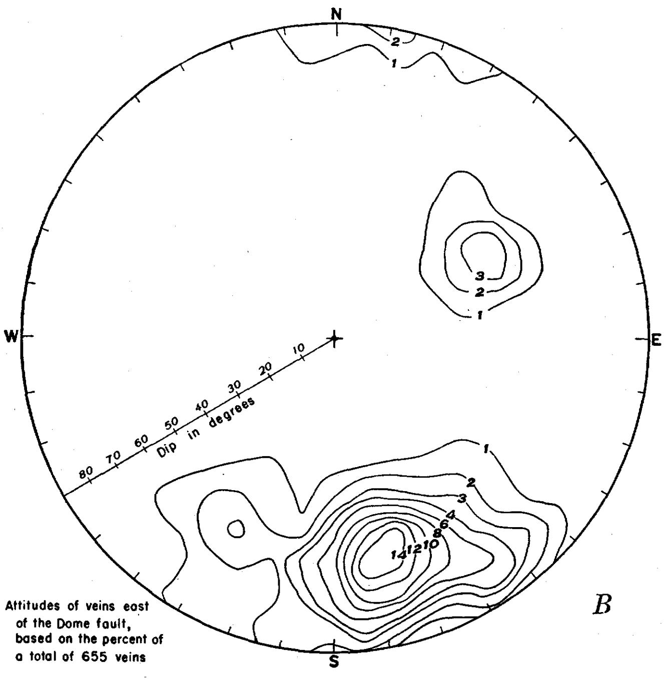


FIGURE 5.—Stereographic diagrams illustrating orientation of vein and joint systems in Castle Dome mine, Gila County, Ariz.

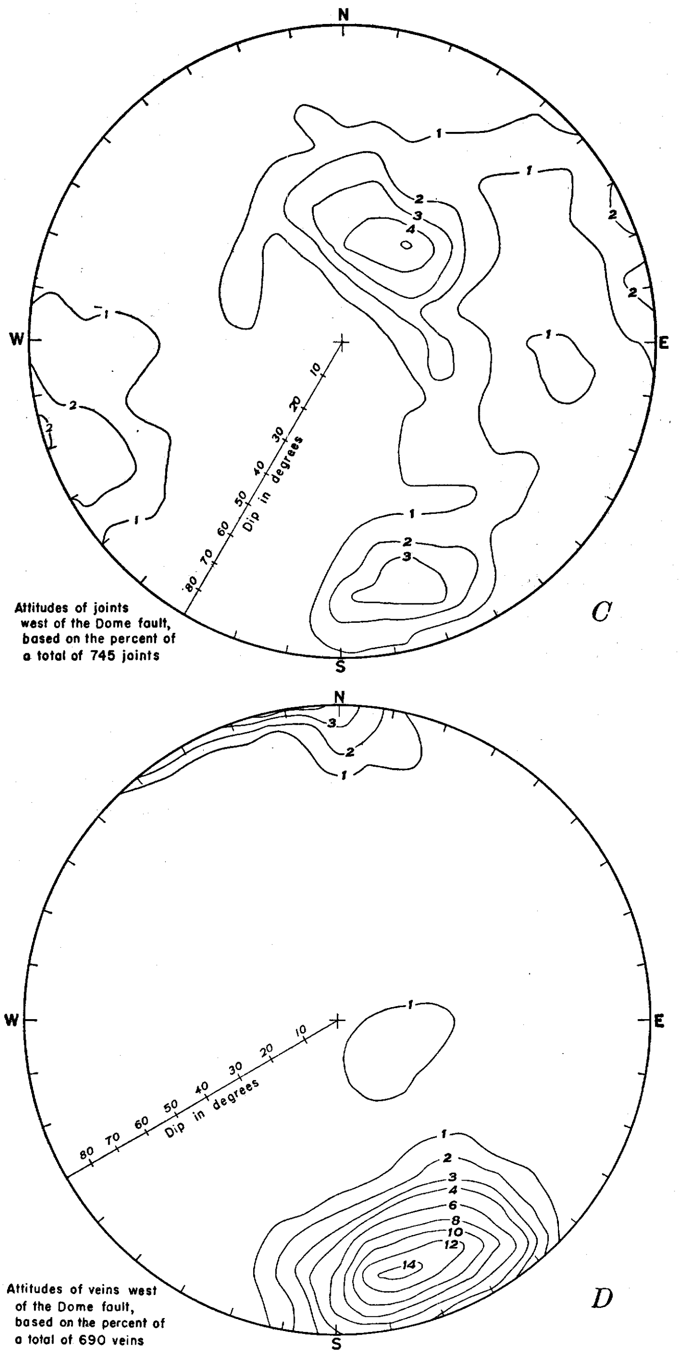


FIGURE 5—Continued.

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ranging in strike from east to northeast and dipping moderately to steeply toward the south. Many of these fractures can be followed continuously for 40 or 50 feet in some large outcrops, and nearly every one is occupied by a thin, continuous veinlet of cross-oriented muscovite with or without quartz. A second, less uniform set of barren shear joints in the granite crosses the strike of the veins at a large angle, trending generally northward to north-northwestward and dipping steeply to the east. Both the barren shear joints and the muscovite veins cross pegmatites, aplites, and large barren quartz veins, as well as each other, without noticeable offset. Indeed, some have bisected large orthoclase phenocrysts in the granite without displacing the two halves.

The vein fractures are probably later than the diabase sills in the mine, although generally the sills are thoroughly brecciated instead of being systematically fractured like the quartz monzonite into which they were injected. Mineralization of the brecciated sills has produced numerous intersecting veinlets with no definite patterns, but in a few places, veins belonging to the main set in the quartz monzonite continue into the diabase sills. Furthermore, none of the sills was injected along fractures paralleling the veins, and had such pronounced fractures existed before the diabase injections, some of them would surely be occupied by diabase dikes.

The vein fractures in the Castle Dome area were probably developed only a little before mineralization, for they are developed in the granite porphyry, which is the last intrusion known to predate mineralization. If, as is likely, the granite porphyry intrusion was essentially contemporaneous with the intrusion of the Schultze granite, the set of fractures in the granite, now occupied by muscovite veinlets, is probably of the same age as the similarly oriented set of vein fractures in the Castle Dome mine. They cut, not only the Schultze granite, but also the pegmatites, aplites, and barren quartz veins that are late phases of the intrusion of the Schultze granite.

Where the vein fractures cut older structures, they cause no offset and must be regarded as a regional set of joints. Perhaps displacements occurred along some of the original fractures, but the only known faults parallel to the vein fractures were formed after the period of mineralization and have been recognized in the mine east of the Dome fault. Movement along them has been small, but they are notable features in the mine, for in many places the bench faces break along them. They are slickensided and contain as much as half an inch of black gouge, which commonly contains crushed pyrite. They are probably related to movements that occurred after the period of mineralization along the Dome fault, from which some of them appear to branch.

The set of barren joints in the mine, striking generally north-northwest and dipping to the east, may be either contemporaneous with, or younger than, the vein fractures and the mineralization. As shown by the diagram (fig. 5*B*), a minor set of veins trends north-northwestward and dips 45° E., indicating that some fractures that predate mineralization have such a trend. This minor set actually represents very few veins, and if it is compared with the set of barren joints whose attitude is generally similar (fig. 5*A*), a considerable difference in the form of the contours that outline the sets on the two diagrams is clear. Although they dip at about the same angle, the strike of the veins differs much less than the strike of the barren joints. In the field no conclusive evidence has been found to determine their relative ages. The fact that most of the fractures in this set are barren suggests that most of them were formed after the period of mineralization.

The difference between the fracture sets on the two sides of the Dome fault suggests that the joints are in some manner related to the fault. The attitude of the main set of veins only is the same on opposite sides of the fault; however, the actual relation between the Dome fault and the joints is not indicated. At present no satisfactory explanation of the vein and joint sets in the Castle Dome area is evident, and in seeking an explanation it must be emphasized that the fracture system described is clearly regional in character and that any explanation of it must be applicable to a large region and not to the Castle Dome area alone. In view of the regional development of the fracture system, it seems unwise, where conclusive evidence is lacking, to attempt an explanation until the extent of the fracturing and the details of the regional geology are known.

The diabase sills in the mine have a variety of attitudes, but all of them, except those in the Dome fault zone, dip gently to the north, northwest, or west. Most of those shown on the mine sections (pl. 4) have not yet been exposed by mining operations and are known only from exploratory-drilling records. Their broad extent and their continuity in the quartz monzonite suggest that they were injected along several continuous fractures, probably faults. However, where they have been exposed in the bench faces, the quartz monzonite along their margins is not noticeably more brecciated than elsewhere, and no set of fractures parallel to the sills has been recorded.

CASTLE DOME COPPER DEPOSIT

HISTORY OF MINING

The history of mining in the Miami district, which includes the Castle Dome area, began in 1874 when the Globe claim, now a part of what is generally known as the Old Dominion mine, was located by the same party of prospectors that located the Silver King mine north of Superior. During the years that followed, interest shifted to other locations in the Globe area where rich silver ores were found. Although the presence of widespread copper mineralization was recognized, the deposits received little serious attention until about 1881, when a small copper furnace was erected on the Western Pass road about 6 miles nearly due west of Globe.

In 1901, when F. L. Ransome, of the U. S. Geological Survey, made his first study of the Globe-Miami area, some chrysocolla ore was being shipped from the Keystone mine. At this time 4-percent copper ore was considered to be the lowest grade that could be profitably worked. In 1904 disseminated chalcocite was discovered in what afterwards proved to be the east end of the Live Oak ore body, and 2 years later a small mill was built to treat the low-grade ore. The venture was not successful, but it attracted considerable attention to the possibility of exploiting this type of low-grade copper deposit by means of large-scale operations. There followed a period of intensive exploration by shafts, adits, and drill holes that revealed great reserves of disseminated chalcocite containing approximately 2 percent copper. In 1911, the production of copper concentrate was begun by the Miami Copper Co., and in 1915, the Inspiration Consolidated Copper Co. began operations on a large scale, using a flotation process to recover the copper sulfides.

The first mining location in the Castle Dome area was the Continental claim, filed about 1881. Other locations followed on the south flank of Jewel Hill and along Gold Gulch, where terrace gravels cemented by copper carbonates were found. The Continental mine attracted little attention until 1896, after considerable development work had been done. In 1899, the mine was purchased by the Old Dominion Co. Although a little rich copper ore had been found, none had been produced up to the time of Ransome's visit to the mine in 1902. The mine was operated intermittently from 1906 to 1941, mainly by leasers.

From 1905 to 1910, many claims were located on Porphyry Mountain and throughout the Cactus mineralized area on Pinto Creek, 2 miles south of Porphyry Mountain. During this period most of the claims were acquired by the Arizona National Copper Co., the Pinto Copper Mining Co., and the Cactus Development Co. The Arizona National Copper Co. obtained a controlling interest in the Cactus Development Co. and carried on some development work.

The Cactus Development Co. was later reorganized as the Cactus Copper Co. by the firm of Gay & Sturgis, of Boston. The new company leased the property of the Arizona National Copper Co. and, between 1908 and 1910, sank 2 shafts and 15 churn-drill holes to explore the Cactus property on Pinto Creek.

The Castle Dome Development Co. was incorporated December 22, 1915, under the laws of Maine, with a capitalization of $3,000,000. At the time of incorporation its holdings consisted of eight patented claims on Porphyry Mountain and a 97-percent interest in the Inspiration Extension Copper Co. The latter company owned 29 patented claims on Porphyry Mountain and 95 claims adjoining the west side of the Inspiration Consolidated Copper Co.'s property in Webster Gulch. The Castle Dome Development Co. drove seven adits, totaling 3,500 feet, and during 1918 and 1919 put down seven churn-drill holes on what is now the site of the Castle Dome mine on Porphyry Mountain.

In 1920 the Susquehanna Trust and Safe Deposit Co. obtained title by foreclosure to the property of the Arizona National Copper Co., which had been mortgaged to them for a loan of $250,000.

The Pinto Valley Co. was incorporated April 7, 1921, with a capitalization of $500,000. Garret Mott became president and T. R. Drummond manager. This company leased and optioned the property of the Arizona National Copper Co. from the Susquehanna Trust and Safe Deposit Co. for a period of 3 years. Under the terms of the option the assets of the Arizona National Copper Co. were to be acquired in exchange for 110,000 shares of Pinto Valley stock whenever Mott or his associates paid into the treasury of the Pinto Valley Co. $100,000 for a like amount of stock. The option was eventually exercised, and the Pinto Valley Co. acquired the assets of the Arizona National Copper Co., including the Pinto Valley Copper Co. and the Cactus Copper Co., which it controlled.

In February 1924, the Pinto Valley Co. acquired the assets of the Castle Dome Development Co. by means of a merger, the terms of which called for the exchange of 600,000 shares of the latter's stock for a like amount of Pinto Valley stock and the payment of $15,000 in cash. In order to accomplish this merger, the capitalization of the Pinto Valley Co. was increased from $500,000 to $4,000,000. Thus the Pinto Valley Co. obtained title or control of 55 patented and 45 unpatented claims on Porphyry Mountain and Pinto Creek. The 37 claims on Porphyry Mountain became known as the Castle Dome division, and the original property of the Pinto Valley Co. on Pinto Creek was called the Cactus division. Four churn-drill holes were put down on the Castle Dome division in 1924.

During the ensuing years the Cactus division attracted greater interest than the Castle Dome division. However, sufficient ore was indicated at Castle Dome by the 11 holes that had been drilled up to this time to warrant the retention of a metallurgist to conduct tests with a view to the possibility of the economic extraction of copper from the Castle Dome ore by a leaching process. During 1926, metallurgical investigations were conducted in a small pilot plant on the property. The ore used in the tests was obtained from the Indicator adit on the south slope of Porphyry Mountain.

Two more churn-drill holes were driven in 1926 and 1928 and a diamond-drill hole, 1,085 feet deep, in 1929. Also in 1929, a contract was made with the Inspiration Consolidated Copper Co. to treat 3,000 tons per day of the Castle Dome ore over a period of 8 years. The estimated reserves at this time consisted of 9,000,000 tons of ore containing 1.27 percent copper and 20,000,000 tons containing 0.8 percent copper. The proposed plan of operation called for the development of an underground mine employing a system of block caving similar to that in use in the Miami Copper Co. mine. The Pinto Valley Co. was unable to finance the venture and suffered severe financial difficulties during the ensuing years of business depression.

In May 1940, the Miami Copper Co. purchased those properties of the Old Dominion Co. that were located in the Miami district and thus acquired a portion of the Continental group of mining claims that covered the east end of the Castle Dome ore body. In the fall of 1940, the Miami Copper Co. took an option on the adjoining Castle Dome group of claims from the Pinto Valley Co. A drilling campaign was begun in July of the same year for the purpose of exploring the ore body. By the end of 1941, sixty-one drill holes had been completed, and the results were so favorable that the Miami Copper Co. exercised its option. The Castle Dome Copper Co., Inc., a wholly owned subsidiary of the Miami Copper Co., was organized in November 1941 to hold and operate the claims.

Also in the latter part of 1941, arrangements were made with the Defense Plant Corporation, a subsidiary of the Reconstruction Finance Corporation, for developing and equipping the mine and for the construction of a concentrating plant as a national defense project. Construction and development work was begun in January 1942, and milling operations started June 10, 1943. Since that date the mine has been in continuous operation by the Castle Dome Copper Co., Inc., under a lease agreement with the Defense Plant Corporation.

PRODUCTION

In the early days a few hundred tons of cemented terrace gravels containing copper carbonates was mined from deposits along the sides of Gold Gulch; also, a little gold is said to have been washed from the gravels in Gold Gulch; but until the development of the Castle Dome open-pit mine, the only recorded production in the Castle Dome area came from the Continental mine, 3,000 feet northeast of Porphyry Mountain. From 1906 to 1941, the Continental mine produced 34,000 tons of ore that contained approximately 3,600 ounces of gold, 134,000 ounces of silver, and 2,000,000 pounds of copper.

The production of copper concentrate from the Castle Dome mine was begun June 10, 1943. A summary of copper production to December 31, 1946, is shown below. During this period 5,842 ounces of gold and 295,965 ounces of silver were recovered from the copper concentrate.

Copper production of Castle Dome mine

Year	Ore (tons)	Copper (pounds)
1943	1, 714, 205	18, 020, 066
1944	4, 107, 975	49, 743, 367
1945	4, 183, 769	53, 324, 969
1946	4, 102, 566	56, 590, 107
Total	14, 108, 515	177, 678, 509

Although no record of the production of gem turquoise is available, this product of the mine should not go unmentioned. Undoubtedly the Castle Dome mine is one of the most important sources of gem turquoise at the present time. The quantity marketed during the first 3 years of operation was probably between 10,000 and 20,000 pounds.

The Castle Dome ore contains 0.01 to 0.02 percent MoS_2; as yet, however, no attempt has been made to recover molybdenum, mainly because an adequate supply of fresh water for use in a recovery unit is not available.

MINERALS OF THE CASTLE DOME COPPER DEPOSIT

Here follows a tabulation of the minerals present in the Castle Dome copper deposit grouped according to their mode of origin.

Minerals of the quartz monzonite host rock:

Quartz.	Biotite.	Zircon.
Orthoclase.	Apatite.	Ilmenite.
Oligoclase.	Sphene.	Magnetite.

Minerals formed by hydrothermal alteration of the host rock:

Quartz.	Allophane.	Epidote.
Sericite.	Bleached biotite.	Clinozoisite.
Pyrite.	Leucoxene.	Calcite.
Beidellite.	Rutile.	Stilbite.[1]
Hydrous mica.	Chlorite.	

Hypogene vein minerals:

Quartz.	Molybdenite.	Specularite.[1]
Sericite.	Sphalerite.	Barite.
Adularia.	Galena.	Wavellite.[1]
Pyrite.	Gold.[2]	Metatorbernite.[1]
Chalcopyrite.	Silver.[2]	Fluorite.[1]

Supergene minerals:

Chalcocite.	Goslarite.	Cimolite.[1]
Covellite.	Anglesite.	Kaolinite.
Cuprite.	Sulfur (native).	Canbyite.
Malachite.	Limonite.	Quartz.
Azurite.	Jarosite.	Chalcedony.
Copper (native).	Turquoise.	Opal.
Chalcanthite.	Halloysite.	Gypsum.[1]
Molybdite.	Endellite.	Silver (native).

Minerals of uncertain origin:

Libethenite.[1]	Wulfenite.[1]

[1] Noted in one or more places but not general in distribution.
[2] Indicated by analysis.

On page 130 is a general list, arranged alphabetically, of all the minerals of the Castle Dome deposit.

HYPOGENE MINERALIZATION

GENERAL DESCRIPTION AND TENOR OF THE PROTORE

The hypogene mineralization in the Castle Dome deposit is largely confined to the quartz monzonite and the granite porphyry bodies. Where it extends into the adjacent rocks, it is generally very weak. The effects of supergene alteration are so superficial that the unaltered protore can be observed in many places in the mine, particularly in the footwall of the ore body. The hydrothermal alteration that accompanied the mineralization has not affected the appearance of the host rock as greatly as in most other disseminated copper deposits, and in most of the mineralized area the texture of the quartz monzonite is perfectly preserved. The only readily apparent difference in appearance between protore and unmineralized rock, in many places, is a slight bleaching caused by the chalklike character of the altered plagioclase and local bleaching of biotite.

Pyrite and chalcopyrite are the most abundant sulfide minerals. The ratio in which the two sulfides are present varies from place to place in accordance with a pattern of hypogene zoning. They occur in, or associated with, a set of narrow, closely spaced, generally parallel veins in which quartz is the most prominent mineral. Many of the veins contain only quartz, others contain quartz and pyrite, and some are largely pyrite with little or no quartz. Veins containing pyrite are generally bordered by coarse sericite. In the typical protore, containing a maximum of about 0.3 percent copper, the amount of chalcopyrite present is necessarily very small. Much of it is disseminated or occurs in threadlike stringers in the wall rock between and bordering the veins. A few quartz veins contain discontinuous stringers of chalcopyrite, and some of the quartz-pyrite veins also contain a little chalcopyrite. Veins containing massive chalcopyrite are not common except in or near the diabase sills.

Molybdenite, although a very minor constituent of the ore, is widely distributed and is easily recognizable in many parts of the mine. It is especially conspicuous where it forms selvages on the walls of pyrite veins. The rock cleaves along the selvages, exposing large surfaces plated with molybdenite. The general distribution, as determined by visual estimates, is not uniform, but the mill head usually averages 0.01 to 0.02 percent MoS_2, depending on the part of the mine from which the ore is obtained.

The only other hypogene sulfides, sphalerite and galena, are present in a few scattered veins that appear to represent a decidedly later period of mineralization than the pyrite-chalcopyrite-molybdenite phase. Some of the sphalerite-galena veins are parallel to the principal vein system but are wider and more persistent than the earlier veins. Much of the lead-zinc metallization is associated with the Dome fault system, in which it occurs as small shoots, more commonly in minor branches than in the main fault. One large shoot, 5 to 6 feet wide, in the east branch of the fault is exposed in the face of the 4,590-foot bench. This body is mainly sphalerite and quartz, with minor amounts of galena and chalcopyrite.

The amount of gold and silver in the ore is very small. The mill concentrate over a period of 9 months averaged 0.0324 ounces of gold

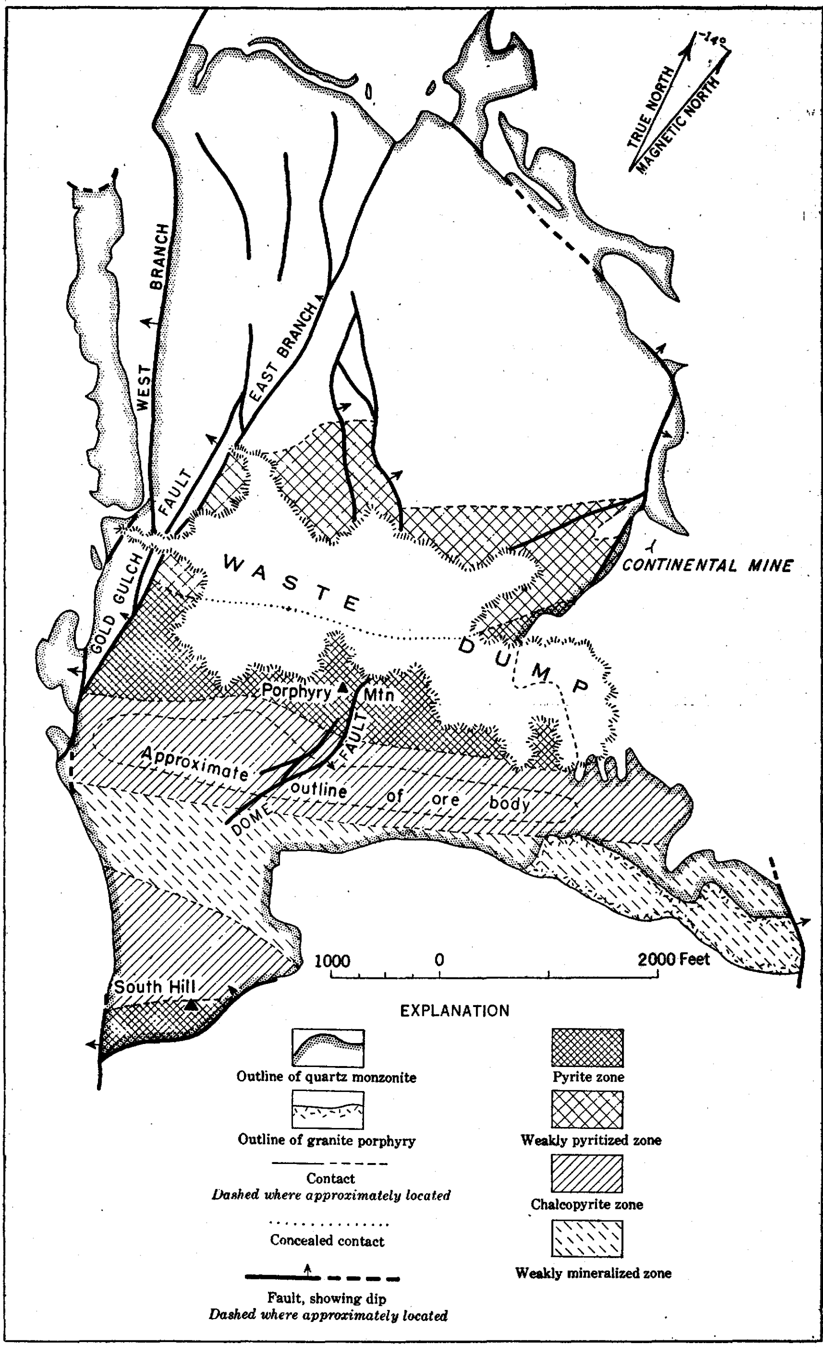

FIGURE 6.—Outline of quartz monzonite, Castle Dome deposit, Gila County, Ariz., showing generalized pattern of mineralization zones.

and 1.613 ounces of silver per ton. A few scattered assays indicate the highest gold content to occur in the vicinity of the Dome fault.

The gangue minerals of the ore are mainly the same as the constituents of the quartz monzonite host rock, together with introduced quartz, clay, sericite, and a little barite.

In a zone trending about N. 75° E. across the quartz monzonite body and through the summit of Porphyry Mountain, the mineralization consists mainly of quartz and pyrite, with very little copper (fig. 6). The hydrothermal alteration of the host rock in this zone is generally more intense than elsewhere in the mineralized area. Quartz and sericite are abundantly developed along the borders of the quartz-pyrite veins and, where the veins are especially numerous, replace all the rock minerals except original quartz. Where sericitization is less intense, the plagioclase of the rock is altered to clay minerals.

Between the pyrite zone and the south edge of the quartz monzonite body is a zone relatively rich in chalcopyrite, in which the ore body occurs. Quartz-sericite alteration is less intense in this zone, but clay alteration is generally strong. North of Porphyry Mountain the waste dumps from stripping operations overlap the pyrite zone and cover a large area that has not been studied and has been only partly explored by drilling. The few holes that were drilled showed a high proportion of pyrite and very little copper, but these holes are all under the southern half of the dump. Northward from the dump the alteration and metallization fade rapidly. A zone of weak propylitic-type alteration borders the mineralized area on the north and south.

The tenor of the protore is probably lower than that in most other disseminated copper deposits now being mined. It generally ranges from 0.1 percent copper in the pyrite zone to 0.3 percent in the chalcopyrite zone; hence very little copper sulfide is visible. Locally, in and near thin diabase sills, the rock contains massive chalcopyrite veins, and the tenor commonly increases to 1.5 percent or more. Several diabase sills 1 to 15 feet thick are exposed in the mine or were cut by exploratory drill holes. The largest and most continuous sill crops out along the south edge of the ore body and dips gently toward the north or northwest through the ore body.

EXTENT OF MINERALIZATION AND RELATION TO ROCK TYPES

The area of hypogene mineralization includes approximately the southern half of the quartz monzonite body and all of the granite porphyry body that crops out along the south margin of the quartz monzonite. At the south, mineralization extends to the contacts of the quartz monzonite with the granodiorite and schist but rarely continues more than a few feet or tens of feet beyond the contact. Along the southern part of the east contact, mineralization is relatively strong in the quartz monzonite and granite porphyry but fades abruptly in the diabase, although weak mineralization extends eastward for a distance of 1,000 feet into the diabase. Along the west side, the Gold Gulch fault may be considered the boundary, although slight mineralization extends beyond it in some places. Toward the north, mineralization in the quartz monzonite decreases gradually. Its extent in that direction is strongly influenced by structure; that is, mineralization extends farther to the north in the vicinity of northward-trending faults than elsewhere. The maximum dimensions of the mineralized area are roughly a mile from north to south by a mile from east to west, but the total area is somewhat less than a square mile.

In general, metallization is confined largely to the quartz monzonite porphyry phase of the intrusive body, but this spatial relation is considered to be coincidental and not due to any favorable characteristic of the porphyry. Within the mine area, which is the only place where a reliable comparison can be made, no appreciable difference either in character or tenor can be recognized between the protore in adjacent bodies of the quartz monzonite porphyry and the porphyritic quartz monzonite.

The hypogene metallization in the granite porphyry appears to be weaker than that in the adjacent quartz monzonite, although few data are available on which to base a comparison. Only three exploratory holes have been drilled in the granite porphyry. Two of these, at positions E. 1400, A, and E. 1400, C, on the mine coordinate grid, near the east edge of the mineralized area (pl. 1), averaged 0.22 and 0.145 percent copper, respectively, whereas a hole drilled in the quartz monzonite at position E. 1400, E, 200 feet farther north, averaged 0.487 percent copper. These average values are probably not a just comparison of the tenor of the hypogene metallization in the respective rocks, because the hole in the quartz monzonite shows more chalcocite enrichment than those in the granite porphyry. However, it appears reasonable to conclude that the quartz monzonite in the vicinity of the holes carried at least 0.3 percent copper compared to less than 0.2 percent in the granite porphyry. The third hole, at position E. 400, E, probably passes into quartz monzonite above the lower limit of oxidation and leaching. Even if the primary metallization were of the same grade as that in the quartz monzonite, the possibility of finding an ore body in the granite porphyry is remote, for there appears to be less secondary enrichment in this rock. Probably this is due to the fact that the granite porphyry is less pervious to the percolation of ground water.

The granodiorite is mineralized only near its contact with the granite porphyry, and there only slightly. In general, it shows very little evidence of hydrothermal alteration. Similarly, in the diabase intruded along the east side of the quartz monzonite and the granite porphyry, the mineralization is extremely weak. In contrast, the fine-grained diabase sills intruded into the quartz monzonite in the mine area are richly mineralized. They consistently contain at least twice as much copper as the adjacent quartz monzonite. Whether or not the fine-grained sills are offshoots of the main diabase intrusion has not been determined; however, both diabases predate mineralization. The diabase in the sills is more favorable as a host rock, probably because the sills are thin and brittle and therefore were more thoroughly brecciated than the larger masses of diabase or the quartz monzonite.

A little mineralization occurs in the Devonian limestone where it is in fault contact with pre-Cambrian quartzite on the south side of Jewel Hill east of the Castle Dome mine and, in exactly the same relation to the quartzite, in the Devonian limestone where it is exposed in Gold Gulch west of the mine. The mineralization is of the pyrometasomatic type and consists mainly of garnet, magnetite, and hematite, which replace certain limestone beds. A little oxidized copper is exposed in a few shallow workings. In both localities the limestone is separated from the quartz monzonite by diabase and is in depressed blocks on opposite sides of the quartz monzonite horst. The source of this mineralization is not known, but there is strong, if not conclusive, evidence that the mineralization is earlier than the diabase. In each example the faults along which it occurs are interpreted as predating the diabase. The small block of limestone in Gold Gulch about 800 feet south-southwest of mine coordinate W. 3600, A, is mineralized along a fault older than the diabase. It is completely surrounded by diabase and is undoubtedly an inclusion. The mineralization does not extend into the surrounding diabase; hence it is considered to have occurred before the diabase intrusion and is therefore earlier than that in the Castle Dome deposit. It may be related to the intrusion of the quartz monzonite.

HYDROTHERMAL ALTERATION OF THE HOST ROCK

GENERAL DESCRIPTION

The hydrothermal alteration that accompanied the copper mineralization in the Castle Dome area has been discussed by the authors in a previous article (Peterson, Gilbert, and Quick, 1946). It consists of three phases.

A very mild phase of the propylitic type, in which biotite and plagioclase are partly altered to sericite, epidote, clinozoisite, chlorite, and calcite, occurs in an outer zone, where mineralization is very weak, and extends beyond the area where mineralization and alteration can be recognized in the field. This zone surrounds an area of stronger alteration where most of the plagioclase and a little of the orthoclase and biotite are replaced by a montmorillonite-type clay mineral. The clay or "argillic" phase of alteration, to borrow the term used by Lovering (1941, p. 236) to describe the effects of hydrothermal alteration that has resulted in the prominent development of clay minerals, is most intense in the mine area and diminishes gradually toward the north and south, where more fresh plagioclase, together with the alteration products of the propylitic phase, is evident.

The third phase of alteration, which will be referred to as the quartz-sericite phase, is related to numerous small quartz-pyrite veins along which the wall rock is replaced by quartz, sericite, and a little pyrite and adularia. Each quartz-pyrite vein is bordered by a white zone, a fraction of an inch to several inches wide, where the wall rock has been mostly replaced by quartz and sericite. Although veins bordered by quartz-sericite alteration occur throughout the entire mineralized area, they are most numerous and largest and have the widest alteration borders near the summit of Porphyry Mountain. This area of most intense quartz-sericite alteration is entirely within the clay zone and might be said to represent a third and inner zone of alteration. However, clay alteration is strong in the wall rock between the veins, and the two phases cannot be separated areally. Thus, although quartz-sericite alteration should be regarded as a separate phase, it is superimposed on the clay phase as well as the propylitic phase and actually does not constitute a separate zone.

PROPYLITIC ALTERATION

In the marginal part of the mineralized area, entirely outside the mine, the host rock shows the effects of very weak alteration, which produced minerals characteristic of propylitization. So feeble is the alteration that the rock, in hand specimens, looks fresh except for a slight greenish coloration of the biotite due to replacement by chlorite. The minerals produced during this phase of alteration, as seen under the microscope, include pyrite, chlorite, epidote, clinozoisite, sericite, calcite, and leucoxene, but their occurrence and small amount suggest that they are the result essentially of recrystallization without appreciable change in the bulk chemical composition of the rock.

Of the original rock minerals, quartz, orthoclase, and apatite were not affected by propylitic alteration. Biotite and plagioclase are invariably altered but nowhere completely replaced. Sphene is partly or completely converted to leucoxene, but ilmenite shows only thin leucoxene rims. Sphene apparently is most susceptible to attack and is commonly altered to leucoxene where the other minerals are largely fresh.

Most of the biotite remains unaltered, but some of it is partly replaced along cleavages by green chlorite with an anomalous blue birefringence, and a few completely replaced crystals are chlorite pseudomorphs. Epidote and a little clinozoisite, in addition to chlorite, partly replace some biotite; where they are absent, calcite commonly occurs. Some of the oriented rutile needles that form inclusions in most of the unaltered biotite, and possibly some small inclusions of ilmenite and sphene, are recrystallized to scattered rutile granules or leucoxene.

Plagioclase invariably contains plates and veinlets of sericite oriented approximately parallel to the twinning plane (010) and at right angles to it. Sericite is scattered uniformly through some plagioclase crystals, but in others it occurs in patches. It is commonly accompanied by stringers and tiny, scattered grains of clinozoisite or epidote. In contrast to the altered biotite, in which more epidote than clinozoisite occurs, altered plagioclase contains much more clinozoisite than epidote. A few crystals are replaced almost completely by clinozoisite. A little chlorite is present in veinlets and as rosettes in some of the plagioclase, and where epidote and clinozoisite are absent, calcite occurs.

CLAY ALTERATION

Throughout the mine area, except in the sericitized borders of quartz-pyrite veins, the quartz monzonite looks almost fresh. The main alteration noticeable in hand specimens is in the plagioclase, which has a chalky appearance and rarely shows any twinning striations. Some of the biotite also is altered to a soft, buff micaceous mineral in the form of either pseudomorphs or aggregates. Generally, however, the rock has retained its original color and texture. Under the microscope the plagioclase appears to be largely replaced by clay minerals. Because of these clay minerals, it slakes quickly in water, and the broken rock in the mine crumbles rapidly during rainy weather. In some places the plagioclase is stained light brown, probably by included limonite in the clay. When it is moist, it commonly is pale green, owing to adsorbed copper sulfate.

The minerals of the quartz monzonite were selectively attacked in the course of the clay alteration. Quartz and orthoclase were stable except in a few places where the alteration was most intense and a few tiny veinlets and small patches of clay replace both minerals. Biotite was relatively stable but was bleached and partly replaced by clay where the alteration was severe. Plagioclase was the least stable of the original minerals, and in many places where the biotite remains nearly fresh the plagioclase has been completely replaced by clay. Sphene and ilmenite were converted to leucoxene or fine-grained rutile, and most of the apatite was destroyed.

Variations in the intensity of the clay alteration in the mine are rarely evident except under the microscope. The study of thin sections clearly shows that the amount of clay varies from place to place, but no systematic arrangement of these variations has been recognized. The alteration was probably more severe along some of the larger fractures, but every part of the rock within the clay zone was affected to some degree. This is in marked contrast to the localization of quartz-sericite alteration along veins that cut both the argillized and propylitized zones. The only systematic variation observed in the general clay alteration is a decrease in the amount of clay minerals both north and south of the mine area.

The zones of clay and propylitic alteration overlap as an ill-defined zone where both are represented. In this zone of overlap, which is best developed north of Porphyry Mountain, clay occurs in plagioclase together with scattered sericite and a little clinozoisite, epidote, and chlorite; biotite is partly replaced by chlorite and epidote but mostly is unaltered. Much of the original sphene is converted to leucoxene, but apatite is fresh. The propylitic alteration of the marginal zone is undoubtedly a feebler phase of alteration than the clay phase. The overlap might suggest that clay alteration encroached on propylitic alteration, but in the mine, which is entirely within the clay zone, nothing indicates that the rocks were propylitized before the clay alteration occurred. Fresh original biotite is common, and even slight replacement of it, or of plagioclase, by chlorite is decidedly rare. No epidote nor carbonates have been found in the mine. However, scattered fine sericite that is not apparently related to veins occurs in the plagioclase throughout the mine and probably developed ahead of the clay alteration.

The dominant clay mineral is a colorless or slightly yellowish montmorillonite-type clay that has been tested by C. S. Ross, of the Geological Survey, by differential heating methods. He reports in a personal communication: "This [test] shows that the material gives a typical montmorillonite curve, and also shows very clearly that the exchangeable base is calcium." J. S. Axelrod, also of the Geological Survey, made an X-ray analysis of the argillized plagioclase and found that it contains a member of the montmorillonite group, muscovite, and a very little plagioclase. The clay mineral has a refractive index for beta of 1.545 and a birefringence of 0.02+. It is tentatively identified as beidellite because it corresponds approximately to beidellite, described by Lovering (1941), in the altered rocks that border tungsten veins in Boulder County, Colo. Examination of the argillized plagioclase by immersion methods reveals minor amounts of another clay mineral whose refractive index is near 1.52 and a very small amount of an isotropic mineral of low refractive index that is probably allophane.

Beidellite occurs alone in veinlets, as very fine aggregates, and as scattered larger flakes. It is generally associated with some sericite and a micaceous mineral whose indices of refraction are between those of beidellite and sericite and which is undoubtedly hydrous mica. Hydrous mica cannot be distinguished from muscovite by means of X-ray patterns.

Under the microscope the replacement of plagioclase by clay can be seen in various stages of completion. Where it is incomplete, irregular patches and replacement veinlets of the clay mixture occur in the plagioclase apparently uncontrolled by the crystal structure of the plagioclase. Though the clay mixture may replace any part of the original plagioclase grains, it commonly has replaced the central part, leaving a less altered rim where replacement is less complete. In some parts of the mine the original plagioclase has been entirely replaced, and where this has happened orthoclase and biotite also have commonly been partly replaced by clay minerals, indicating a local area of more intense clay alteration.

Nearly everywhere small blades and veinlets of sericite are scattered through both plagioclase and the clay that replaced it. Whether the sericite occurs in the plagioclase or in the clay, it is generally oriented in two directions, one parallel to the twinning striations of the plagioclase and the other approximately at right angles to them. Where replacement of the plagioclase by clay is not complete, there is less sericite in the clay than in the remaining plagioclase; where the replacement is complete, sericite is less abundant or even absent. This observation by itself might indicate either that sericite replaced plagioclase more readily than clay or that clay replaced sericite developed in the plagioclase ahead of the clay alteration. Because the sericite in the clay has the same orientation that it has in the plagioclase, it seems likely that this sericite formed early and was later partly replaced by clay. Indeed, some sericite flakes scattered through the clay are ragged, as though slightly replaced, and in some partly altered plagioclase, tiny sericite veinlets along twinning planes (010) continue from the fresh plagioclase into the clay as ragged trains of sericite flakes. A few plagioclase crystals completely replaced by clay contain some large flakes of beidellite scattered through

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a very fine grained clay matrix. Many of the beidellite flakes are oriented, and some occur in parallel stringers, perhaps suggesting that the early sericite in the plagioclase has been replaced by beidellite.

A very few grains of blue tourmaline occur here and there in the argillized plagioclase.

Much of the biotite in the clay zone is unaltered, but in areas of relatively intense clay alteration it is commonly bleached. The area of most general bleaching is in the west-central part of the mine, suggesting stronger clay alteration there than elsewhere. The original brown biotite is converted to a colorless or buff mineral of micaceous habit that generally replaces the original biotite. The mineral looks much like sericite in hand specimens but is softer. Where the bleaching is incomplete, fresh biotite grades into the bleached product. The mineral has a variable but generally low birefringence and a refractive index that varies but has a minimum value of about 1.550; it is biaxial, with a negative sign and an optic axial angle ($2V$) that is small but variable. The bleaching process is probably a simple leaching of the bases from biotite without noticeable recrystallization; the ultimate product appears similar to kaolinite. In some places beidellite replaces both fresh and bleached biotite.

In the clay zone sphene and ilmenite, which are generally associated with biotite in the fresh rock, are recrystallized to leucoxene or fine-grained rutile. The outlines of euhedral sphene crystals are commonly preserved by the rutile aggregates. Where biotite is altered, the rutile inclusions originally present in it have likewise been recrystallized to tiny granules.

Apatite also is generally associated with biotite in the fresh rock and is most abundant as inclusions in the biotite. In the clay zone it is rare or absent. Where fresh biotite remains, cavities in the shape of apatite crystals indicate that apatite was originally present, but only a few grains, which are commonly corroded, remain.

All the plagioclase in the diabase sills in the mine has been replaced by clay. Although the composition of the original plagioclase in these sills is unknown, it was undoubtedly more calcic than the oligoclase in the quartz monzonite. Some of the green biotite in the sills has been bleached and partly replaced by clay, and a few veinlets of clay cut the sparse quartz grains that are generally present in the diabase. Leucoxene, developed by the alteration of ilmenite, is abundant. In addition to the clay that replaces the minerals of the diabase, thin veinlets of beidellite occur along many fractures in the brecciated sills. The rock usually breaks along these seams, and the fragments are coated with the white clay.

QUARTZ-SERICITE ALTERATION

Each quartz-pyrite vein is bordered by a narrow zone of alteration where the rock is replaced by sericite and quartz with a little pyrite and adularia. Along tiny veinlets the alteration extends only a fraction of an inch into the wall rock, but along larger veins and particularly where pyrite is abundant, the sericitized borders are commonly several inches wide. Within these glistening white border zones the original texture of the rock is either obscure or completely destroyed. The general effect is that of a layer of intense alteration separating each vein from rock of more normal appearance, and because the veins are numerous and generally parallel, the alteration borders along them appear as thin white stripes that cut through rock in which the original color and texture are generally preserved. In a few small areas in the mine, particularly north of the ore body where veins are large and most numerous, the alteration borders along the veins coalesce, and the rock is completely replaced by sericite and quartz. The outer margin of the quartz-sericite alteration along veins is not abrupt, though in hand specimens it commonly appears to be.

Sericitization persists farthest from the veins in those original minerals that are most susceptible to alteration. Close to the veins, plagioclase and clay are completely replaced by sericite, orthoclase and biotite are replaced by both quartz and sericite, and original quartz is partly replaced by sericite. Rutile and zircon are the only minerals not attacked. Beyond a certain limit original quartz and orthoclase are essentially unaltered, but sericitization of plagioclase and clay and of some biotite persists farther from the vein. Thus plagioclase and clay are most susceptible to alteration, whereas biotite is more stable but breaks down more readily than quartz or orthoclase. Apatite is rare, having largely been leached during the clay phase of alteration. What commonly appears in hand specimens to be a sharp outer limit of the sericitized border along a vein is simply that limit where orthoclase ceases to be sericitized and a change in the general color of the rock occurs because of the contrast between the fresh reddish orthoclase and the glistening white sericite.

Relatively little quartz-sericite alteration occurs along veins without pyrite or veins containing only a little pyrite associated with other sulfides. In a single small area, quartz-pyrite veins bordered by pronounced zones of alteration may be present along with other veins, containing barren quartz or chalcopyrite or molybdenite with quartz, that cut sharply through what appears to be unsericitized rock. Indeed, some chalcopyrite veinlets cut through large orthoclase phenocrysts without producing noticeable alteration, but microscopic study generally reveals a little sericitization of the wall rock along chalcopyrite- and molybdenite-bearing veins, and feldspars cut by these veins may be slightly replaced by quartz. However, none of the original minerals is completely replaced, and in hand specimens the alteration is not evident. Pyrite-free veins occur close to, or actually intersect, quartz-pyrite veins and therefore occur in the alteration zones along pyrite veins, but all positive evidence indicates that intense quartz-sericite alteration is related to the quartz-pyrite veins and that only minor silicification and sericitization accompanied the chalcopyrite and molybdenite phases of the mineralization.

Veins containing adularia can rarely be recognized in the field, but thin sections show that a small amount of adularia is common in veins containing chalcopyrite or molybdenite. Adularia also occurs in the walls of the quartz-pyrite veins, but none has been seen in these veins with the exception of a single specimen in which adularia crystals occupy a vug in a quartz-pyrite vein. Only a little adularia penetrates the wall rock as a replacement mineral, and where it does it is invariably associated with sericite. In specimens containing the most adularia sericite blades and rosettes replacing quartz commonly are separated from the quartz by thin selvages of adularia, and stringers cutting through the quartz contain sericite in the center and adularia along the margins. Probably sericite first replaced quartz, and later selective replacement of sericite by adularia occurred along the quartz-sericite boundaries. This relationship occurs in some vein quartz as well as in original quartz grains, and together with the apparent association of adularia with chalcopyrite and molybdenite, it suggests that adularia was formed during the later stages of sulfide mineralization and quartz-sericite alteration.

Not only were some of the original minerals more susceptible than others to the quartz-sericite alteration, but the character of the alteration in the various minerals is distinctive. Quartz close to the veins has been partly replaced by coarse blades and rosettes of sericite. In orthoclase, also, the sericite occurs as coarse blades, but the blades are generally oriented along planes approximately at right angles, as if controlled by the orthoclase structure, and invariably some silicification of the feldspar has taken place. In contrast, plagioclase and clay near quartz-pyrite veins have been replaced by extremely fine grained sericite without any replacement by quartz. The original plagioclase can generally be recognized in completely altered rocks along the veins by microcrystalline aggregates of sericite in the shape of feldspar grains. The probable explanation for the lack of silicification in plagioclase is that both plagioclase and the clay developed in it are very susceptible to sericitization and were rapidly and completely replaced by sericite at the beginning of the quartz-sericite phase of alteration. The plagioclase having been sericitized, replacement by quartz was probably prevented. There is no definite evidence that quartz has replaced sericite anywhere in the Castle Dome area.

Biotite has been replaced by both sericite and quartz in the alteration borders along quartz-pyrite veins. The product is a pseudomorph composed of coarse sericite and quartz containing small aggregates of rutile. In some specimens the place of the original biotite is indicated only by these rutile aggregates.

Rutile and leucoxene, developed from original sphene during the clay alteration, have likewise been recrystallized in the quartz-sericite borders along the veins to aggregates of larger grains. Commonly these aggregates also contain quartz, and a few of them preserve the outlines of original euhedral sphene.

Quartz-pyrite veins in the diabase sills, as in the quartz monzonite, are bordered by zones of intense quartz-sericite alteration.

AREAL DISTRIBUTION OF ALTERATION PHASES

Alteration and mineralization are almost entirely confined to the quartz monzonite and the granite porphyry intruded into it. Figure 7 shows a generalized arrangement of the alteration phases into several zones. These zones are not concentric but rather extend across the quartz monzonite as east-northeastward-trending bands. The boundaries between them are gradational and ill-defined. To the east, west, and south the alteration either ends at the margins of the quartz monzonite or is very weak beyond them, whereas to the north, in the quartz monzonite, the alteration decreases gradually until it becomes unrecognizable in the field.

The quartz-sericite phase of alteration is not confined within any well-defined zone but occurs throughout the mineralization area along pyrite veins and veinlets cutting through both argillized and prophylitized rock. However, a central area where pyrite mineralization and the associated quartz-sericite alteration are most intense can be clearly recognized. This is a zone passing through the summit of Porphyry Mountain and extending eastward nearly to the edge of the quartz monzonite and westward to the Gold Gulch fault (fig. 7). The amount of pyrite and quartz-sericite alteration decreases gradually both to the north and to the south; therefore, any northern or southern limit of this central zone is arbitrary and depends on the criteria chosen to define it. Because intense local sericitization and silicification occur along many faults, it is difficult to outline a single zone within which intense quartz-sericite alteration is confined, but examination in the field leaves no doubt that as Porphyry Mountain is approached either from the north or from the south, the average in-

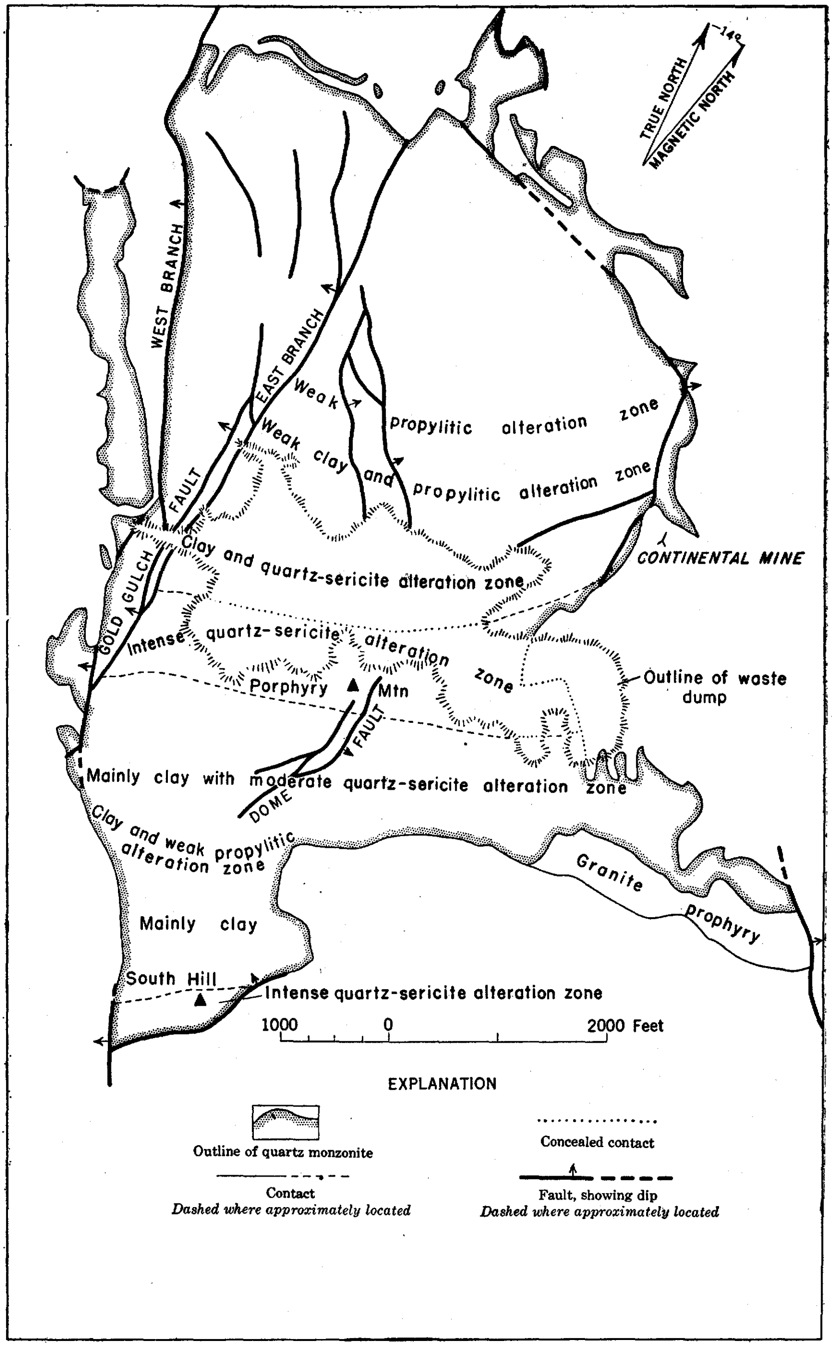


Figure 7.—Outline of quartz monzonite, Castle Dome deposit, Gila County, Ariz., showing generalized hydrothermal alteration pattern.

tensity of quartz-sericite alteration increases greatly. Reasonably definite outlines of the central zone can be described only if the zone is defined as that area containing the greatest concentration of pyrite and the largest and most numerous quartz-pyrite veins, as well as the greatest average intensity of quartz-sericite alteration. Thus defined, the central zone of generally strong quartz-sericite alteration is arbitrarily made to correspond with the pyrite zone in hypogene mineralization. Its southern boundary is in the footwall of the ore body; its northern boundary is beneath the dump. To the east and west it ends abruptly near the margins of the quartz monzonite.

The most intense quartz-sericite alteration has taken place entirely within the area where strong clay alteration occurs. Indeed, clay alteration can be seen within the central zone wherever the quartz monzonite is not completely sericitized and silicified. To the south and probably beneath the dump to the north, quartz-pyrite veins are fewer and smaller, quartz-sericite alteration is generally less, and clay is the dominant alteration product. Farther from Porphyry Mountain the amount of clay as well as quartz-sericite alteration decreases, and in the marginal parts of the mineralized area to the north small veins with thin borders of quartz-sericite alteration occur in the propylitized rock containing no clay. Clearly the veins and the quartz-sericite alteration are imposed on the slightly earlier clay and propylitic phases of alteration, but all are apparently zoned about a center on Porphyry Mountain and undoubtedly all are stages in a single mineralization process.

A second, smaller center of intense quartz-sericite alteration is on the south side of South Hill (fig. 7). Strong alteration comparable to that in the mine area extends well down the north slope of the hill. It is separated from the strongly altered area on Porphyry Mountain by relatively weak clay and propylitic alteration associated with seemingly weak pyrite-chalcopyrite mineralization, whereas on South Hill copper mineralization is shown by three exploratory drill holes to be relatively strong.

Faults controlled the variations in quartz-sericite alteration in many parts of the mineralized area. The intense alteration on South Hill (fig. 7) was undoubtedly controlled by the fault zone separating the Pinal schist from quartz monzonite on the south side of the hill. The fault itself is mineralized, and alteration decreases northward away from it. West of the mine, the Gold Gulch fault is the western boundary of strong alteration and mineralization, although the fault itself is mineralized and altered. West of it both alteration and mineralization are very weak. To the north, the east branch of the Gold Gulch fault is the western boundary of the area in which mineralization and quartz-sericite alteration are pronounced, and the alteration extends farthest northward along this fault as well as along the group of north-northwestward-trending faults a few hundred feet to the west. Intense quartz-sericite alteration also occurs along the northeastward-trending fault about 500 feet northwest of the Continental mine, and local intense sericitization is evident along the Dome fault in the southwestern part of the Castle Dome mine and along numerous minor faults elsewhere in the mine. Probably many of these same faults also influenced the distribution of both clay and propylitic alteration, but there is no direct indication that this is true.

In addition to the general study of alteration throughout the mineralized area, a more detailed study was undertaken within the Castle Dome mine. Only the exposed faces could be studied, because the benches are often regraded with waste rock brought from other parts of the mine. Each face was therefore subdivided into units 100 feet long, and each unit was carefully examined and given an average rating for each of several variable features of metallization and alteration. As the faces are 40 or 45 feet high and differ in slope from vertical to about 60°, the average area represented by each unit is 100 feet long by about 20 feet wide. In order to obtain a representative distribution of the unit areas studied, it was decided to study areas centering at the intersections of 100-foot coordinates. This plan was followed at first, but in the later part of the work it could not be coordinated with the rapid progress of mining operations and areas were studied wherever they filled large gaps between those already examined. An attempt was also made to examine adjacent areas on different days in order to reduce the personal error in estimation, but in the later stages of the work this plan could not always be followed and areas were studied in the order in which they were exposed by mining operations.

The maps (pl. 5) based on this study show variations in the intensity of alteration and metallization within the mine, but they show only those features that can be rated consistently by visual estimates—that is, amounts of pyrite, molybdenite, introduced quartz, sericite, and altered biotite. At best the estimates are approximate and relative. Silicification is probably the most difficult feature to rate accurately and consistently, because much of the quartz visible in the rock is original and most of the replacement quartz near the veins is obscured by sericite. Nevertheless, as most of the introduced quartz is in veins, the estimates probably are reasonably consistent. The recorded differences in the amount of pyrite are reliable in the unoxidized rocks but are very approximate where much oxidation and leaching have occurred. The relative amounts of molybdenite, sericite, and altered biotite can be estimated with reasonable consistency throughout the mine. Clay alteration can be recognized by examination in the field, but the differences in the amount of clay present cannot be judged accurately except by the study of thin sections under a microscope. Therefore, no map was made to show the variations in clay alteration in the mine.

Considering the mineralized area as a whole, the Castle Dome mine lies entirely within the area of greatest alteration and metallization. For this reason, the maximum variations in the degree of alteration and metallization are not included within the mine, and the area is too small to show clearly the general alteration zoning that has been described. It would be desirable to include the entire mineralized area in the detailed field study of alteration if such a plan were practical. Not only do the dumps cover large and critical areas, but any estimates that could be made from the outcrops available would not be comparable with those made in the mine. Outside the mine, only three general divisions based on the degree of alteration can be consistently recognized—much altered, little altered, and unaltered—whereas within the mine the much-altered rock has been separated into three subdivisions. Furthermore, only within the mine is there reliable information concerning the variations in sulfide deposition to which the variations in alteration are related.

Several significant features are shown on the alteration maps of the mine. Silicification is most intense north of, or in the footwall of, the ore body near the top of Porphyry Mountain. Here also the amounts of pyrite and sericite are greatest. The intensity of silicification is less in the area of the exposed ore body, and a corresponding decrease in sericitization is less apparent but nevertheless occurs there. The increased silicification and sericitization in the southern part of the mine are probably related to faults and intimate fracturing; certainly they are related to the Dome fault between mine coordinate positions W. 1800, I; W. 2400, C; and W. 2800, E. The increase in the amount of pyrite to the north is better illustrated by the sections showing the distribution of hypogene and supergene sulfides along lines W. 1800 and W. 2920 (pls. 6, 7).

All the altered biotite books resemble sericite when seen in hand specimens, and in fact many of them have been sericitized. If all the altered biotite were sericitized, the map (pl. 5*E*) showing its distribution should agree with that showing the distribution of sericite (pl. 5*D*). However, the two maps disagree in the west-central part of the mine where biotite is most altered and the amount of sericite is least. Examination under the microscope by immersion methods shows that the biotite in this area is bleached and partly altered to clay instead of being sericitized. Because biotite remains fresh in many parts of the mine during clay alteration, the general bleaching of the biotite in this area suggests that clay alteration was more intense here than in most other parts of the mine.

During the detailed study of alteration within the mine, a record was made of the distribution of two types of biotite, one occurring as books and the other as aggregates of small flakes. It was thought that the aggregate biotite, which is undoubtedly developed by recrystallization of books, might be a result of the hydrothermal alteration associated with the copper metallization. The study clearly shows that the distribution of aggregate and book biotite is erratic but that the aggregates are more abundant in the southern part of the mine and the books most abundant to the north near the top of Porphyry Mountain. In the quartz monzonite south of the mine nearly all the biotite occurs as aggregates, whereas north of the mine only books have been found. Thus there is a progressive increase from north to south in the relative amount of aggregate biotite, and there is apparently no relationship between its distribution and the alteration zoning. The study of thin sections shows that the biotite aggregates were formed before any of the products of propylitic, clay, or quartz-sericite alteration. As discussed on page 27, the recrystallization of the original biotite books may be a metamorphic effect produced by the granite porphyry intrusion.

CHEMICAL AND MINERALOGICAL CHANGES

Chemical analyses of fresh and altered quartz monzonite from the Castle Dome area are shown in table 1.

The samples representing the fresh and altered rocks were necessarily collected from widely separated areas. Samples 1 and 2, representing the fresh rock, were collected north of the dump about half a mile from the area in the mine where the altered samples were taken. Although both samples appeared fresh, microscopic examination showed a slight weathering of the plagioclase and the presence of some minerals of the propylitic phase. In order to appraise the accuracy of the sampling, two samples of the fresh rock, judged by inspection to be as similar in composition and texture as possible, were collected about 500 feet apart. To insure accurate sampling of such a coarse-grained rock, large samples were cut. These were crushed and reduced to convenient-size pulps for chemical analysis by approved sampling methods.

The two analyses agree very well in respect to those oxides that represent the essential constituents of the rock. The differences between the two are believed to represent actual differences in composition rather than errors in sampling and analysis. In the tables

TABLE 1.—*Chemical analyses of fresh and altered quartz monzonite, Castle Dome area, Ariz.*
[Norman Davidson, analyst]

	Samples				
	1 (fresh)	2 (fresh)	3 (altered)	4 (altered)	5 (altered)
SiO_2	69.60	70.68	69.20	70.33	72.15
Al_2O_3	13.99	13.82	14.33	14.53	11.77
Fe_2O_3	1.82	2.03	1.03	.50	1.00
FeO	2.15	1.83	.99	1.03	.96
MgO	.86	.44	.73	.59	.43
CaO	2.12	1.61	.25	.32	.23
Na_2O	2.66	2.64	.72	1.02	.28
K_2O	4.56	4.48	5.53	6.10	3.46
H_2O- (110° C.)	.14	.14	1.75	1.08	.02
H_2O+ (110° C.)	1.00	1.26	2.81	1.76	1.82
TiO_2	.62	.55	.66	.51	.43
P_2O_5	.23	.19	.13	.13	.08
MnO	.13	.11	.01	.02	.01
FeS_2	.02	.05	1.72	1.76	7.05
Total	99.90	99.83	99.86	99.68	99.69
Sp. gr. (bulk)	2.7	2.6	2.3	2.3	2.8
Sp. gr. (powder)	2.88	2.74	2.69	2.60	2.99

Supplementary spectrographic analyses by K. J. Murata

BaO	0.02	0.02	0.02	0.03	0.02
BeO	.0007	.0005	.0003	.0003	.0004
CoO	.0007	.0007	.001	.001	.0005
Cr_2O_3	.002	.002	.001	.001	----------
CuO	.002	.003	.1	.08	.005
MnO	.08	.08	.007	.01	.008
MoO_3	----------	----------	.001	.002	.001
NiO	.0002	.0002	.0002	.0005	.0001
SnO_2	----------	----------	----------	----------	.001
SrO	.02	.02	.01	.01	----------
V_2O_5	.008	.008	.006	.005	.007
Y_2O_3	.01	.01	.01	.01	.008
Yb_2O_3	.001	.001	.0007	.0007	.0006
ZrO_2	.07	.07	.07	.06	.06

Not found in any sample: As, Bi, Cb, Cd, Ge, In, La, Pt, Pd, Au, Ag, Re, Ta, Th, Tl, Sb, W, Zn, B.

1, 2. Unaltered quartz monzonite porphyry; half a mile due north of Porphyry Mountain.
3. Altered quartz monzonite porphyry, clay phase; Castle Dome mine.
4. Altered quartz monzonite porphyry, clay-sericite phase; Castle Dome mine.
5. Altered quartz monzonite porphyry, quartz-sericite phase; Castle Dome mine.

that follow, an average of two analyses is used to represent the composition of the fresh rock.

Table 2 is an attempt to recast the analyses for the purpose of showing the approximate mineral composition of the fresh and altered rocks. Although in detail the computed mineral composition is not strictly accurate, the major constituents conform with the general mineral composition observed during the study of about 125 thin sections. The boldest assumption involved in the construction of the table is that the original composition of all the samples before alteration was the same. Obviously this is only approximately true. Some very minor constituents were ignored, and the exact composition of many of the alteration products is uncertain. For example, the nature of the chemical changes in the bleaching of biotite is not known, nor is the degree of base exchange in the clays. The amount of orthoclase in samples 3 and 4 is based on the assumption that in these phases all the orthoclase remains unaltered—that is, the weight of the orthoclase in equivalent volumes of the fresh and the argillized rock is the same (see table 4). The study of thin sections amply justifies this assumption.

TABLE 2.—*Approximate mineral composition of fresh and altered quartz monzonite, Castle Dome area, Ariz.*

	Samples			
	1 and 2 (fresh)	3 (altered)	4 (altered)	5 (altered)
Quartz	32.66	35.03	36.23	57.10
Orthoclase	22.29	[1] 25.68	[1] 25.68	----------
Albite	22.55	6.09	8.69	2.38
Anorthite	7.03	.49	.85	----------
Biotite	7.27	----------	----------	----------
Apatite	.50	.28	.28	.17
Ilmenite	.61	----------	----------	----------
Sphene	.63	----------	----------	----------
Magnetite	2.67	----------	----------	----------
Kaolinite	[2] 3.01	----------	----------	----------
Bleached biotite	----------	3.83	3.08	----------
Rutile	----------	.66	.51	.43
Beidellite	----------	16.92	8.70	.21
Sericite	----------	[3] 6.35	[3] 11.86	29.33
Limonite	----------	2.61	2.01	2.52
Pyrite	.03	1.72	1.76	7.05

[1] Orthoclase assumed to remain unaltered in these phases. Percentage computed on the basis of equal weights in equivalent volumes of fresh and altered rocks (see table 4).
[2] Excess alumina computed as kaolinite; includes minor amounts of other alteration minerals.
[3] Hydrous mica computed as sericite.

The gains and losses of the constituents in equivalent volumes of the fresh and the altered rocks can be seen in table 3, which shows the computed weights of the various constituents in 1 cubic centimeter of rock. Although the values for samples 3 and 4 are not entirely accurate because of the difficulties encountered in determining the bulk specific gravities of these highly argillized rocks, they are considered sufficiently reliable to illustrate the general changes in composition.

Table 4 shows the weights of the mineral constituents in 1 cubic centimeter of rock. In studying the chemical and mineralogical changes illustrated by these two tables, the reader should realize that the relationship between the samples is not necessarily one of progressive change from fresh quartz monzonite through an argillized phase, as represented by sample 3, and an intermediate phase, represented by sample 4, to a rock almost completely replaced by quartz and sericite. The intermediate sample (4) was probably never as completely argillized as sample 3, and the higher plagioclase content should not be construed to indicate that plagioclase has been introduced, but rather that the plagioclase in this sample is not as completely altered as that in sample 3.

TABLE 3.—*Weight, in grams, of each constituent in 1 cubic centimeter of fresh and altered quartz monzonite, Castle Dome area, Ariz.*

	Samples			
	1 and 2 (fresh)	3 (altered)	4 (altered)	5 (altered)
SiO_2	1.858	1.592	1.617	2.020
Al_2O_3	.368	.329	.334	.330
Fe_2O_3	.051	.024	.012	.028
FeO	.053	.023	.024	.027
MgO	.017	.017	.014	.012
CaO	.050	.006	.007	.006
Na_2O	.071	.017	.023	.008
K_2O	.119	.127	.140	.097
H_2O-(110° C.)	.004	.040	.025	.001
H_2O+(110° C.)	.030	.065	.041	.051
TiO_2	.016	.015	.012	.012
P_2O_5	.006	.003	.003	.002
MnO	.003			
FeS_2	.001	.040	.041	.197

TABLE 4.—*Weight, in grams, of each mineral constituent in 1 cubic centimeter of fresh and altered quartz monzonite, Castle Dome area, Ariz.*

	Samples			
	1 and 2 (fresh)	3 (altered)	4 (altered)	5 (altered)
Quartz	0.865	0.803	0.833	1.599
Orthoclase	.591	[1].591	[1].591	
Albite	.597	.140	.200	.067
Anorthite	.186	.011	.019	
Biotite	.192			
Apatite	.013	.003	.006	.005
Ilmenite	.016			
Sphene	.017			
Magnetite	.070			
Kaolinite	.080			
Bleached biotite		.088	.071	
Rutile		.015	.012	.012
Beidellite		.389	.200	.003
Sericite		[2].146	[2].273	.821
Limonite		.070	.046	.030
Pyrite	.001	.040	.041	.197

[1] All orthoclase assumed to remain unaltered in these phases.
[2] Hydrous mica computed as sericite.

In the alteration of the fresh rock to the clay phase, only plagioclase and biotite suffer noteworthy change. The decrease in silica (table 3) represents an actual leaching that accompanies the replacement of oligoclase by clay. Quartz (table 4) decreases slightly; however, inasmuch as microscopic study shows that quartz is rarely replaced in the clay phase, the decrease shown by sample 3 may be at least partly due to a primary difference in composition. Magnesia remains constant, whereas lime, soda, and alumina decrease because of the destruction of plagioclase. Both ferric and ferrous iron are leached. Potash is added and appears as sericite and hydrous mica associated with the clay. It is highest in sample 4, in which the plagioclase contains a higher ratio of sericite to clay. Total water reaches a maximum in the most highly argillized rock and decreases as the proportion of sericite to clay increases. Titanium oxide remains virtually constant in the change from primary ilmenite and sphene to rutile and leucoxene in the altered rock. About half the apatite is leached at this stage of alteration. The residual plagioclase becomes progressively more sodic as the degree of alteration increases.

In the final stage of quartz-sericite alteration, represented by sample 5, all the original minerals of the rock are destroyed except quartz, which suffers only a minor degree of replacement by sericite. Silica increases mainly through the introduction of vein quartz. Alumina remains nearly constant after the plagioclase is destroyed. Water expelled below 110° C. decreases to a minimum with the complete substitution of quartz and sericite for clay. Potash decreases when orthoclase is replaced, but even after complete replacement of orthoclase the potash content is only a little lower than in the fresh rock. The relatively high ferric iron content of the altered rocks, especially in the quartz-sericite phase, is due to supergene products, mainly limonite, resulting from the oxidation of pyrite. Sample 5 contains a small excess of lime and magnesia that is not accounted for by any mineral recognized in thin sections of the rock. The changes in mineral composition from fresh quartz monzonite to the altered phases in equivalent volumes of rock are illustrated in figure 8.

PARAGENESIS OF THE HYPOGENE MINERALS

All the veins in the Castle Dome deposit except those containing sphalerite and galena are considered to be essentially of one age. Nevertheless, they differ sufficiently in the predominating sulfide minerals that they are referred to as pyrite, chalcopyrite, or molybdenite veins. Quartz is by far the most abundant vein mineral associated with the sulfides, and some veins consist entirely of quartz. Although most of the hypogene sulfides are confined entirely to the veins, a large proportion of the chalcopyrite, estimated at 30 to 50 percent, occurs as discrete grains disseminated in the wall rock between the veins. In most parts of the deposit the amount of sulfides is so small that, except in a few of the larger veins, the ore minerals are generally not in contact; hence their age relations are not always apparent. However, the mineral relationships are not complicated by multiple stages caused by succeeding waves of mineralization initiated by reopening of the vein fractures. All the mineralization except the lead-zinc phase appears to have taken place as a single uninterrupted event that was probably of relatively short duration.

Slight changes in the type of deposition, indicated by changes in the texture of the quartz gangue and by replacement of early minerals

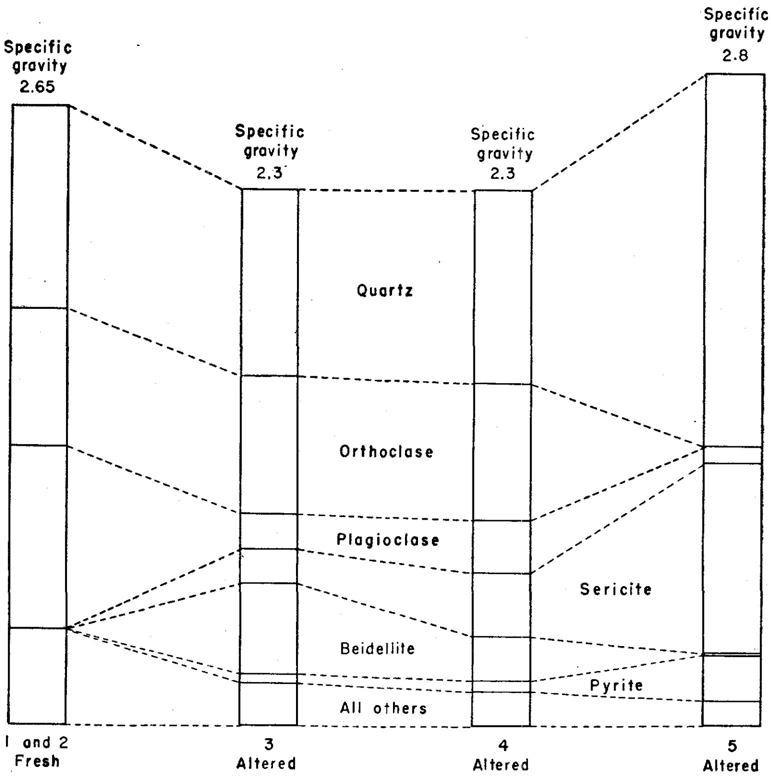


FIGURE 8.—Diagram showing mineralogic changes in alteration of 1 cubic centimeter of quartz monzonite, Castle Dome area, Gila County, Ariz.

by later ones, are commonly seen in thin sections of the vein filling, but only a very few of the veins studied show any evidence that could be interpreted to mean reopening and subsequent deposition along the new breaks. Veins commonly cross, but at the intersection they blend into one another and rarely, if ever, show relationships that can be construed to indicate difference in age even though the veins are of different types. For example, a quartz-molybdenite vein may intersect a quartz-pyrite vein and the chain of pyrite grains in the latter appear to be continuous across the intersection, yet the quartz grains in the two veins may interlock as if both had formed at the same time. If a mineralized vein were cut by a fracture that was in turn mineralized, one might expect the intersection of the two veins to show some evidence of brecciation in the earlier vein filling and recognizable continuity in the later vein at the crossing, a condition that has not been clearly recognized either in the field or in thin sections under the microscope. In the mine, quartz-chalcopyrite veins can be seen that appear to cross quartz-pyrite veins—in other words, the chain of chalcopyrite grains in one vein appears to be continuous across the pyrite filling in the other, but there are also veins showing the opposite relationship. Commonly both relationships can be observed along a single vein and even in a single hand specimen. This is probably due to zoning within the veins; that is, pyrite was deposited where the vein fractures offered least resistance to the passage of mineralizing solutions and chalcopyrite in the less permeable and hence cooler parts of the fractures. Thus, with the exception of those veins that contain sphalerite and galena, the various types of veins appear to have been deposited more or less simultaneously; therefore, the age relationships of the minerals can be determined only where they are in contact.

The age relationships, or sequence of deposition, of the various hypogene minerals is illustrated in figure 9. Quartz shows a long range of deposition, beginning before the earliest sulfides and continuing until the latest sulfides ceased to form. Most of it, however, was contemporaneous with the sulfides. The earliest deposition in the veins consisted of quartz alone followed by sulfides accompanied by quartz and coarse sericite. The age relationships of vein sericite to other minerals are difficult to determine because of the strong tendency of sericite to develop its own characteristic crystal faces. In contact with other minerals it is always euhedral. Foils and rosettes of sericite either penetrate pyrite or have pyrite molded against them. The sericite appears to be contemporaneous with pyrite and the accompanying quartz. Where chalcopyrite occurs not associated with pyrite, sericite is usually lacking also, but where chalcopyrite and sericite are associated, particularly in quartz-pyrite veins, the chalcopyrite is molded against sericite and commonly against euhedral quartz as if it had been deposited in vugs lined with projecting crystals of quartz and sericite. A little quartz occurs as veins in pyrite and, in places, appears to replace pyrite.

Where pyrite is in contact with other sulfides, the relationships indicate that in some places pyrite was clearly earlier, whereas in others all were deposited simultaneously. Chalcopyrite occurs as veinlets in pyrite grains or fills spaces between fractured grains, as in "exploded bomb" structures. Pyrite and chalcopyrite grains in contact generally show mutual boundaries that are considered by many students to indicate contemporaneous deposition. The position in the mineral sequence of the chalcopyrite disseminated in the wall rock cannot be determined, but it was probably deposited at the same time that pyrite was forming in the veins where the temperature was too high to favor the deposition of chalcopyrite.

The most easily discernible occurrence of molybdenite in the mine is as thin selvages along one or both walls of pyrite seams, but molyb-

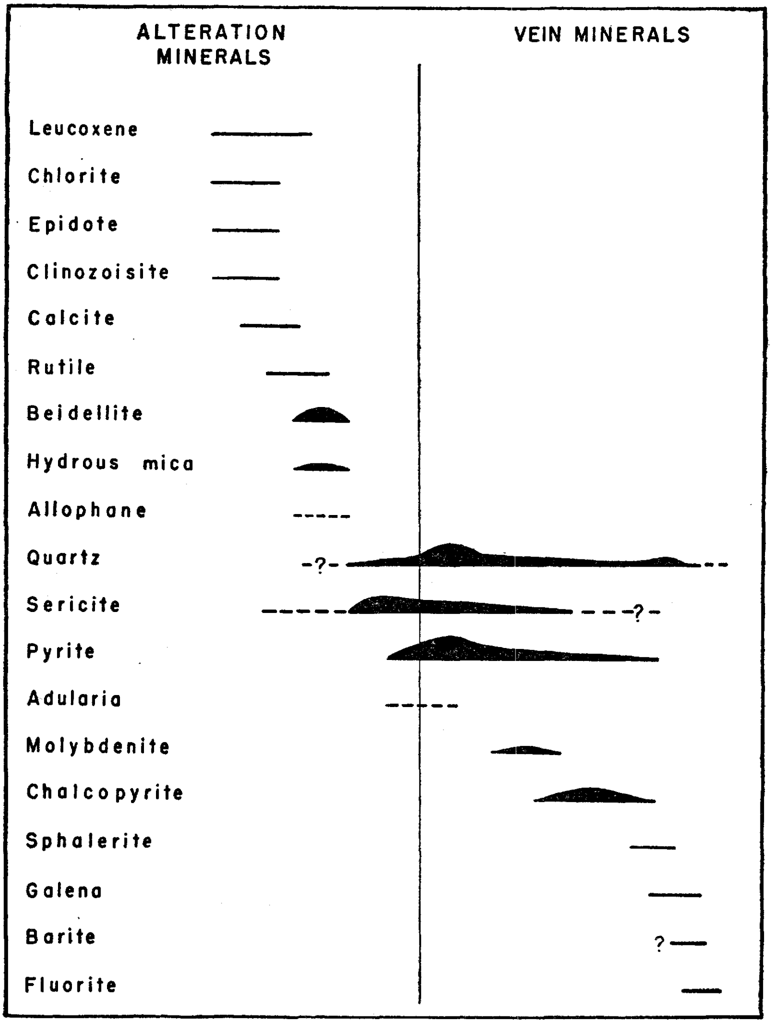


FIGURE 9.—Chart showing age relations of hypogene minerals, Castle Dome deposit, Gila County, Ariz.

denite occurs, also, in quartz veins or in the quartz associated with chalcopyrite and pyrite. In all types of veins it is generally near the margins. On polished surfaces of the veins it is seen as small rosettes, foils, and plates either in chains or disseminated among the quartz grains. Although molybdenite is commonly in contact with pyrite and chalcopyrite, the interpretation of age relations is uncertain because, like sericite, it has a strong tendency to develop its own characteristic crystal faces. Where molybdenite and chalcopyrite are associated, euhedral foils penetrate into chalcopyrite or are completely enclosed by it. In many examples chalcopyrite is intergrown with plates of molybdenite as if the two minerals formed contemporaneously. The boundaries between them are generally smooth and straight, but here and there molybdenite occurs as irregular veinlike structures and forms embayments in chalcopyrite, suggesting replacement. In contact with pyrite, molybdenite occurs as foliated masses along the margins of the pyrite or fills the spaces between pyrite grains. The boundaries are smooth but curved, and the foliations are bent, roughly paralleling the outlines of the pyrite masses. No decisive evidence indicating replacement of pyrite by molybdenite was seen; however, all the observed relationships suggest that the molybdenite is younger than most of the pyrite.

The zinc-lead veins are a separate phase of the hypogene mineralization, apparently deposited after the main period of copper mineralization. Sphalerite and galena are found only along faults and major fractures and occur in larger masses and in wider and more persistent veins than are characteristic of the more general pyrite-molybdenite-chalcopyrite phase. The crude banded structures of the veins parallel to the walls suggest that they were deposited in open fractures. The spatial relation of the minerals is consistent in most veins. A band of quartz, generally enclosing a little pyrite, lies next to the walls. In some veins the quartz shows coarse comb structure, with prisms up to an inch long projecting toward the middle. Next is a band of sphalerite that is partly replaced by chalcopyrite, especially near the outer edges. Coarsely crystalline galena commonly forms the middle part of the vein. Many of the vein fractures are not completely filled, and coarse crystals of galena project into the open spaces.

Microscopic examination of polished surfaces of the vein filling shows that pyrite is the earliest sulfide and is always associated with quartz. Some pyrite appears to be replaced by sphalerite, but commonly euhedral grains of pyrite are completely enclosed by sphalerite. Both chalcopyrite and sphalerite fill spaces between fractured pyrite grains in typical "exploded bomb" structures. Some of the sphalerite is massive, but most of it is intergrown with quartz. Where chalcopyrite and galena are in contact, they show mutual boundaries and were probably contemporaneous; both clearly replace sphalerite.

Gold and silver are present in very small amounts, and no minerals of these metals were found in any of the specimens examined. They probably occur in very minute quantities in chalcopyrite or sphalerite. The few samples of the ore that have been assayed for gold and silver indicate that the highest values are concentrated along the Dome fault system.

Hematite of definite hypogene origin is so rare that its age relations to other minerals cannot be determined with certainty. In one polished section specular hematite forms a fine-grained aggregate with quartz on the borders of a quartz-pyrite vein. In this one specimen the hematite and associated quartz are later than the pyrite, which they clearly replace.

Barite, although fairly common in parts of the mine, was not found associated with sulfides. Clusters of euhedral crystals coat the walls of open fractures that are partly filled with limonite and jarosite. Many of the barite crystals are covered by thin crusts of euhedral quartz that is probably supergene. Although limonite commonly is abundant in the barite fractures, it is of the transported type, and no evidence was found that sulfides were ever present in the fractures. However, pyrite and chalcopyrite commonly occur disseminated in the wall rock less than a quarter of an inch away. The barite is undoubtedly hypogene and probably represents a late phase of the mineralization.

In some of the faults northwest of the mineralized area, plates of barite are intergrown with massive vein quartz. The outcrops are oxidized and contain some limonite, and most of the barite has been leached out, leaving negative pseudomorphs of the platy crystals in the quartz. Small, scattered pyrite grains are present in the quartz in these veins.

Fluorite (CaF_2) has been found only in the southern part of the mine in fractures that are within the Dome fault system. It occurs most commonly as small cubic crystals, 1 to 2 millimeters across, attached to the walls of fractures. In a few places, cavities up to 3 inches across are filled with vuggy masses of large cubic crystals intergrown with thin plates of barite and a little comb quartz. The fluorite cubes formed around the barite plates and commonly engulf them. The relationship of the crystals indicates that some fluorite continued to form after the deposition of barite had ceased.

Another group of minerals, which may either represent the final ebbing stage of the copper mineralization or belong to a later period of mineralization from an independent source, includes wavellite [$3Al_2O_3.2P_2O_5.13(H_2O,HF)$], metatorbernite ($CuO.UO_3.P_2O_6.8H_2O$), wulfenite ($PbO.MoO_3$), and possibly libethenite ($4CuO.P_2O_5.H_2O$). These minerals have been discussed previously by Peterson (1947). They are present in relatively small amounts and thus far have been found only within definite small areas in the mine.

Wavellite is the most abundant and widely distributed mineral of this group. It occurs mainly in a small elongate area between coordinate positions W. 1000, G, and W. 1600, G (pl. 3), and appears to be localized along fractures, with a moderate northerly to northwesterly dip, that cut the steeply southward-dipping veins at approximately right angles. Where open spaces occur along fractures, wavellite forms encrustations on the walls and on earlier minerals deposited on the walls of the fractures. The crusts consist of coalescing hemispherical masses up to 4 millimeters in diameter with radiating internal structure. The surfaces of the hemispheres glisten with light reflected from countless crystal faces. Some of the crusts are clear and colorless; others are gray, yellow, or pale green. In the weathered zone the crusts are usually a translucent white but may be stained by iron.

The metatorbernite in the Castle Dome deposit is always associated with wavellite and, almost without exception, was deposited on wavellite crusts. It occurs as very thin, rectangular, bright-green plates that are usually arranged in tiny rosettes of extraordinary beauty. In some places the crystals form a mosslike mat covering the wavellite crusts. Examined in immersion oils, the crystals are seen to be rectangular or to show rectangular cleavage. Plates lying face up show no birefringence and have an index of refraction of 1.623. The other index is slightly higher. Tilted plates show anomalous purple to violet interference colors and slight pleochroism from yellowish green to greenish blue. These properties identify the mineral as metatorbernite I, described by Hallimond (1920), who first recognized it, as an artificial mineral obtained by the dehydration of torbernite (Hallimond, 1916).

The first inversion of torbernite in contact with water occurs at 75° C. It is accompanied by a marked change in optical properties and a loss of four molecules of water. A second inversion takes place at 130° C. with further loss of water. The two products have been designated as metatorbernite I and II, respectively. Hallimond was unable to bring about a reversal of the transformations. Metatorbernite I was later recognized as a natural mineral in Spain and Cornwall by N. L. Bowen (1919).

Libethenite, a rare copper phosphate, has been found only along the Dome faults and related fractures. It occurs as crusts on the walls of open fractures. The crusts are generally composed of tufts of small, emerald-green, orthorhombic prisms or of a drusy mat of acicular crystals. In places discrete crystals, some so small they cannot be recognized without the aid of a lens, others up to 1.5 millimeters long, lie flat on the rock surfaces.

Jewell Glass, of the United States Geological Survey, studied several specimens of the mineral and described its physical and optical properties as follows:

Crystals are usually short prisms two to five times as long as they are thick, passing into acicular forms grouped in radiating clusters, united in druses. A few crystals show pyramidal habit, resembling octahedrons. Good prismatic cleavage, luster vitreous, color emerald to yellowish green, transparent.

Optically the mineral is biaxial negative, $2V$ large—about 85°. An optic axis emerges normal to a prismatic cleavage or a side pinacoid and looks like the familiar optic axis on epidote grains. The plane of the optic axis is across the elongation. Twinning is observed occasionally. Dispersion is strong $r>v$. Thin fragments are pale green; pleochroism faint, very seldom distinguishable. Extinction is inclined 0° to 20°, greater than 20° in a few grains. Indices of refraction, average of several readings on each specimen, are: $\alpha=1.703$, $\beta=1.746$, $\gamma=1.789$. Birefringence=0.086, all, ±0.002.

Wulfenite, the molybdate of lead, has been found in only one locality in the mine, near coordinate position W. 2400, E, on the 4,350-foot and 4,310-foot benches, where it occurs in close association with libethenite. It was identified on fragments of rock broken by blasting and has not been seen in place. Instead of the usual tabular crystals, the Castle Dome wulfenite occurs as small, pointed, tetragonal prisms, rarely over 3 millimeters long and generally much smaller, which are attached to fragments of rock that apparently formed the walls of open fractures. The surfaces of the rock, as well as some of the crystals, are coated with limonite, jarosite, canbyite, and malachite, all of undoubted supergene origin.

Although the libethenite and wulfenite may well have been formed by supergene processes, the lack of valid evidence of supergene origin in the case of wavellite and metatorbernite suggests that they were deposited by late hypogene solutions. No minerals have been recognized in the deposit from which uranium could have been derived by weathering processes. If it had been leached from undetectable traces present in the country rock, one would expect the metatorbernite to have a more general distribution instead of being concentrated in a relatively small part of the mineralized area. The absence of an obvious source of uranium is probably not conclusive evidence that metatorbernite was deposited by hypogene solutions; however, no evidence has been found to date that suggests a supergene origin.

Metatorbernite is less widely distributed than wavellite, which commonly occurs alone; however, metatorbernite is never found unaccompanied by wavellite. It usually occurs on wavellite crusts and is clearly the younger mineral. Metatorbernite has not been recognized in the leached capping. Most of it has been found in the lower part of the secondary sulfide zone, but the maximum depth at which it occurs cannot as yet be determined. In a few places, metatorbernite has been dissolved by supergene solutions, leaving a very little residue of a yellowish material that resembles kaolinite.

The fact that torbernite inverts to metatorbernite I at 75° C. and to metatorbernite II at 130° C. may indicate that metatorbernite I formed at a temperature between 75° and 130° C., which is certainly above the range of supergene solutions. However, it has not been produced artificially except by dehydration of torbernite; hence there is as yet no proof that metatorbernite I cannot be precipitated from solution at temperatures lower than 75° C.

Jewell Glass, in a personal communication, describes an instance in which the inversion of torbernite to metatorbernite I unquestionably took place at room temperature. A specimen of torbernite from Spruce Pine, N. C., stored in a mineral cabinet in the Geological Survey laboratory, was reexamined after a lapse of about 10 years and was found to have inverted to metatorbernite I. The time required for the inversion to take place at room temperature is not known, because no interim examination of the specimen was made; however, it was observed that after a few months, or perhaps years, the emerald-green color of the torbernite had faded to a dull siskin green and the crystals had lost their transparency.

Very little wavellite has been found in the weathered zone. In a few places in the upper part of the ore body, the wavellite shows evidence of attack by supergene solutions and some of it is stained by iron oxides, but much of it occurs in fractures where chalcopyrite is only slightly tarnished by chalcocite and where no other supergene minerals are present. Wavellite commonly crusts pyrite and chalcopyrite, and in one place it was deposited on fresh galena and sphalerite. Many examples of wavellite crusts on barite crystals have been observed.

Wavellite probably forms as a supergene mineral in the weathering of phosphate deposits, but where it occurs in vein deposits it is most likely hypogene, although opinions differ on this point. Lindgren and Ransome (1906, p. 176) describe hypogene wavellite associated with adularia in the Cripple Creek district. Russell (Hey, Bannister, and Russell, 1938, p. 51) describes wavellite in the Castle-an-Dinas wolfram mine in Cornwall so as to suggest to the reader that it is a late hydrothermal mineral. The wavellite in the tin veins of Llallagua, Bolivia, is considered hypogene by Turneaure (1935, p. 60) and Bandy (1942, p. 330), but Samoyloff (1934, p. 495) and Ahlfeld (1931, p. 253) include it among the supergene minerals. In a later discussion of the same tin deposit, Ahlfeld (1936, p. 220) describes wavellite as one of the latest hypogene minerals.

Libethenite crystals are always present on those surfaces of rock fragments to which wulfenite crystals are attached, but many specimens contain libethenite crystals unaccompanied by wulfenite. Both minerals occur in fractured rocks where oxidation and leaching of sulfides are probably complete. The fractures contain abundant limonite, canbyite, malachite, and jarosite. These minerals are clearly younger than the wulfenite, but their relationship to libethenite is not clear. In some places libethenite crystals are engulfed by them; in other places libethenite crystals appear to have formed on limonite crusts. Clumps of libethenite crystals are commonly attached to wulfenite in such a manner as to suggest that libethenite is decidedly the younger mineral. Thin botryoidal crusts of a black manganiferous mineral resembling pyrolusite are often found on the same rock surfaces with libethenite, which in some examples forms crusts on the manganese mineral. No other occurrence of manganese oxides has been noted in the Castle Dome deposit; hence it is considered to be related to this phase of mineralization. On some rock fragments, barite crystals are attached to the same surfaces as are wulfenite and libethenite, and both minerals appear to have formed on barite crystals.

Although wulfenite and some of the libethenite are clearly older than such definitely supergene minerals as limonite, malachite, canbyite, and jarosite, little evidence can be cited to support a postulation that they were deposited by late hypogene solutions. The usual difficulty in accounting for a source of molybdenum to produce wulfenite by supergene processes is solved in this deposit by the presence of molybdenite, which, although very resistant to oxidation, has been altered to the oxide to some extent; therefore, some molybdenum may have been transported by supergene solutions. Likewise, a source of lead is available in galena, which is more abundant along the Dome fault system than elsewhere in the deposit. However, Dittler's experiments (1914) on the synthetic production of wulfenite indicate that an alkalic environment is necessary and therefore that wulfenite is not likely to be formed by supergene solutions. He concludes that it is formed by rising alkaline solutions that have reacted with lead carbonate.

Sources of the constituents of libethenite also are readily available in this deposit. The copper could have been derived from hypogene chalcopyrite, and the phosphate ion from apatite or wavellite. Lindgren (1905, p. 118) found small crystals of libethenite in cavities and seams in quartzite in the oxidized zone of the Coronado vein in the Clifton-Morenci district of Arizona. He considered it to be a supergene mineral.

On the basis of field evidence, wulfenite and libethenite are clearly late minerals and may well be of supergene origin; however, wulfenite is earlier than libethenite, and the two minerals are not necessarily genetically related. Their association could easily be coincidental.

ZONING IN THE HYPOGENE MINERALIZATION

Evidence of zoning in the distribution and relative abundance of hypogene copper and iron sulfides is clearly recognizable. That the distribution of pyrite and chalcopyrite is the result of zoning rather than other causes is confirmed by its conformity with the hypogene alteration pattern, which shows the same intensity relations as the sulfide minerals in accordance with the generally observed and accepted sequence.

The distribution and relative abundance of pyrite in the mine area, as determined by visual estimates in the field, is shown on plate 5*B*. In spite of minor inconsistencies and the difficulty of estimating the amount of pyrite in near-surface areas where much of it has been destroyed by oxidation, the chart shows a distinct increase from south to north. The similarity of this chart to those showing the distribution and relative abundance of sericite (pl. 5*D*) and introduced quartz (pl. 5*C*) shows the close relationship of the three minerals. The relative abundance of sericite and introduced quartz serves as a measure of the intensity of hydrothermal alteration.

The increase in the amount of pyrite from south to north is even more strikingly illustrated by the pyrite graphs shown on plate 6 and plate 7. These graphs show the relative abundance of pyrite based on quantitative estimates made by means of polished briquettes of concentrate panned from churn-drill sludges. On section W. 1800 (pl. 6) the relative amount of pyrite doubles within a distance of 700 feet, and on section W. 2920 (pl. 7) the increase is even more accentuated. Comparison of these sections with plate 5*B* shows that the interval represented by the sections is rated "little" to "intermediate" in pyrite content as determined by field estimates, and northward the increase continues to a maximum near coordinate line *O*, where sericitization and silicification also reach maximum intensity. Pyrite, although it commonly has a long range of deposition, is generally one of the earliest sulfide minerals and commonly represents deposition at relatively high temperatures.

As the variations in the amount of pyrite are gradational, the delineation of a high-pyrite zone is necessarily arbitrary. The same is true of the zone of maximum quartz-sericite alteration, but the two zones are closely related, and if properly selected standards are applied, they seem approximately to coincide. The high-pyrite zone trends about N. 75° E. across the quartz monzonite body. Its southern boundary is approximately on coordinate line *K*; its northern limit is concealed by the dump on the north side of Porphyry Mountain but is probably near coordinate line *S*. This zone represents the most intense phase of metallization and host-rock alteration.

Chalcopyrite, on the other hand, is generally considered to represent a lower-temperature phase of mineralization than pyrite and hence would be expected to show maximum concentration in a zone bordering the area of most intense alteration and pyritization. An analysis of the distribution of chalcopyrite is complicated by supergene leaching and enrichment, by the stronger metallization associated with the diabase sills in the mine area, and also by lack of information regarding the northern part of the mineralized area, where few exploratory holes have been drilled and much of the surface is concealed by waste dumps. However, enough information is available to show clearly that the average copper content of the protore in the vicinity of the ore body was higher than that in the high-pyrite zone.

The zone of maximum copper deposition lies south of the pyrite zone and is bounded approximately by coordinate lines *E* and *K*. It extends eastward nearly to the edge of the quartz monzonite and westward to the Gold Gulch fault. The Castle Dome ore body is largely within these limits. Between coordinate line *E* and the south edge of the quartz monzonite, the protore appears to have been poor in copper except in a small area south of the west end of the mine. This area, in which the protore was probably comparable in grade to that in the main chalcopyrite zone, is in a lobe of quartz monzonite extending about 1,800 feet beyond the main south contact and underlies South Hill (pl. 1), a small rounded hill south of Gold Gulch. Along the west side of this lobe, the quartz monzonite is in contact with diabase; at the south and southeast it is in contact with schist along a fault dipping 70° N. In a zone about 200 feet wide along the fault, the quartz monzonite is highly sericitized and is probably relatively poor in chalcopyrite, although no accurate estimate can be made on the basis of capping studies because of the high pyrite-to-chalcopyrite ratio of the mineralization. North of this sericitized zone enough chalcopyrite was deposited to produce a protore that was raised to ore grade by supergene enrichment, as shown by three churn-drill holes. A poorly mineralized belt from 600 to 1,200 feet wide, showing weak propylitic-type alteration, lies between the two chalcopyrite zones.

The difference in the tenor of the protore attributable to zoning is small. The copper content of the rock unaffected by supergene alteration ranges from about 0.1 percent in the pyrite zone to about 0.3 percent in the chalcopyrite zone. Six exploratory holes drilled in the pyrite zone show average grades of 0.10 to 0.15 percent copper and an appropriately high pyrite content. In the chalcopyrite zone, drill-hole samples of rock unaffected by supergene enrichment and well removed from the influence of diabase sills range from 0.15 to 0.4 percent copper.

One might well expect to find a symmetrical arrangement of alteration and metallization surrounding the area of greatest intensity. North of Porphyry Mountain, the intensity of alteration decreases from a maximum near the summit to a minimum represented by the north limit of mineralization. Between these limits there is probably a zone of intermediate alteration corresponding to that on the south slope of Porphyry Mountain, but no evidence has been found to support an assumption that there is a corresponding zone of higher copper content on the north slope. If it is present, it is entirely covered by the waste dumps.

Seven exploratory holes have recently been drilled in the mineralized area north of the waste dumps at locations where residual limonite textures in the capping appeared to indicate a relatively high concentration of chalcopyrite compared with other parts of that area. The sludges from these holes showed a relatively weak mineralization that consisted mainly of pyrite. The copper assays of rock unaffected by supergene leaching and enrichment are remarkably uniform in these holes, averaging about 0.10 percent copper.

Judging from what can be seen on the surface now exposed and the data obtained from these drill holes, the type of mineralization characteristic of the high-pyrite zone continues northward, but with decreasing intensity and with no apparent increase in the amount of chalcopyrite in the intermediate zone.

Molybdenite is widely distributed, and traces of it were found in nearly every part of the mineralized area; however, an accurate quantitative estimate of its distribution is difficult because the amount present is very small compared with the other sulfides. As shown on plate 5*F*, which is based on visual estimates made in the field, the major concentration occurs in the area in which the copper content is highest—that is, mainly within the limits of the ore body but also along the footwall of the ore body in the transitional zone in which the amount of chalcopyrite decreases and pyrite increases. Very little molybdenite was found in the churn-drill sludges, undoubtedly because the flaky mineral was easily lost in panning. For this reason an estimate of its distribution based on a study of the concentrate is considered unreliable and is not shown on plates 6 and 7. However, the field observations clearly indicate that the distribution of MoS_2 is related to the general pattern shown by the chalcopyrite and pyrite zones. It is most abundant in the chalcopyrite zone and in the southern part of the pyrite zone. This apparent overlap suggests that molybdenite is intermediate between pyrite and chalcopyrite with respect to the temperature at which deposition took place.

Sphalerite and galena are considered to represent a late phase of mineralization associated with major fractures and faults that formed or were reopened by renewed movement after the main period of pyrite-molybdenite-chalcopyrite deposition. Consequently this later mineralization is superimposed on the earlier phase instead of being in a marginal zone as might be expected. The amount of sphalerite found in the drill-hole sludges was very small; therefore, the distribution shown on plates 6 and 7, which is based on the few grains identified in the sludge concentrate, is probably not significant. The more general occurrence along section W. 1800, however, could be taken as evidence to confirm the field observation that sphalerite is generally associated with faults—particularly the Dome fault system, which is very prominent on this section.

The distribution of barite within the mine area also suggests a relationship to the sulfide zoning. Barite was first recognized in the initial cuts on the 4,310-foot and 4,265-foot benches, where it was fairly abundant in the southern part of the chalcopyrite zone. The distribution illustrated on plates 6 and 7 shows that it is not generally associated with sphalerite and galena but offers some support to the observation that it occurs mainly in the chalcopyrite zone.

Thus the hypogene mineralization on the south slope of Porphyry Mountain shows a succession of minerals in roughly parallel zones (fig. 6), with pyrite predominating in the most intensely sericitized and silicified area at the north followed by molybdenite, chalcopyrite, and barite in the progressively less altered zones toward the south. Sphalerite and galena represent a distinctly later phase of the mineralization confined mainly to faults and major fractures that cut across all the other zones. North of Porphyry Mountain, zones corresponding to those on the south slope have not been recognized, and the mineralization of the pyrite zone appears to decrease in intensity toward the north in proportion to the decrease in hydrothermal alteration.

STRUCTURAL CONTROL

Although the hypogene mineralization is distributed throughout the southern half of the quartz monzonite body, the principal locus of deposition in this area is a set of closely spaced, nearly parallel fractures. These fractures, now represented by the veins, dip steeply to the south and in general strike N. 75° E., roughly parallel to the south contact of the quartz monzonite body. The local variations in strike and dip, as recorded in various parts of the mine, are shown on plate 3, and a statistical analysis of the vein pattern is illustrated by figure 5. The veins are spaced a few inches to a foot or more apart; the greatest number are probably between 6 and 10 inches apart. There are a few places in the mine where they cannot be recognized, but in these places they are probably obscured by intense alteration or later fractures. The pattern of mineral zoning seems clearly governed by the veins; that is, the zones extend across the quartz monzonite body parallel to the strike of the veins. Whether or not the localization of metallization in the southern half of the area was the result of a greater concentration of fractures in that locality is uncertain. The fractures are certainly more obvious where they contain mineral veins and where fresh rock surfaces have been exposed by mining operations than they are in the weathered outcrops of the unmineralized area to the north.

Northwest-striking joints and fractures with a northeasterly dip are abundant throughout the mine area, especially east of the Dome fault. Only a very few of these are mineralized.

The vein fractures were probably only secondary structures by means of which the solutions were disseminated through the mineralized area. The main feeder channels along which solutions were conducted from their source to the minor fractures have not been identified. The most intense mineralization and alteration occur in a zone passing through the summit of Porphyry Mountain; this suggests that the main conduit lies in that vicinity, possibly parallel to the vein system. If such a structure exists, it is entirely concealed by dumps, and no direct evidence of its presence has been found.

Of the known faults that predate mineralization and may have served as feeder channels, the Dome and Gold Gulch fault systems are the most prominent. Both are mineralized to some extent, but the copper content of the rock in the vicinity of the faults is not appreciably higher than elsewhere. The alteration along the Dome fault system is perceptibly stronger than average but not as strong as would be expected along a major channel. The mineralization associated with the fault is mainly of the late phase consisting of sphalerite, galena, and a little chalcopyrite and pyrite. The Gold Gulch fault is the western limit of strong alteration and mineralization. The fault zone is mineralized, but mainly with quartz, barite, and limonite, the latter probably derived from a little pyrite. Wall-rock alteration adjacent to the east side of the fault is exceptionally strong, whereas that on the west is very weak, indicating that the course of hypogene solutions was influenced by the fault. Southeast of South Hill the fault that forms the contact between quartz monzonite and schist (pl. 1) was probably the channelway for the copper-bearing solutions that produced the small ore body underlying the hill.

Alteration in the quartz monzonite is strong along the fault and decreases gradually toward the northwest. There is only slight alteration in the schist south of the fault. Several other faults north of Porphyry Mountain were at least minor channels for solutions, because alteration and mineralization penetrated farther northward in areas adjacent to one or both sides of the faults than elsewhere (fig. 6).

Although the faults mentioned in the preceding paragraph controlled to some extent the course followed by the mineralizing solutions, they appear to have had little effect on the general distribution of copper minerals. The vein fractures provided the principal openings through which the solutions pervaded the rocks and in which the minerals were deposited. Collectively they may constitute a sufficiently permeable zone to represent the main channel. It is possible, however, that the localization of mineralization was not primarily controlled by structure but is related to some inscrutable feature, the nature of which can be only a subject of speculation.

INFLUENCE OF THE DIABASE SILLS

Several thin diabase sills from 1 to 15 feet thick crop out along the south edge of the chalcopyrite zone and dip at low angles toward the north or northwest through the chalcopyrite and pyrite zones. The diabase in these sills is thoroughly brecciated and strongly mineralized. The mineralized veins, which dip approximately at right angles to the dip of the sills, generally do not continue through them, but abundant pyrite and chalcopyrite were deposited in random fractures in the brecciated diabase. Much of the metallization in the sills occurs as films of sulfides in the cracks between the breccia fragments, but large, discontinuous, irregular veins containing massive chalcopyrite and a little quartz are common in the diabase and in the quartz monzonite near the sills.

In most places the maximum copper content coincides with the position of the diabase sills as recorded in the drill logs. Part of the increase is due to supergene enrichment, but there is also a marked increase in the hypogene metallization, as is illustrated by the hypogene copper and pyrite graphs on plates 6*B* and 7*B*. The pyrite graphs show prominent bulges at or close below the diabase, and because pyrite is strictly of hypogene origin, the increase must be interpreted as stronger hypogene metallization. The drill sludges show no evidence that the lower pyrite content above the diabase resulted from the destruction of pyrite by supergene oxidation and leaching.

The copper assays in the diabase sills within the chalcopyrite zone range from 0.75 to 7 percent, and the average is about 1.5 percent. The concentration of copper in and near the sills is clearly related to the pattern of hypogene zoning. As shown by drill-log data, the copper content of the sills within the high-pyrite zone is in proportion to the lower average grade of the quartz monzonite within that zone. A typical example is the drill hole at position W. 2400, M, which cuts a diabase sill about 15 feet thick. The average copper content in 200 feet of quartz monzonite below the leached cappings is about 0.12 percent; for 15 feet of diabase it is 0.33 percent, compared to an average grade of about 1.5 percent for the diabase in the chalcopyrite zone.

Although not as richly mineralized as the sills, the quartz monzonite adjacent to one or both sides of the sills contains considerably more copper than the general average of the zone in which the sills occur. The effect of the sills on the metallization of the adjacent rock is clearly illustrated along the north side of the ore body, where, in many places, tongues of ore 50 to 100 feet thick extend northward into the high-pyrite zone, following the diabase sills. As the sills continue deeper into the high-pyrite zone, the tongues contract in thickness and their copper content decreases; however, the copper content of the diabase in the sills remains at least double that of the adjacent rock.

The stronger metallization associated with the sills appears to be the result of favorable physical rather than compositional characteristics of the rock. Wherever the sills have been exposed by mining, they are thoroughly shattered. A few veins in the quartz monzonite continue into the sills, but most of them end at or near the contacts. The age of the vein fractures with respect to the sills is uncertain. The lack of continuity of the veins through the diabase might suggest that the sills were later than the vein fractures and were not subjected to the same stresses. On the other hand, if they were earlier, the slight adjustments that produced the parallel fractures in the quartz monzonite may have caused shattering in the brittle diabase.

The vein system in the quartz monzonite is approximately at right angles to the gently dipping sills; solutions rising in the vein fractures were diverted by the diabase, and the brecciated sills became channels in which solutions collected and moved, thus favoring the concentration of ore minerals in and near them. The rich copper metallization associated with the diabase sills was an important contributing factor in the formation of the ore body.

CHARACTER OF THE MINERALIZING SOLUTIONS

What little can be said concerning the character of the mineralizing solutions must be inferred from the effects produced in the rocks within the present range of observation, approximately 900 feet in vertical extent. The minerals formed in the Castle Dome deposit indicate that the solutions were rich in silica, iron, and sulfur and also contained copper, molybdenum, zinc, lead, barium, and probably a little phosphoric acid, fluorine, and carbon dioxide. Probably many other metals and radicals were present in the solutions, but the physical and chemical conditions existing at the horizon now exposed did not favor their deposition as minerals. Most of the sulfur was deposited as metallic sulfides, but some barite formed during the late stages of mineralization. The minerals produced in the altered host rock are among those generally considered to form in at least a mildly alkalic environment. Although the solutions may be assumed to have been acid nearer their source, no evidence has been found to suggest that they were still acid when they reached the rocks now exposed. Whether or not they carried potash from their magma source is debatable.

The general pyrite-molybdenite-chalcopyrite mineralization and the alteration of the host rock are considered to have taken place as a single event of relatively short duration. Along the most permeable channels leading from the source of mineralization, quartz and pyrite were deposited. Bordering these channels, the wall rock was altered to quartz and sericite. Here potash was removed to some extent, owing to the replacement of orthoclase by quartz and sericite. True, it is commonly believed that potash is added during sericitization, but although this is generally correct in the case of basic rocks, it is not necessarily so for acid or intermediate rocks.

As mineralization continued, the solutions spread out into the adjacent less permeable and cooler rock, where they deposited quartz, molybdenite, and chalcopyrite. Their character was changed by reaction with the wall rocks; their ability to produce sericite decreased, but they continued to leach lime, soda, iron, silica, and alumina and altered the plagioclase in the rock to clay minerals, mainly beidellite. Some potash was introduced to form the sericite and hydrous mica disseminated in the clay. If it is assumed that silicification and sericitization are more intense at depth, the potash added in the clay zone could easily be accounted for as having been leached from deeper rocks.

The nearly spent solutions that soaked farthest from the main paths of circulation deposited a little pyrite and carbonate and altered certain minerals in the surrounding quartz monzonite to chlorite, epidote, clinozoisite, calcite, and sericite. These solutions, in order to produce this propylitic-type alteration in the outlying rock, must have contained some H_2S and CO_2. A little magnesia, lime, potash, and iron also may have been added; at least the solutions were sufficiently concentrated in these bases to prevent further leaching. Propylitization appears to be largely a process whereby the constituents of certain rock minerals are rearranged to form minerals more stable under mild hydrothermal conditions. In general its effect is greatest on basic rocks in which the constituent minerals are generally less stable under hydrothermal attack than the minerals characteristic of the acid rocks. Analyses commonly show that magnesia, lime, and CO_2 were introduced during propylitization.

As the solutions continued to flow and the temperature of the rock between the channels increased, the various zones of metallization and alteration spread farther from the principal channels, and the more intense phases of alteration progressively overlapped the less intense phases. Quartz-sericite alteration was superimposed on the zone of clay alteration and even spread along some of the vein fractures out into the propylitic zone.

Following the main period of mineralization, during which quartz, pyrite, molybdenite, and chalcopyrite were deposited along the small vein fractures, a minor structural adjustment occurred in the mineralized area, mainly along the Dome fault system. New channels were opened in which sphalerite and galena were deposited together with a little pyrite and chalcopyrite. Minerals like those of the earlier period were probably deposited at much deeper levels in these same channels at the same time that sphalerite and galena formed nearer the surface at the horizon now exposed. As far as can be determined at the present stage of mine development, the sphalerite and galena mineralization is confined to the Dome fault system and a few major fractures. Barite and fluorite were probably deposited at the same time or closely following the lead-zinc veins.

The wavellite and metatorbernite were clearly later than the sphalerite and galena and also later than the barite and fluorite. They appear to have been deposited by late hydrothermal solutions not necessarily directly related to the earlier mineralization. Wulfenite and libethenite were both later than barite and fluorite but preceded the main period of supergene alteration. Their origin is uncertain, but no substantial field evidence has been found to indicate that they were not deposited by supergene solutions.

SOURCE OF THE HYPOGENE SOLUTIONS

Although the source of the solutions that produced the hypogene mineralization is not known, some speculation as to their possible source is believed justified.

The mineralization was certainly later than the diabase sills and was also later than the diabase intruded around the west, north, and east sides of the quartz monzonite body. It is also later than the granite porphyry, which intrudes both the main diabase and the quartz monzonite; hence it is clear that the solutions did not have their source in the same magma reservoir from which the quartz monzonite was evolved. The association of similar copper metallization with granite porphyry intrusions both at Castle Dome and at Copper Cities, near Sleeping Beauty Peak, about 6 miles to the northeast, and with small granite porphyry intrusions in the Pinal schist in the Cactus property, 2 miles to the southwest, suggests that the solutions may have emanated from the same magma reservoir that supplied the granite porphyry.

The granite porphyry at Castle Dome may be an offshoot of the Schultze granite, but no direct proof of this relationship has been found. The Schultze granite intrudes the quartz monzonite that crops out south of the schist in the southern part of the Castle Dome area and is undoubtedly the same as the porphyritic quartz monzonite in the main mass, but the age of the Schultze granite with respect to the diabase has not been determined. The granite porphyry is very similar to the porphyry that is commonly a border facies of the Schultze granite.

The mineralization is most likely genetically related to the igneous rock most nearly contemporaneous with it, which could be either the granite porphyry or the Schultze granite. Ransome (in Emmons, 1917, p. 213) considered the mineralization in the Globe and Miami areas to be related to the Schultze granite.

AGE OF THE HYPOGENE MINERALIZATION

From the foregoing discussion it is evident that the hypogene mineralization at Castle Dome is later than the diabase and therefore younger than the Pennsylvanian limestones. Ransome (1919, p. 173) has shown that the supergene enrichment in the Miami and Inspiration ore bodies is related to the topography antedating the Whitetail conglomerate and therefore that the hypogene mineralization must have occurred before the Whitetail conglomerate and the dacite were laid down. In the Castle Dome ore body, enrichment is related to the present topography; the fact that no mineralized quartz monzonite has been found in the Whitetail in the Castle Dome area is not conclusive, because only a few fragments of quartz monzonite and only one boulder of granite porphyry have been found in the Whitetail. E. N. Pennebaker reported a single fragment of granite porphyry containing pyrite as having been found in the cuttings of a hole drilled for water in the Whitetail conglomerate northwest of Jewel Hill. Thus, on the basis of the available information, the age of the copper mineralization at Castle Dome cannot be tied down as closely as that in the Old Dominion, Miami, and Inspiration ore bodies. It is, however, probably of the same age; that is, it occurred after the intrusion of the diabase in post-Pennsylvanian time and before the extrusion of the dacite flows. The intrusion of the Schultze granite stock probably took place in late Mesozoic or early Tertiary time and was followed by a period of metallization that produced the large copper ore bodies of the Miami district.

SUPERGENE MINERALIZATION

RELATION TO THE WATER TABLE

All the available information concerning the permanent water table in the Castle Dome area has been obtained from logs of the exploratory churn-drill holes. As the water table is well below the ore body, only a small proportion of the holes are deep enough to reach it. The level at which a hole showed appreciable inflow or the level to which water rose in a hole is taken to represent the permanent water table at that place.

As illustrated on the longitudinal section of plate 4*G*, the water table slopes westward from an approximate elevation of 3,980 feet at coordinate E. 400 to an elevation of 3,800 feet at coordinate W. 3,500. This slope, projected westward, carries the water table to the bottom of Gold Gulch, where it courses along the west side of the quartz monzonite intrusive. On several cross sections where enough holes have been drilled to give reliable information, the water table shows a northward component of dip, suggesting that its general dip is a little north of west. This suggestion is strengthened further by the fact that the level of Gold Gulch where it runs parallel to the south edge of the quartz monzonite is slightly higher than the water level shown in the drill holes to the north. That the bottom of the gulch is approximately at the level of the permanent water table is indicated by seepage and by a slight flow of water that persists throughout most of the year.

Undoubtedly the water table was much higher when the mineralized area was covered by Tertiary deposits and the present drainage system was being developed. The lowering of the water table to its present level was apparently too rapid to affect the shape of the chalcocite body; that is, the water table was not stabilized at any level long enough to permit an accumulation of chalcocite so situated as to suggest a possible relation to it. The irregular bottom of the chalcocite enrichment indicates that the supergene process operated well above the permanent water table, which was probably near its present level during most of the enrichment cycle.

The cycle is not far enough advanced to have caused a recognizable accumulation of chalcocite at the present water level. Every drill hole shows a gradual decrease in the amount of enrichment with depth, and in most of the holes the lower limit of chalcocite is well above the water level. Along the south edge of the ore body several holes drilled from the lower flanks of Porphyry Mountain show enrichment extending down to or a little below the water table, but even in these holes the amount of enrichment decreases with depth and there is no concentration of chalcocite at or near the water table.

The enrichment in the Castle Dome deposit is considered to have taken place in a zone of transient saturation in which, for the most part, the rocks were moist rather than saturated. A condition approaching saturation occurred only after seasons of maximum precipitation, and only then did appreciable downward movement of ground water occur. As concluded by Spencer (1916, p. 80) in his report on the Ely district, oxidation took place mainly when the sulfides were merely moist rather than flooded, because the necessary oxygen could then be supplied to the ground water as fast as it was expended in the reactions of oxidation. After periods of heavy rainfall the products of oxidation were flushed out and carried downward in solution. The depth to which surface water penetrated before its dissolved copper was precipitated by reaction with chalcopyrite depended on the permeability of the rocks, which varied from place to place. Consequently the lower limit of the chalcocite zone is irregular and shows little conformity with either the topography or the water table.

RELATION OF ENRICHMENT TO TOPOGRAPHY

Except for minor irregularities caused mainly by deep leaching along faults, the top of the chalcocite zone shows a striking conformity with the present land surface (pl. 4). The outlines of the ore body on the west, south, and east sides at the various levels are nearly parallel to the contours and reflect even minor irregularities in the topography (pl. 8). The lack of conformity of the bottom of the chalcocite zone with the topography is equally striking.

In the southern part of the mineralized area there are places where enrichment extends down to the water level, but northward there is a progressive decrease in the amount and depth of the enrichment.

The reason for the deep enrichment at the south is not entirely clear, but it is probable that during the development of the present drainage system, the mineralized quartz monzonite was first exposed where Gold Gulch cut through the cover of Tertiary sediments and lavas. Then, as the gulch was cut deeper, the quartz monzonite was progressively uncovered toward the north. Thus the enrichment processes have been active for a longer period of time on the lower flank of Porphyry Mountain than in the more recently uncovered areas to the north.

Gold Gulch is clearly the oldest drainage unit that serves the mineralized area. It has a comparatively low gradient and is probably aggrading slowly at the present time where it courses along the south edge of the quartz monzonite body. At one time it became so stabilized that deposits of coarse gravel accumulated in its channel to a depth of 50 feet or more, and the water table was correspondingly higher than it is at present. Recent entrenchment of the stream left terraces of these old gravels at several places along its course south of the mine. Enrichment probably began at about the time the older gravels were being deposited and the rate of erosion in the areas bordering the gulch had decreased to such an extent that oxidation and leaching progressed more rapidly than mechanical disintegration. That the entrenchment occurred in relatively recent time is shown by the presence of fresh chalcopyrite in the surface rocks exposed by the removal of the gravels.

In contrast, the tributaries draining the area north of Porphyry Mountain have steep gradients and narrow channels in which the bottoms are scoured clean of debris. The rock forming the bottoms of the scoured channels contains fresh sulfides, indicating that, in this part of the area, stream erosion is progressing at a faster rate than complete oxidation and leaching. This observation is confirmed by seven exploratory holes recently drilled in the area north of the waste dumps. The hypogene mineralization cut by these holes averaged about 0.10 percent copper. The average depth to which this small amount of copper has been leached, as indicated by assays of the sludges, is 40 feet; the maximum depth is 70 feet. Only two holes show leaching of copper below the level of the nearest canyon bottom. In one hole, according to the sampler's record, the sulfides were completely leached to a depth of 45 feet, but in all the others pyrite was found within 10 feet of the surface.

PARAGENESIS OF THE SUPERGENE MINERALS

Chalcocite is the most abundant supergene sulfide mineral and accounts for a large part of the copper in the ore body. All but a very insignificant part of it formed as a pseudomorphic replacement of chalcopyrite brought about by reaction with copper sulfate carried in solution by ground water; therefore, the mode of occurrence is the same as that of the replaced mineral, chalcopyrite. The replacement is rarely complete except in very small grains near the top of the chalcocite zone. Elsewhere nearly all grains show remnants of chalcopyrite, and the proportion of unreplaced mineral increases progressively with depth. Replacement begins at the surface or gangue boundaries of the grains and spreads inward toward the centers. Many grains show replacement veinlets spreading from fractures.

In a few places where the copper content of the protore was unusually high—for example, in and near the diabase sills—a little chalcocite formed as a replacement of pyrite, but only after nearly all the available chalcopyrite had been replaced. The pyrite grains show only thin shells of chalcocite, whereas the associated chalcopyrite grains are either completely replaced or contain minute remnants of the original mineral. In most places pyrite is not affected. Sphalerite, though relatively rare, is more susceptible to replacement than chalcopyrite in spite of the fact that it occurs in much larger masses. The sphalerite masses are generally granular, and replacement takes place simultaneously from all the grain boundaries, so that on polished surfaces the chalcocite is dotted with rounded islands of unreplaced sphalerite. Thumbprint structures produced by nearly complete replacement are characteristic of this chalcocite.

The chalcocite that replaced chalcopyrite is of various shades of light blue, whereas that replacing pyrite and sphalerite is generally light gray to white.

Covellite was formed in two ways, either by oxidation of chalcocite or by replacement of chalcopyrite. The amount of covellite formed by the latter process is small compared with the amount of chalcocite, but the mode of occurrence of the two minerals is the same. The relations of covellite and chalcocite to chalcopyrite suggest no clue as to the controls that determined which mineral was formed. It is common to see in the same chalcopyrite grain a chalcocite veinlet in one fracture and covellite in another, both formed under apparently identical conditions.

Most of the covellite was formed by oxidation of chalcocite, probably through reaction with ferric sulfate and sulfuric acid. The greatest concentration of covellite is at the top of the secondary sulfide zone, where the chalcocite may be entirely altered to covellite. In general the chalcocite-covellite ratio increases progressively with depth until only traces of covellite are present. On polished surfaces of the supergene sulfides the covellite formed in this manner appears as a grating of bladed crystals oriented along crystallographic directions of the chalcocite. The alteration proceeds inward from the rims of the grains or along fractures, as in the case of direct replacement, but generally there is some unaltered chalcocite between the covellite and the remnants of chalcopyrite.

Small masses of cuprite were noted in a few places in the oxidized zone. It apparently formed by oxidation of chalcocite in places where insufficient pyrite remained after chalcocite enrichment to provide the necessary acid to carry away the copper. This condition occurred where oxidation was active in and near diabase sills—that is, where the abundant chalcopyrite associated with the sills had been replaced by chalcocite that later came within the zone of oxidation.

The cuprite was further altered by hydration and carbonation to form copper carbonates, and it is almost invariably covered by a shell of malachite and azurite. Malachite is by far the more abundant of the two carbonates. A little azurite may have been formed directly, but in most places its relationship to malachite strongly suggests that it is an alteration product of that mineral. Small amounts of malachite are present throughout the leached capping. Some of it was undoubtedly precipitated from copper sulfate solutions by atmospheric carbon dioxide.

Native copper is exceedingly rare in the Castle Dome deposit. A few small masses have been uncovered in the rich ores along the diabase sills. It is always associated with cuprite and apparently was formed under the same conditions as cuprite. It represents the ultimate product in the oxidation of chalcocite in the absence of excess acid.

Abundant chalcanthite ($CuO.SO_3.5H_2O$) is being deposited in the mine where ground water, supplied by current rains, seeps to the surface and evaporates. Where fracture zones intersect the faces of the benches, the rock has, in little more than a year, taken on a distinct blue color caused by the efflorescent crusts. A little chalcanthite was observed in near-surface fractures when the ore body was being stripped, but most of that seen today has been deposited since mining began.

Molybdenite is usually stable even where the other sulfides have been completely destroyed, but in a few places some of it has oxidized and small quantities of molybdite [$Fe_2O_3.3MoO_3.7\frac{1}{2}(?)H_2O$] have resulted. The derivation of molybdite by oxidation of molybdenite is unquestionable in most specimens.

Galena has altered to the lead sulfate, anglesite, which forms spongy masses in the oxidized veins or white crusts on galena crystals. Sphalerite has oxidized to the zinc sulfate, goslarite, which forms a white coating on the sphalerite or is carried away in solution. Most of the goslarite noted in the mine has formed since the sphalerite was exposed by mining operations. No lead or zinc carbonates have been recognized.

Native sulfur is found wherever sphalerite and galena are undergoing oxidation. It occurs in well-formed orthorhombic crystals in open spaces in the veins and also as thin crusts in fractures in sphalerite.

Though not heavily iron-stained, the leached capping contains a fair amount of residual limonite derived, for the most part, by the oxidation of pyrite and copper sulfides. The term "limonite," as used here, denotes hydrated iron oxides, generally mixtures of goethite and lepidocrocite. Most of it is of the transported type, which forms botryoidal crusts lining the walls of open fractures. In the high-pyrite areas, especially near the top of Porphyry Mountain, the rock is striped by limonite veins in which the botryoidal crusts were deposited on quartz in fractures formerly occupied by quartz-pyrite veins. The limonite was deposited by descending meteoric water, which carried in solution ferrous sulfate and sulfuric acid derived from oxidation of pyrite. The ferrous sulfate was readily oxidized to ferric sulfate, which, with decreasing acidity, hydrolyzed to the basic ferric sulfate and finally to ferric hydrate. The ferric hydrate was precipitated and converted to limonite. Where potash salts were present, the basic ferric sulfate, jarosite, probably formed. Limonite boxwork structures characteristic of leached pyrite-chalcopyrite mixtures can be recognized in many places in the outcrops and in the leached capping. Relief limonite of the type generally considered to result from the oxidation of chalcocite was common in the capping on the 4,430-foot and lower benches.

More or less jarosite ($K_2O.3Fe_2O_3.4SO_3.6H_2O$) is present in all parts of the mine, but it is especially abundant on the upper levels where the sulfides were mainly pyrite. It occurs as bright-yellow powdery masses in the open spaces in limonite-crusted fractures that were once filled with pyrite. Where barite is present in the oxidized zone, it is commonly associated with jarosite, which, together with limonite, forms coatings on and between the barite crystals. The association of jarosite with barite is believed to be coincidental; the barite veins were open and became channels of circulation for sulfate-bearing solutions. Where pyrite is undergoing oxidation at the surface, as in the rock exposed in the canyons north of Porphyry Mountain, jarosite forms abundantly and is later oxidized to limonite after the pyrite is largely destroyed, but in the limonite-crusted fractures in the mine jarosite appears to be the latest mineral and is deposited on the limonite crusts.

Turquoise ($CuO.3Al_2O_3.2P_2O_5.9H_2O$) is fairly widespread throughout the chalcocite zone and in the leached capping. A hard variety of gem quality occurs in small veinlets up to a quarter of an inch thick and also in concretionlike masses up to half an inch thick and several inches across. Its color ranges from sky blue to bluish green, the blue variety predominating. It is generally associated with clay minerals and sericite, which form selvages on the veins and masses and are commonly included in the turquoise. Seen under the microscope, it is a microcrystalline aggregate with an index of refraction of 1.62.

Much more abundant than the hard blue turquoise is a soft, chalky variety that occurs in small masses on the walls of open fractures or fills small fractures. The masses and seams clearly were not formed by alteration or replacement of wall rock, but appear to be accumulations of material transported into the fractures either in solution or in suspension in ground water. The color of this turquoise ranges from almost white to light blue when dry and deeper shades of blue when moist. The powdered mineral in immersion oils is almost identical in appearance with the powdered hard variety and has the same index of refraction. The powder turns brown when ignited, becomes soluble in acid, and gives positive reactions for copper and phosphate ions. All gradations between the hard blue turquoise and the white, chalky variety can be found.

A few specimens of a light-green, iron-bearing copper aluminum phosphate were found occurring in the same manner as the hard blue turquoise. It is a microcrystalline aggregate whose index of refraction is 1.67. Charles Milton, of the United States Geological Survey, identified the mineral as a ferrian turquoise. He reports as follows:

> The light-green mineral is identified as turquoise, rather than chalcosiderite. These two minerals are considered to be isomorphous with Al_2O_3 and Fe_2O_3, respectively. The X-ray pattern is similar to that of turquoise from Lynch, Va., which presumably would hardly be distinguishable from chalcosiderite. However, the density (2.75 by pycnometer) and the indices of refraction indicate turquoise.

Turquoise is generally conceded to be a supergene mineral formed by the action of ground-water solutions, carrying copper and phosphate ions, on kaolin and possibly sericite. Various writers describe all stages in the alteration from copper-stained kaolin to hard blue turquoise, but it is not usually clear from the published descriptions whether the mineral is kaolinite or some other clay. Kaolinite is not a common mineral in the Castle Dome deposit, and its association with turquoise has not been noted; however, turquoise is commonly associated with beidellite, halloysite, and sericite. No evidence has yet been found to suggest that it formed from clay. The hard blue turquoise grades into the soft, white, claylike variety, but the indices of refraction of all of it are near 1.62, which is much higher than that of any of the clays present. None of the material could be construed as being transitional between turquoise and the associated clays.

Although turquoise is most abundant in the upper part of the secondary sulfide zone, it is commonly found in deep fractures where chalcopyrite in the wall rock is only slightly replaced by chalcocite or not at all. If turquoise was formed by supergene solutions, the phosphate ion would have to have been derived from either wavellite or apatite, which are the only earlier phosphate minerals. Wavellite is apparently dissolved by supergene solutions, but much turquoise occurs in parts of the deposit where no wavellite has been recognized. In the Castle-an-Dinas mine in Cornwall (Hey, Bannister, and Russell, 1938, p. 41), wavellite spheres are commonly covered with minute botryoidal aggregates of bright-green turquoise, but no such association has been seen at Castle Dome. Most writers look to apatite as the probable source, even though they admit in some cases that the country rock contains very little apatite.

Although clay is an abundant constituent of the ore body, most of it is of hypogene origin and is a product of the hydrothermal alteration of host-rock minerals, particularly plagioclase. A relatively small amount seems clearly to be supergene and was probably formed by the action of acid-bearing solutions on sericite, feldspar, and perhaps hypogene clay minerals. All the supergene clays have been transported and deposited in fractures or in open spaces in brecciated rock. They commonly engulf earlier minerals, particularly barite, which had formed on the walls of fractures.

The isotropic clay, halloysite, is the most common. It occurs as small seams or masses in the capping and in the upper part of the secondary sulfide zone. These masses, commonly several inches thick, are white to creamy white and break with conchoidal fracture. Endellite occurs in exactly the same manner and cannot be distinguished from halloysite except by a difference in the index of refraction.

Some of the halloysite contains very finely divided particles of a zinc mineral like the tallow clay described by Short (1943, p. 104) as occurring in the Magma mine. J. J. Fahey, of the United States Geological Survey, examined the material for Short and suggested that the zinc mineral is probably hemimorphite. The inclusions in the Castle Dome clay are very small, rounded grains whose index of refraction is near 1.62, much higher than that of the clay. They are probably hemimorphite.

Another isotropic clay with an index of refraction near 1.515 occurs in the same manner as halloysite and endellite. Only a few specimens have been found. Its optical properties are the same as those of cimolite, which Larsen and Berman (1934, p. 50) list as a doubtful clay mineral.

Some kaolinite is undoubtedly present in the oxidized zone, but no certain identification has been made because all specimens in which its presence was suspected were contaminated by limonite and sericite. Some of the mineral believed to be kaolinite is rendered almost opaque by fine, dustlike inclusions that are probably limonite.

The claylike hydrous ferric silicate, canbyite, is abundant throughout the oxidized and leached zone. It is chocolate brown in color and has a waxy luster. It coats the walls of fractures and has the appearance of a dried slimy mud. Under the microscope it is seen to consist of very thin, amber-yellow flakes. Its refractive indices vary somewhat but are near α, 1.56; γ, 1.58. Its optical character is negative, and the optic axial angle ($2V$) is very small.

In a few places very thin crusts of quartz crystals, probably deposited by supergene solutions, coat barite and rock minerals on the walls of open fractures. Small amounts of opal and chalcedony were noted in a few places in the leached capping.

According to Ransome (1903, p. 160), the oxidized ores in the Continental mine, 3,000 feet northeast of Porphyry Mountain, contained a little native silver. It occurred chiefly in calcite tinged green and blue by carbonates of copper. None was recognized in the Castle Dome deposit.

DETAILS OF THE ENRICHMENT PROCESS

Conditions in the Castle Dome area would seem to favor deep oxidation, leaching, and transportation of copper. The water table is relatively deep, and although the rock is not thoroughly brecciated, joints and minor fractures are abundant throughout the mineralized area. The gangue and host rock are relatively nonreactive, and the amount of pyrite present appears ample to provide sufficient acid to permit transportation of the copper set free by oxidation of the hypogene sulfides. Although the present climate is semiarid, the rainfall is comparable with that of other regions in the Southwest where oxidation and enrichment have been deep. In the nearby Miami and Inspiration deposits, oxidation and leaching commonly reach a depth of 600 feet below the top of the mineralized schist and porphyry; the average thickness of the leached capping over the ore bodies is probably more than 250 feet. In places the hypogene sulfides have been partly or wholly replaced by chalcocite for 300 to 400 feet below the bottom of the leached capping. Nevertheless, the effects of supergene alteration in the Castle Dome deposit are superficial, and the cycle is not far advanced.

The thickness of leached capping over the Castle Dome ore body ranges from a few feet to a maximum of about 250 feet; the average thickness is about 80 feet. In some places along the south side of the ore body, oxidation penetrates to the water level, but rarely does complete leaching extend more than 150 feet below the surface. The leached capping generally contains less than 0.1 percent copper, and the transition from waste to ore containing 0.4 percent or more copper usually occurs in less than 15 feet.

The importance of supergene enrichment is illustrated graphically by plates 6*B* and 7*B*, which are typical cross sections through the ore body. The copper graphs plotted along the drill holes show the approximate amount of copper originally present in the protore and also the amount of copper added by supergene enrichment. They clearly show that the supergene copper amounts to a substantial increase in the tenor of the ore, especially in the southern part of the ore body.

The graphs were prepared from data obtained by estimating the amount of supergene replacement of sulfide minerals observed in polished briquettes of concentrate panned from churn-drill sludges. Dried sludge samples of about 850 grams each, representing 5 feet of churn-drill hole, were panned, and the concentrate was cleaned and weighed to determine the percent of total sulfide minerals. The concentrate was mounted in bakelite and polished for microscopic examination with reflected light. Visual estimates were made of the proportion of pyrite, the proportion of chalcopyrite replaced by chalcocite and covellite, and the ratio of chalcocite to covellite. From the proportion of pyrite and the amount of total sulfide recovered, the percent of pyrite in the sample was determined. Some of the sulfide was unquestionably lost in panning; this figure is therefore lower than the true value for each sample. However, assuming that the proportion lost is approximately constant, the graphs plotted on the left sides of the drill holes show the relative distribution of pyrite with reasonable accuracy.

The relative proportion of hypogene and supergene copper, plotted on the right-hand side of each drill hole, was determined by the formula

$$X=\frac{100C}{AK+B}$$

in which X is the weight of copper present as chalcopyrite before enrichment—that is, the weight of copper originally present in the protore; A is the estimated percentage of chalcopyrite replaced by chalcocite and covellite; B is the percentage not replaced; C is the weight of sulfide copper determined by analysis of the sludge sample; and K is a factor representing the increase in copper content in the conversion of chalcopyrite to chalcocite and covellite. This factor varies according to the ratio of chalcocite to covellite in the replaced mineral.

These computations are based on the assumption that all the chalcocite and covellite formed as a replacement of chalcopyrite, which is essentially true. Only two briquettes showed slight replacement of pyrite. The very small amount of sphalerite and galena present in some of the briquettes was ignored in the computations. The data used in plotting the graphs were obtained by studying every fifth sample in each hole. Each sample represents the sludge from 5 linear feet of hole; hence the samples studied represent 5-foot segments 20 feet apart vertically. The data are necessarily approximate because of incomplete recovery of the sulfides by panning and because of the uncertainty of estimating the volume of mineral replaced from the surfaces exposed in the polished briquettes. Another minor source of error is the small amount of copper in the form of oxidized minerals shown on a few of the graphs. As the oxidized minerals probably formed largely from chalcocite or covellite, a part of their copper content may have been added by supergene enrichment. However, in spite of these minor inaccuracies, the graphs illustrate the relative magnitude of the effects of supergene enrichment on the hypogene mineralization.

The graph sections show the relatively deep oxidation and enrichment in the southern part of the mineralized area and the decrease in the amount and depth of enrichment farther north. Although it is doubtful that more than a few hundred feet of quartz monzonite have been eroded from the top of Porphyry Mountain since the cover of Tertiary rock was removed, considerably more erosion undoubtedly took place on both the north and south flanks; during the early stages, erosion of the quartz monzonite probably progressed more rapidly than oxidation and leaching, so that no enrichment occurred. Later, when Gold Gulch ceased cutting deeper and its channel began to aggrade, the rate of erosion on the south flank decreased and the condition was reversed; that is, oxidation and leaching penetrated faster than the quartz monzonite was worn away. For a time erosion may have nearly kept pace with the leaching of copper.

At least in the southern part of the ore body considerably more copper has been added by enrichment than can be accounted for by leaching of the present capping. Much of this excess copper was leached from capping subsequently removed by erosion, but there has undoubtedly been some southward migration of ground water, roughly parallel to the slope of the surface toward Gold Gulch, and a part of the excess supergene copper present in the lower southern part of the ore body was probably leached from higher ground farther to the north.

The large dump-covered area on the north side of Porphyry Mountain could not be examined, and it contains only four exploratory drill holes, W10S, W16W, W27Q, and W29U (pl. 1). All four holes show a relatively high pyrite content and are apparently in the pyrite zone. They show leaching of copper to a depth of 50 to 100 feet. Hole W27Q contains 85 feet of ore averaging 0.58 percent copper, but the other three show only slight chalcocite enrichment. No information is available concerning the area between coordinate line *W* and the north edge of the dump. In the remainder of the mineralized area north of the dump, fresh sulfides are exposed in the bottoms of all the canyons, indicating that there has been too little leaching and enrichment to raise the protore from 0.10 percent copper to ore grade. Practically no chalcocite and no residual limonite indicative of leached chalcocite were found associated with the sulfides exposed in the outcrops. The results of recent exploratory drilling in the area north of the dump are described on pages 100 and 110.

The oxidation of the primary sulfides and the leaching, transportation, and redeposition of copper formed a continuous and progressive process. The irregularity of the bottom of the chalcocite zone and the progressive decrease in the amount of chalcocite with depth suggest that the process took place above the permanent water table. The fact that a relatively large proportion of the chalcopyrite remains unreplaced throughout the secondary sulfide zone indicates that the process is not far advanced. The supergene cycle began after the mineralized quartz monzonite was uncovered and exposed to weathering and after the rate of erosion had become stabilized to a degree that mechanical disintegration progressed more slowly than chemical processes.

The first step in the cycle began with near-surface oxidation of pyrite and chalcopyrite, forming copper sulfate, sulfuric acid, and ferrous sulfate that were carried downward in solution by ground water. When the acid became largely neutralized by reaction with rock minerals, the copper sulfate reacted with chalcopyrite, forming chalcocite, which was fixed as a pseudomorphic replacement of chalcopyrite. Undoubtedly a little covellite also was formed by direct replacement of chalcopyrite. Ferrous sulfate was quickly oxidized to ferric sulfate, which, in the presence of sulfuric acid, is an active solvent of sulfide minerals. Where an excess of free acid was not available, ferric sulfate was hydrolyzed to ferric hydrate, which soon became fixed as limonite.

As oxidation progressed deeper, it encroached upon the newly deposited chalcocite zone, where chalcocite then became the first mineral to be attacked, followed by chalcopyrite. Pyrite, being the most resistant, remained to some extent in the partly leached rock but finally oxidized, yielding sulfuric acid, ferrous sulfate, and eventually ferric sulfate. Where the latter, carried downward in acid solution, came in contact with chalcocite, it altered the chalcocite to covellite by removing half the copper. However, much of the ferric sulfate was hydrolyzed, and its iron was deposited in fractures as the botryoidal crusts of limonite that are so abundant in the Castle Dome deposit. Covellite is much more resistant to solution than chalcocite and probably more so than chalcopyrite. It is well known in connection with the metallurgical leaching treatment of chalcocite ores, not only in processes employing ferric sulfate but also in those in which oxygenated water is used as a solvent, as at Río Tinto, Spain, that cupric sulfide forms as an intermediate product in the decomposition of chalcocite. The half of the copper freed by the alteration of chalcocite to cupric sulfide is quickly and easily leached; the remaining half is recovered only by prolonged treatment unless heat is applied, in which case the reaction may be accelerated. When water and atmospheric oxygen only are used on coarsely crushed material, the first reaction may be accomplished in a few months, whereas the second step usually requires several years.

In order to determine whether the cupric sulfide produced in the metallurgical process is actually the mineral covellite, B. H. McLeod, of the Inspiration Consolidated Copper Co., leached a sample of chalcocite (minus 150-mesh, plus 200-mesh) with a 0.05 percent solution of ferric sulfate (pH, 1.2) until the first step in the reaction was complete. The residue was then imbedded in bakelite, polished, and examined microscopically by means of reflected light. The leached grains were strongly anisotropic and had other optical properties characteristic of covellite. J. M. Axelrod, of the Geological Survey, examined the chalcocite and the leached product by means of X-rays and reported as follows:

> The specimen labeled chalcocite is chalcocite by comparison with data for a chalcocite from Cornwall, England. Its density after decantation with bromoform is 5.44 (Dana: 5.5–5.8). The specimen labeled covellite (?) gave a pattern corresponding to very fine-grained and poorly crystallized covellite. It contained quartz impurity. Its density after decantation with bromoform is 4.20 (Dana: 4.6).

The leached grains appeared to be homogeneous and showed no evidence of increased porosity. As approximately half the copper was removed by the leaching process, the alteration did not take place by volume-for-volume replacement. The grains apparently contracted in size, for had the alteration taken place by a simple removal of half the copper, the density of the product should have decreased to about 3.36.

Thus a horizon rich in covellite was formed in the upper part of the chalcocite zone. The effect of this process is illustrated on plates 6*A* and 7*A*, which show a greater proportion of covellite in the upper part of the drill holes just below the leached capping. Although a little of the covellite formed as a direct replacement of chalcopyrite, no evidence was seen in the polished briquettes that it is a transitional mineral in the replacement of chalcopyrite by chalcocite.

As the cycle continued, the proportion of chalcocite to chalcopyrite in the secondary sulfide zone increased, and the copper-bearing solutions descended deeper and deeper before coming in contact with sufficient chalcopyrite to precipitate all their copper. Eventually a stage was reached in which no more chalcopyrite remained unreplaced and the copper-bearing solutions began to attack pyrite.

The study of the supergene mineralization in the Castle Dome mine shows clearly that pyrite is not replaced until after nearly all the chalcopyrite has been used up—that is, until only very small kernels insulated by thick shells of chalcocite are left. This selective action of the replacement can probably be explained by a slight difference in potential between chalcopyrite and pyrite that causes one mineral to be attacked and exerts a protective influence on the other. Only two of the samples studied showed slight replacement of pyrite by chalcocite.

One of these samples, which contained 4.4 percent copper, is from a diabase sill in the upper part of drill hole W29D (pl. 7). The chalcopyrite was 95-percent replaced by chalcocite, and the pyrite grains, which amounted to less than 5 percent of the total sulfides, showed slight replacement. This sample is not typical of the deposit in that it contained such a high concentration of copper and so little pyrite. There are probably very few places in the deposit where chalcocite replacement is as complete as is represented by this sample. Although the bottom of the leached zone is only a short distance above the diabase sill, the sample contains no covellite, indicating that it has not yet been affected by oxidation, whereas the sample 25 feet higher in the hole shows a chalcocite-to-covellite ratio of 1:4, proving that in it oxidation of chalcocite had begun. If the zone of oxidation continued to descend and reached the rich chalcocite ore in the diabase, there would probably be too little pyrite present to produce enough acid to carry away the copper. Consequently the chalcocite would be converted directly to oxide and carbonate with little or no movement of copper.

An example illustrating this condition is shown in the diabase sill near the top of drill hole W18D (pl. 6*A*). The sample in the sill contained 1.37 percent copper, of which 1.24 percent was in the form of oxide and carbonate. Only a trace of pyrite remained. Thus, if oxidation and leaching continued, the copper oxide would eventually be converted entirely to carbonates and remain as such in the capping.

The most common occurrence of oxidized copper minerals is in and near diabase sills that have been exposed by stripping operations. In most places in the mineralized quartz monzonite there was no such deficiency of pyrite. Generally the leached capping contains less than 0.1 percent of copper, so that even if all of it is in the form of oxidized minerals, the amount of such minerals remaining is very small.

Galena is present in such small amounts that no examples of its replacement by supergene copper minerals were seen. It was apparently converted to anglesite and remained in the capping. With the probable exception of galena, sphalerite was the least stable sulfide mineral in the chalcocite zone and, wherever present, is usually completely replaced by chalcocite. The liberated zinc was carried away and dissipated. No secondary zinc sulfide has been recognized in the Castle Dome deposit.

Molybdenite was the most stable sulfide mineral in both the oxidized and chalcocite zones. Although it is rarely recognized in the outcrop, there has probably been no leaching of molybdenum. Molybdenite may have been partly oxidized within a foot or two of the surface, but below that depth it appears to have been unaffected. Only in a very few places was any alteration to molybdite noted.

AGE OF THE SUPERGENE ENRICHMENT

Ransome (1919, p. 173) has shown that the supergene enrichment in the Miami and Inspiration deposits is not related to the present topography or to the present water table, and he concludes that it occurred before the dacite was poured out over the surface. In contrast, the close conformity of the top of the chalcocite zone in the Castle Dome deposit with the topography clearly relates the enrichment to the present erosion cycle. The shallow depth to which oxidation and enrichment have penetrated in most parts of the deposit and the incomplete replacement of the hypogene sulfides are proof that the process is an early stage of development. If any enrichment was accomplished in the period of exposure, prior to the time the dacite was laid down, all evidence of it has been erased by recent erosion. It appears more probable that at the time the Miami and Inspiration deposits were undergoing enrichment, the mineralized quartz monzonite at Castle Dome was part of a low-lying area in which the water table was too near the surface to permit appreciable oxidation and leaching. It was probably a basin of deposition, receiving debris washed in from adjacent higher areas as discussed on page 40. The chalcocite enrichment now seen in the ore body began after the dacite and Gila conglomerate were laid down and largely removed from the mineralized area by erosion. It was still in progress when interrupted by the present mining operations.

CASTLE DOME ORE BODY

GENERAL DESCRIPTION AND TENOR

The ore body that is being exploited in the Castle Dome mine underlies the south flank of Porphyry Mountain (pl. 9) between coordinate lines *D* and *K*. Measured on the long axis, which strikes N. 75° E., it ranges in length from 2,000 feet on the 4,430-foot level to about 3,800 feet on the 4,085-foot level. Its upper surface, which in most places is the boundary between leached capping and the top of the secondary sulfide zone, dips from 25° to 45° S., which is a little steeper than the slope of the mountain. The north and lower limits of the ore body, as shown on the plans and cross sections, are assay boundaries determined by the prevailing cut-off grade. The vertical thickness, though not uniform, is estimated to average about 225 feet.

Considered from a purely practical point of view, only such material as is included within the scope of mining operations determined by the original plan of exploitation can be rightly considered ore, but for purposes of geologic discussion the term "ore body" will be used to include all the rock containing 0.4 percent or more copper.

The shape of the ore body is difficult to describe in detail. It can best be visualized by referring to the bench plans (pl. 8) and the cross sections (pl. 4). These show that the outlines and cross sections of the ore body are irregular and differ widely in size and shape from place to place. In detail the shape is influenced by the diabase sills and to some extent by faults. Most of the irregularities in the upper surface of the ore were caused by faults, in some cases by deep leaching along them, in other cases by the diversion of ground water so as to insulate certain parts from the effects of supergene enrichment.

The lower limit of the ore body is irregular and is difficult to outline accurately because, in most places, the exploratory drill holes are too far apart to permit the projection of apparent boundaries from one hole to another. The same handicap is met in attempts to work out the internal structure of the ore body. The drill logs consistently show segments of relatively high-grade ore alternating with segments of low-grade ore or waste. Because of the wide spacing of the holes an intersection of ore in one hole cannot be safely correlated with an intersection of similar grade in another hole without some additional evidence of continuity to justify the projection. These variations in grade of ore may be due to any one of several causes, and the manner in which they are interpreted and shown on the cross sections will depend on the cause. Since the cause cannot always be determined from a study of the drill log, it is often impossible to select the proper criterion for interpretation.

The variations in the grade of the ore may be due to differences in the amount of hypogene mineralization governed by differences in permeability, which in turn are governed by the number and spacing of vein fractures. In such cases, the trend of the bands of different grades of ore should be determined by the strike and dip of the vein system, which would give a steeply southward-dipping pattern on the cross sections.

Where the variation is related to diabase sills, the interpretation must be made on the basis of the strike and dip of the sills; that is, the line of projection from one drill hole to another would be nearly horizontal or dip gently toward the north parallel to the sills.

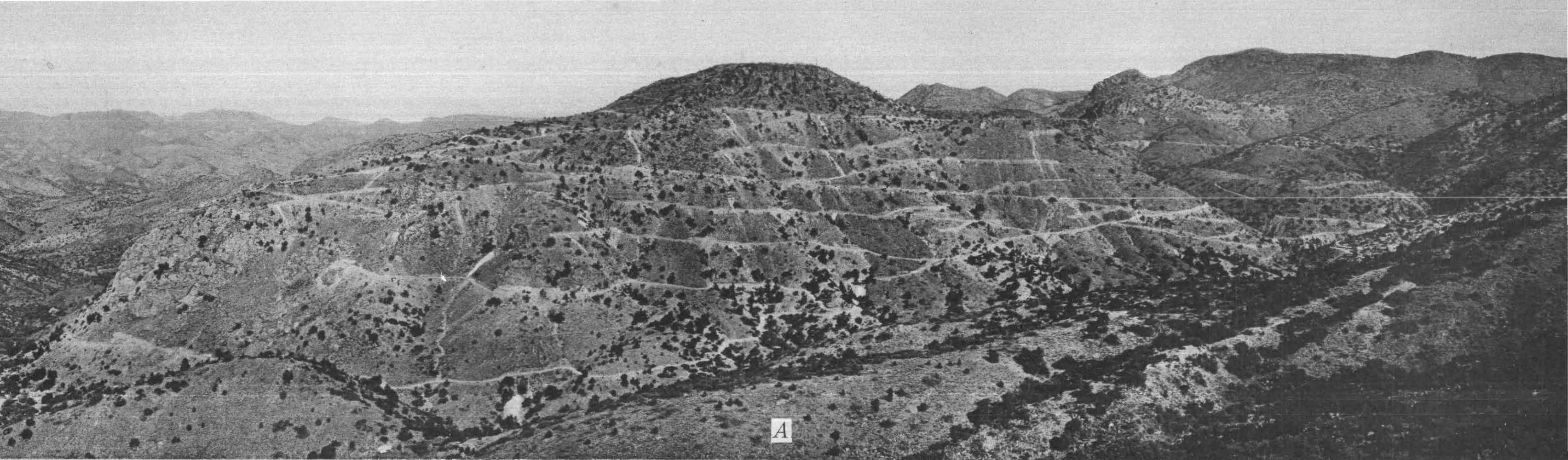
Some low-grade intersections are related to major faults. These are caused either by deep oxidation and leaching along fault zones or by the insulating action of gouge-filled fractures in protecting blocks of protore against both oxidation and enrichment. Features of this type are the most easily recognized and interpreted provided the fault to which they are related can be identified on the surface so as to determine its strike and dip.

Variations in tenor due to supergene enrichment are difficult to identify without detailed knowledge of the mineralogy of the ore involved unless they are clearly related to the lower surface of the leached capping. The interpretation of deeper isolated intersections of chalcocite-rich ore is difficult or impossible, because the structures that guided the course followed by descending ground water cannot be determined by a study of the drill logs.

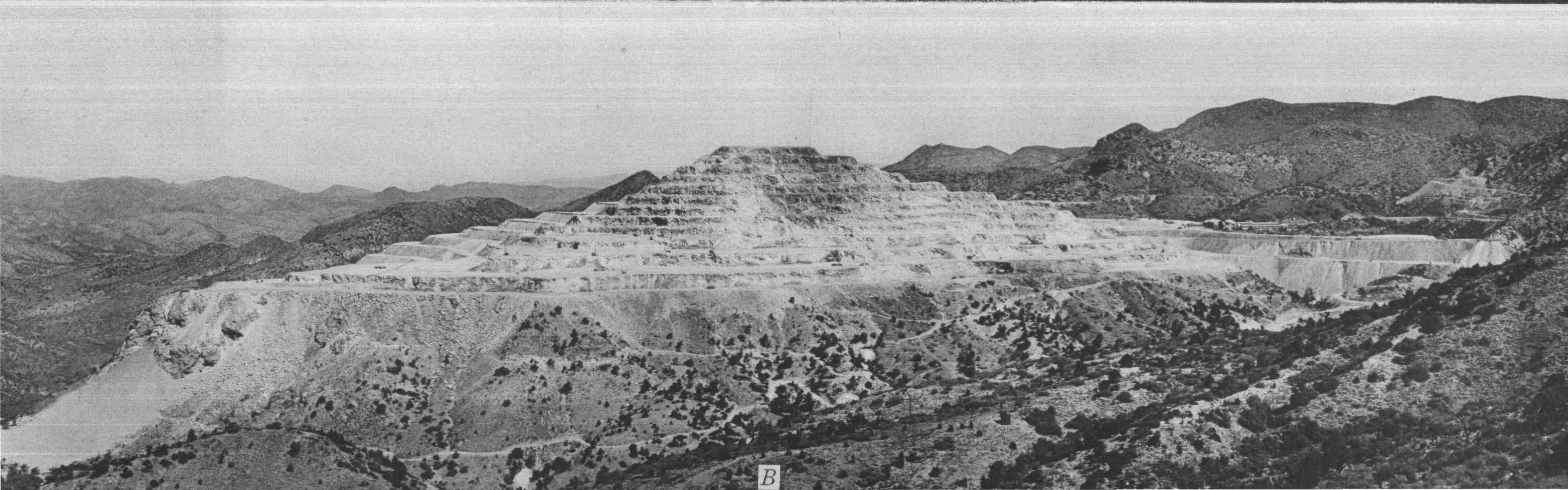
EXPLANATION OF PLATE 9

A. Porphyry Mountain, Castle Dome area, Gila County, Ariz., before mining operations were started. View from the south.

B. Porphyry Mountain, seen from the south after the mine was developed.



A



B

Thus, in order to obtain a true picture of the detailed structure of the ore body, considerably more exploratory drilling would be necessary. However, the general structure of the ore body within the mine limits is fairly well known. The region of greatest uncertainty lies beyond the mine limits, where lack of data is not a practical handicap to the present plan of operation.

The nearly barren capping is easily distinguishable from the underlying light greenish-gray ore by the red or brown coloration caused by residual iron oxides. Below the oxidized and leached zone, the margin of difference between all but the most high-grade ore and rock below cut-off grade is so small that it is usually impossible to distinguish ore from waste by eye. The control for selective mining requires close sampling, which is accomplished by cutting samples from the sludges of every blast hole. In the present operation the mill head is maintained at approximately 0.7 percent sulfide copper by discarding all rock below the 0.4 percent cut-off. A little copper in the form of oxide and carbonate, not recoverable by flotation, is present throughout most of the ore body. The amount differs, depending mainly on the depth below the surface and the nearness to faults and fractures along which there has been deep oxidation. For the ore body as a whole it amounts to 3 to 5 percent of the total copper—that is, about 0.02 to 0.03 percent of the ore.

The structure in the mineralized area is described in detail on pages 54–61. Very few structural features had any apparent effect on the shape or tenor of the ore body. The Dome fault system cuts northeastward through the ore body, dividing it into two parts separated by irregular bodies of waste consisting of leached or unenriched protore related to the fault zone. The amount of displacement on this fault is not known, but from the apparent offset of the diabase sill shown on longitudinal section *G* (pl. 4*G*) a normal displacement of 300 feet to 400 feet might be inferred. However, there is no direct evidence that the east and west segments of the sill were ever parts of a continuous sheet or that the fissures in which the two segments occur were continuous before faulting. At least a part of the movement along the Dome fault system was earlier than the diabase and therefore earlier than the mineralization, because diabase was intruded along the fault in a number of places. The chalcopyrite zone was undoubtedly displaced by movement along the fault after mineralization took place, but the displacement was not sufficient to affect appreciably the continuity of the ore body. It is obvious that because the ore body is to a large extent the result of secondary enrichment, an offset in the protore would have little if any effect on the present continuity of the ore.

Evidence of faulting on a minor scale both before and after the diabase intrusion is seen wherever sills have been exposed by mining operations. Some offsets in the sills are clearly due to minor faulting after the diabase intrusion, whereas others were caused prior to the intrusion by faults that offset the gently dipping fissures into which the sill was later intruded. Evidence of offsets produced by faults formed prior to the diabase intrusion does not commonly appear on the maps, but two such offsets where the diabase was continuous along a fault formed prior to the diabase intrusion and later slightly displaced along the same line are shown on plate 3 at coordinate positions W. 2450, C+60, and W. 1460, M+15. Undoubtedly the sills are not as regular as the mine sections imply; in detail they probably have numerous minor offsets and irregularities. Only their general trend can be determined from the recorded drill-hole intersections.

GEOLOGIC CONTROLS THAT PRODUCED THE ORE BODY

The localization of copper minerals that produced the ore body was due to three major causes: (1) zoning in the hypogene mineralization, (2) the richer-than-average copper metallization associated with the diabase sills, and (3) supergene enrichment. Although the ore body was formed where the maximum effects of all three controls are superimposed, it should not be implied that all three were necessary to produce metallization of ore grade. However, no single one or even two of these controls were important enough in their effects to produce metallization of sufficient extent and value to constitute an ore body that can be exploited under present economic conditions.

The difference in tenor of the protore that can be ascribed to zoning varies from about 0.1 percent copper in the pyrite zone north of the ore body to about 0.3 percent copper in the chalcopyrite zone at the south in the vicinity of the ore body. Thus zoning alone did not produce ore but did provide a body of protore that could be raised to ore grade by a little supergene enrichment, whereas only under the most ideal conditions could the copper in the pyrite zone be concentrated sufficiently to produce an ore body. The information available indicates that leaching was shallow and enrichment slight in the pyrite zone.

The greater concentration of hypogene copper in and near the diabase sills produced ore where the sills are within the chalcopyrite zone, but the concentration is less pronounced in the pyrite zone, as is illustrated by the few drill holes in the northern area. In places in the upper part of the ore body, the original copper content was more than doubled by supergene enrichment, and the relatively rich protore associated with the diabase sills was enriched proportionally.

Each of these three controls affected a much greater volume of rock than is included within the ore body. The entire ore body is included within the chalcopyrite zone, but only a small part of the chalcopyrite zone is ore. The chalcopyrite zone probably continues to considerable depth below the surface, as is shown by drill hole PV-16 at position W. 600, I (pl. 4*B*), which averaged 0.25 percent copper to a depth of 600 feet below the bottom of the ore body. Likewise, the increased copper metallization along the diabase sills continues northward beyond the limits of the ore body, but the dip of the sills carries them below the limits of supergene enrichment north of the mine, so that beyond a certain limit the effects of the sills and of enrichment are not supplementary and therefore do not produce ore in sufficient quantity to warrant development by stripping.

Thus the ore body was formed by a combination of favorable circumstances. The first was the result of primary zoning, which caused a concentration of copper in a relatively small part of the mineralized area—that is, the chalcopyrite zone. This in itself would have given the rock a copper content of only about 0.3 percent. However, in and near the diabase sills where they cut through the chalcopyrite zone, the hypogene metallization was at least twice as rich in copper as in the rest of the chalcopyrite zone. When erosion cut through the overlying younger rocks, it exposed the chalcopyrite zone, so that the diabase sills crop out along its southern limit; not only was all the richer protore associated with the diabase preserved, but the southern part of it was brought within the range of supergene enrichment. The mineralized quartz monzonite was probably first exposed where Gold Gulch cut through the younger rocks; therefore supergene processes have been active for a longer period of time in the vicinity of Gold Gulch than elsewhere in the area.

The topography was carved by erosion in such a way that the maximum amount of supergene enrichment occurred in the chalcopyrite zone so as to produce a chalcocite blanket directly overlying the rich hypogene ore associated with the diabase sills. If erosion had been a little deeper, the rich hypogene ore associated with the diabase would have been carried away and lost. If it had been shallower, the ore body would probably have been diluted by a layer of low-grade rock separating the supergene ore from the rich ore in the diabase. If Gold Gulch had not established itself exactly where it did, supergene processes might not have operated effectively enough to form an enriched ore body worth developing under existing conditions.

CONTINENTAL ORE BODY

The Continental mine is 1,600 feet north of the Castle Dome shops and offices. It is in a small finger of quartz monzonite projecting from the east margin of the main body. The finger is bounded on the north, east, and south by diabase. Ransome (1903, p. 159), who examined the property in 1901, describes it as follows:

> The workings consist of three tunnels with several hundred feet of drifts and crosscuts, two levels, the fourth and fifth, below the third level tunnel, which is the main adit, and some small shafts. The total depth reached is about 350 feet.
>
> * * * Several structurally important faults converge at the Continental mine, and, as is usually the case, the rocks in the vicinity of the intersection are not only disturbed by the main faults but by many minor dislocations as well.
>
> The faults are apparently all normal, the diabase having been dropped against the granite. They probably antedate the period of mineralization in the main, but all show evidence of some recent slipping.
>
> The main adit tunnel runs north for about 215 feet in diabase, and then passes through a fault plane into a triangular block of granite-porphyry, within which are situated the main workings and the ore bodies. So far as known, the ore is confined to this small mass of porphyry and a smaller block lying between it and the Ninety-six shaft * * *, while the faults bounding these blocks, and the surrounding diabase, limestone, and granite-porphyry, contain, so far as known, no mineralization of economic importance. The ore occurs in connection with minor fissures within these two bodies of porphyry, particularly within the larger one, and not in the stronger faults that inclose the latter.
>
> The main vein * * * has a curved course ranging from northeast at its eastern end to nearly northwest at its western end. Below a depth of about 100 feet from the surface the ore occurs as a bunchy vein of quartz, pyrite, and a little chalcopyrite, passing into mineralized porphyry without sharply defined walls, and showing considerable recent movement along the fissure. This sulfide ore is low grade, carrying from 2 to 3 per cent up to an occasional tenor of 20 per cent copper.
>
> For a distance of about 100 feet down from the surface the vein is oxidized and contains some small bodies of rich ore, consisting of cuprite, malachite, and azurite, with native silver. The latter occurs chiefly in calcite, tinged green and blue by carbonates of copper, which often form the gangue of the cuprite.
>
> It is not known whether any chalcocite or other cupriferous sulfides occur between the oxidized zone and the low grade pyrite ore below. None was seen at the time of visit.
>
> The occurrence of goslarite, a hydrous sulfate of zinc, as a fluffy efflorescence of acicular crystals coating some of the drifts, was noted as a rather peculiar feature, as no zinc-bearing minerals were seen in the ore.

No ore had yet been shipped at the time of Ransome's visit, but the deposit was probably developed to a depth below the limit of commercial ore. The underground workings were not examined during the present study of the area.

The earliest recorded production was in 1906 and 1908, when 14,300 tons of ore was mined. The ore produced from 1906 to 1936 contained

5.2 percent copper, 0.058 ounce of gold, and 3.4 ounces of silver per ton. Although the level from which this ore was mined is unknown, it appears probable that it represented an enriched zone where the copper content was highest. Ransome saw no secondary sulfides at the time of his visit; his description of the ore strongly suggests an oxidized chalcocite zone.

After 1936 there was an abrupt change in the tenor of the ore mined. The copper content fell to an average of 0.74 percent; gold increased to 0.15 ounce, and silver to 4.45 ounces per ton. The deeper levels of the mine were probably inaccessible by this time, and the later production doubtless was mined above the adit level where much of the copper had been leached. Some increase in gold content would be expected in the leached zone, and inasmuch as the value of gold increased to $35 per ounce near the beginning of this later period, an effort was probably made to mine ore with as high a gold content as possible. The increased tenor in silver also suggests that the later production came from the higher levels, for although there was probably some solution and transportation of silver, the secondary silver minerals are generally precipitated nearer the surface than secondary copper minerals.

The Continental ore body is probably related to the same period of metallization that produced the Castle Dome ore body. Although it differs in tenor, the hypogene mineralization in the Continental body is essentially the same as that of the Castle Dome deposit.

ALPHABETICAL LIST OF MINERALS OF THE CASTLE DOME DEPOSIT

Adularia
Allophane
Anglesite
Apatite
Azurite
Barite
Beidellite
Biotite
Bleached biotite
Calcite
Canbyite
Chalcanthite
Chalcedony
Chalcocite
Chalcopyrite
Chlorite
Cimolite
Clinozoisite
Copper (native)
Covellite
Cuprite
Endellite
Epidote
Fluorite
Galena
Gold
Goslarite
Gypsum
Halloysite
Hydrous mica
Ilmenite
Jarosite
Kaolinite
Leucoxene
Libethenite
Limonite
Magnetite
Malachite
Metatorbernite
Molybdenite
Molybdite
Oligoclase
Opal
Orthoclase
Pyrite
Quartz
Rutile
Sericite
Silver (native)
Specularite
Sphalerite
Sphene
Stilbite
Sulfur (native)
Turquoise
Wavellite
Wulfenite
Zircon

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INDEX