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

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

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

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

Partial stratigraphic section, Verde formation

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

Partial stratigraphic section, Verde formation—Continued

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

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

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

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

Age

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

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

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

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

OLDER QUATERNARY GRAVELS

Distribution, thickness, and stratigraphic relationship

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

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

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

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

Lithology

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

YOUNGER QUATERNARY GRAVELS

Distribution

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

Thickness and stratigraphic relationship

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

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

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

Lithology

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

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

RECENT RIVERWASH, TERRACE DEPOSITS, AND TALUS

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

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

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

STRUCTURE OF THE OLDER PRECAMBRIAN ROCKS

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

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

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

STRUCTURE OF THE ASH CREEK GROUP

The structure of the Ash Creek group is dominated by two major folds: the Tex syncline and the Mingus anticline (fig. 3). The Tex syncline, in the southwestern part of the area, strikes northwest, whereas the Mingus anticline to the northeast and partly covered by younger rocks, has a sigmoid pattern, owing to deformation after folding.

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

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

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

TEX SYNCLINE

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

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

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

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

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

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

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

Incorrect spelling on the Mingus Mountain quadrangle.

MINGUS ANTICLINE

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

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

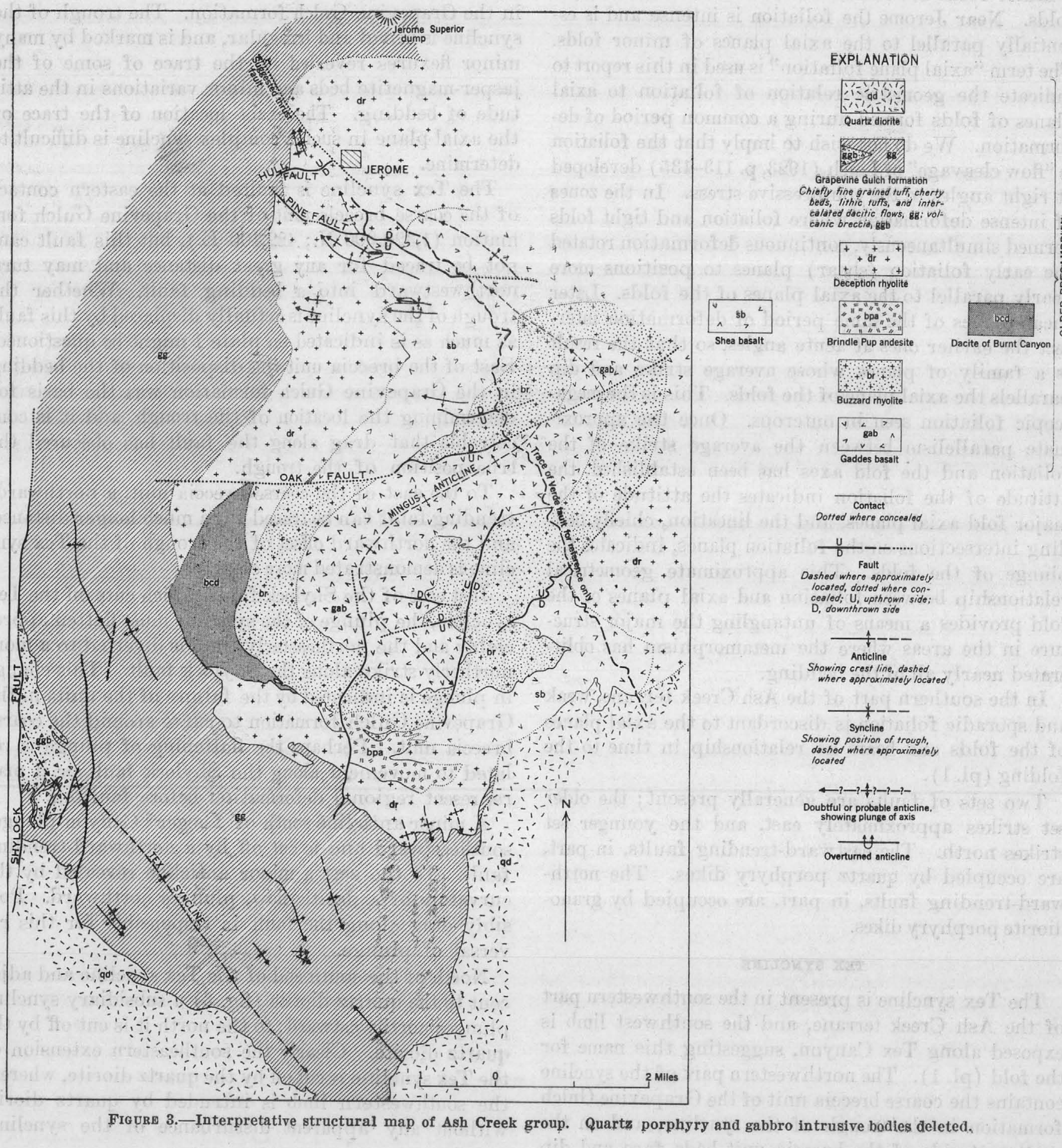

FIGURE 3.—Interpretative structural map of Ash Creek group. Quartz porphyry and gabbro intrusive bodies deleted.

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

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

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

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

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

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

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

Additional evidence for the interpretation of a warped anticline is found in the trend of the contacts on the southern limb of the Mingus anticline westward from Black Canyon. The Buzzard rhyolite-dacite of Burnt Canyon contact trends northwestward near 1,333,000 N.; 434,000 E. in Burnt Canyon where it changes strike to north-northeast. In similar manner, the contact between the Grapevine Gulch formation and older volcanic formations trends west-northwestward in the area south of Black Canyon, but near 1,334,000 N.; 429,000 E. the contact changes strike from northwest to northeast. Both of these areas where the trends of formations change from northwest to northeast are in the correct spatial position to be on the axis of the crossfold that warped the Mingus anticline.

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

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

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

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

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

JEROME-SHEA BLOCK

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

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

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

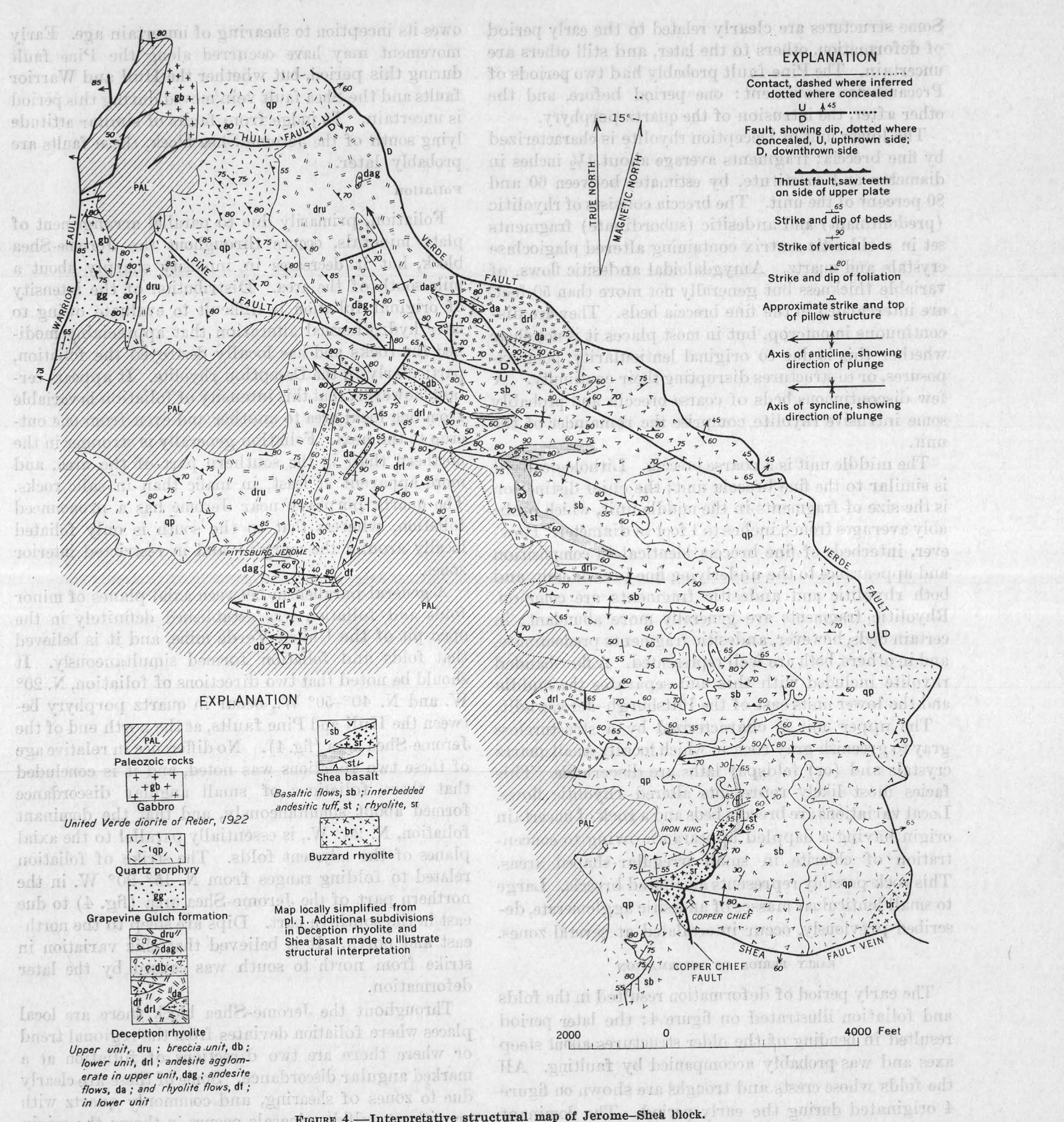

FIGURE 4.—Interpretative structural map of Jerome–Shea block.

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

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

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

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

EARLY PERIOD OF DEFORMATION

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

Foliation

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

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

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

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

Folds

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

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

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

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

Pine fault

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

LATER PERIOD OF DEFORMATION

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

FAULTS OF UNDETERMINED AGE

Hull and Warrior faults

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

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

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

Verde fault

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

Shea fault vein

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

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

Copper Chief fault

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

STRUCTURE OF THE ALDER GROUP

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

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

The later periods of deformation produced two broad zones of distributive movement; one is occupied by the Chaparral volcanics and bounded by the Chaparral and Spud faults, and the other is occupied by the Texas Gulch formation. These two wide "fault" zones divide the rocks of the Alder group into three blocks. Each block has structural continuity, and this continuity permitted us to propose a tentative stratigraphy for the central block. But attempts at stratigraphic correlation between the rocks in these blocks and fault zones were largely unsuccessful. The stratigraphy of the Alder group is thus in doubt, and much of the structure still remains to be unraveled.

Henceforth, for the purposes of clarity and ease of expression the different planar structures will commonly be referred to as *s*1, *s*2, *s*3, and *s*4—*s*1 refers to bedding; *s*2 to the foliation or cleavage produced during the early period of deformation; *s*3 to the foliation produced during later deformation; and *s*4 to the fracture cleavage and late foliation that locally cuts *s*3 and *s*2.

EARLY PERIOD OF DEFORMATION

Within the map area the early period of deformation produced steeply dipping beds (*s*1), foliation (*s*2), and a mineral assemblage containing actinolitic hornblende. The steeply dipping beds reflect isoclinal or nearly isoclinal folds. The foliation, which formed concomitant with the tight folds, dips 70° or more, and strikes north, north-northwest and north-northeast. We ascribe considerable significance to the local predominance of actinolitic hornblende as the dominant mafic mineral, for retrograde metamorphism apparently accompanied later deformation and altered the actinolitic hornblende to chlorite.

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

LATER PERIODS OF DEFORMATION

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

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

STRUCTURE OF ROCKS BETWEEN CHAPARRAL VOLCANICS AND TEXAS GULCH FORMATION

The block lying between the Chaparral volcanics and the Texas Gulch formation including the Indian Hills and the low hills north of the Indian Hills, so far as we know, is not cut by any major fault zones, so that all units lie in their normal stratigraphic position. The block thus is a structural unit. The chronology of the recognizable structural events consists of an early period of isoclinal folding accompanied by formation of axial plane foliation. A second period of deformation, manifest by distributive shearing, locally produced fractures, breccia zones, and foliated zones that intersect the older structures. The local shearing of the quartz diorite in the southwest corner of the Jerome area, particularly along the Spud and Chaparral faults, is ascribed to this period. So far as known, this was the last period of sufficiently intense deformation to foliate rocks. Fracture cleavage, joints, scattered occurrences of folded cleavage, and faults marked by gouge testify to later deformation, but the ages of most of these structures are uncertain. In comparison to the earlier periods of deformation, they indicate mild deformation of no great significance.

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

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

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

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

The planar structures in this block consist of *s*1 (bedding), *s*2 (early foliation), *s*3 (later folation, breccia zones, and fractures), and *s*4 (chiefly fracture cleavage). In general, *s*1 and *s*2 are parallel, but discordances mark outcrops scattered throughout the Spud Mountain volcanics. Most of these probably resulted from the intersection of the foliation with the "noses" of folds. *S*2 and *s*3 are discordant in Spud Mountain and in the hills lying southeast of Spud Mountain. Elsewhere *s*3 appears parallel to *s*1 and *s*2, although the extent to which *s*3 is superimposed on *s*2 in these other areas cannot easily be determined. Apparently the early deformation squeezed the rocks so intensely that the foliation, bedding, and fold axis became essentially parallel.

In the Green Gulch volcanics away from the Chaparral fault and in the breccia unit of the Spud Mountain volcanics in the southwest corner of the Jerome area, the bedding ranges in strike from N. 20° E. to N. 45° W., but the general strike is about north (pl. 1). On the southern flank of Spud Mountain and on the hill to the southeast, the strike of the bedding ranges from north to N. 20° E. On Spud Mountain, however, the strike swings from north to N. 45° W., and then to nearly north again, as if to parallel the Spud fault. Bedding as well as foliation makes the swing, which suggests folding of both structures. This folding is not entirely due to drag on the major fault zone bounded by the Spud and Chaparral faults and marked by the Chaparral volcanics, except where it parallels the Spud fault. *S*3 in the vicinity of Spud Mountain strikes N.

45°–65° E. and dips steeply northwestward. It clearly cuts *s*2, and comprises narrow foliated or brecciated zones and in some places, single planes along which shear has occurred. Drag folds occur along the margins of nearly all the *s*3 zones. Their pattern, similar to that along the Chaparral fault, indicates that the rocks on the north side of the *s*3 zones moved northeastward in relation to those on the south side. There is a possibility that any *s*3 zone also had a vertical component as well as the obvious horizontal component. This possibility becomes a probability for the zone occupied by the Chaparral volcanics and for the later movement that occurred in the andesitic tuffaceous unit of the Spud Mountain volcanics and in the western zone of basaltic flows in the Iron King volcanics. For in these three areas, mineral streaking on the foliation planes plunges steeply, and is believed to be the *a* direction of the structural coordinate system.

In the Indian Hills, foliation ranges in strike from north-northwest to north-northeast, and this range may be due to the presence of both *s*2 and *s*3. If so, the more westward-trending foliation is *s*2, and the more easterly *s*3. In general, actinolitic hornblende is associated with rocks (mostly andesitic breccia) whose foliation has the former trend, and chlorite with rocks (mostly andesitic tuffaceous rocks) whose foliation has the latter.

In both the Indian Hills and the vicinity of Spud Mountain, retrograde metamorphism accompanied the development of the *s*3 planar structure, so that chlorite is the dominant mafic mineral in the *s*3 zones, and actinolitic hornblende is characteristic of the rocks where only *s*2 is present. In other areas where there is only one structural trend, the presence of chlorite as the mafic constituent has been used as evidence for the superposition of *s*3 on *s*2.

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

Several lines of evidence suggest the superposition of *s*3 on *s*2 in part of the block east of Spud Mountain. Eastward from Spud Mountain, *s*3 becomes more and more prominent and the rocks become more and more chloritic; simultaneously, however, the angle between *s*2 and *s*3 becomes more acute, and finally *s*2 and *s*3 appear to merge. At this point the rocks are chloritic. If chlorite is indicative of *s*3, which is certainly true near Spud Mountain, the extent of the later period or periods of shearing can readily be determined. Chloritic rocks crop out from about the contact between breccia and tuffaceous facies of the breccia unit of the Spud Mountain volcanics, which is about 1,000 feet west of the Iron King mine, eastward about to the major fold axis near Humboldt. From this line eastward about to the contact between the Iron King volcanics and the tuffaceous unit of the Spud Mountain volcanics, actinolitic hornblende predominates. Chlorite again prevails from here eastward to the fault that bounds the Texas Gulch formation on the west. The rocks also become much more fissile near this fault, indicating that shearing was distributed through a wide zone. As is discussed in a subsequent section, the quartz diorite in the southwest corner of the map area may be younger than the early deformation. Narrow zones of rocks that appear to be foliated quartz diorite dikes occur near the Iron King mine. If these zones are quartz diorite dikes related to the large mass of quartz diorite to the southeast, then the andesitic tuff near the Iron King mine possibly was sheared during a later period of deformation.

Whatever deformation sheared the rocks near the Iron King mine also sheared the rocks as far east as Humboldt, for all have a common, northward plunging lineation (mineral streaking). It is perhaps more than fortuitous that the lineation in the Chaparral volcanics at the latitude of the Iron King mine also plunges northward, nearly parallel to that in the rocks around the Iron King mine. Certainly the lineation in the Chaparral volcanics is related to *s*3. The lineation in the central and western part of the Indian Hills and possibly that north of Highway 89A plunges northward, but that east of the Indian Hills plunges steeply north and south. The rocks east of the Indian Hills, however, were deformed by Paleozoic and later faulting, so that the variation in the plunge of the lineation might be due to this later faulting.

Lineations were observed only in the chloritic rocks. Where actinolitic hornblende is the predominant mafic mineral, the rocks are not fissile and break along the joints rather than along the foliation. The age and significance of the lineation will remain undetermined until the problem can be solved of whether movement occurred along the foliation in the presently chloritic rocks during the deformation that produced *s*3. The assumption that the lineation is related to *s*2 does not fit well with the southward plunge of the major syncline. The average plunge of the lineation is about

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

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

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

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

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

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

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

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

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

STRUCTURE OF GREEN GULCH VOLCANICS

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

The area underlain by the Green Gulch volcanics is too small to determine the structural pattern; it is only a "fragment" of the whole, about which little is known. Foliation, bedding, lithologic units, and minor fold axes all trend northward, except where dragged into northeast trends along the Chaparral fault. Foliation and bedding generally dip steeply westward. Foliation, bedding, and the axial planes of minor folds generally are all parallel, although in scattered outcrops bedding and foliation intersect along a northward-plunging lineation that appears to reflect a north plunge of the minor folds. Scattered determinations of the direction that beds face, though few and inconclusive, suggest that, except for local reversals due to minor drag folds, the formation faces west.

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

STRUCTURE OF TEXAS GULCH FORMATION

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

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

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

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

Much faulting also occurred within the Texas Gulch formation both between rocks of different lithology and within units of uniform lithology. The elongated mass of andesitic tuffaceous rock enclosed in the Texas Gulch formation in the southern part of the formation and the smaller elongated masses lying near 1,305,500 N.–416,000 E. and 1,328,000 N.–418,800 E. were all detached from the parent formation and squeezed or faulted into their present positions. The fault bounding the east side of the largest or most southerly mass of andesitic tuffaceous rock in places truncates folds and beds. The western fault conforms to the trend of the adjacent rocks, but is marked by a zone as much as 500 feet wide of fine-grained very fissile chlorite schist in which no vestige of the original character of the rock remains. The quartz diorite masses lying near 1,320,700 N.–414,000 E.; 1,341,000 N.–415,500 E.; 1,327,-500 N.–418,000 E.; 1,298,000 N.–417,000 E. also have mechanical contacts. Indications of contact metamorphism are lacking in the purple slates alongside these quartz diorite masses, and the pronounced cleavage passes uninterrupted from slate into quartz diorite. Evidence clearly indicates that the purple slate and the quartz diorite were sheared simultaneously. Angular discordances between foliation or bedding in purple slate and foliation or bedding in adjacent rhyolitic tuff indicate differential movement between the two rock units; and the thickening and thinning of the purple slate units shows differential movement within and along the margins of lithologically uniform units. The sheared-out middle limb of the fold near 1,276,500 N.–414,700 E. reveals a relative concentration of distributive movement in local zones. But despite the large amount of distributive shearing evident in the Texas Gulch formation, there was no great loss of structural continuity.

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

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

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

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

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

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

STRUCTURE OF THE ROCKS EAST OF TEXAS GULCH FORMATION

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

SHYLOCK FAULT

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

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

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

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

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

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

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

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

STRUCTURE OF THE PALEOZOIC AND CENOZOIC ROCKS

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

OLDER STRUCTURE

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

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

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

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

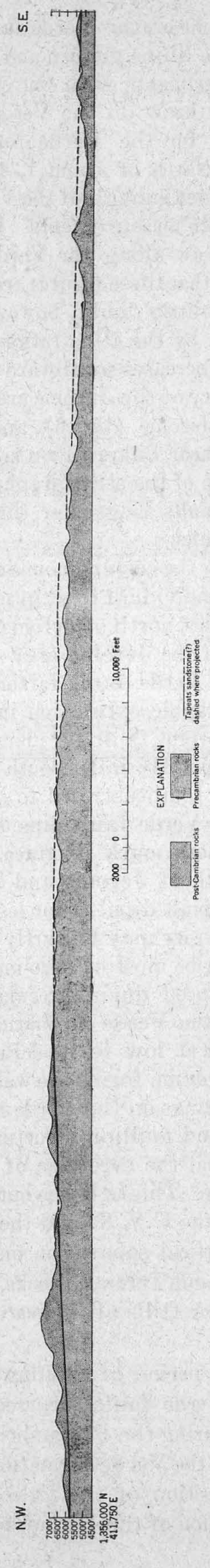

FIGURE 5.—Section through Black Hills showing arching of Tapeats sandstone(?).

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

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

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

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

YOUNGER STRUCTURE

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

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

VERDE FAULT ZONE

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

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

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

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

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

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

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

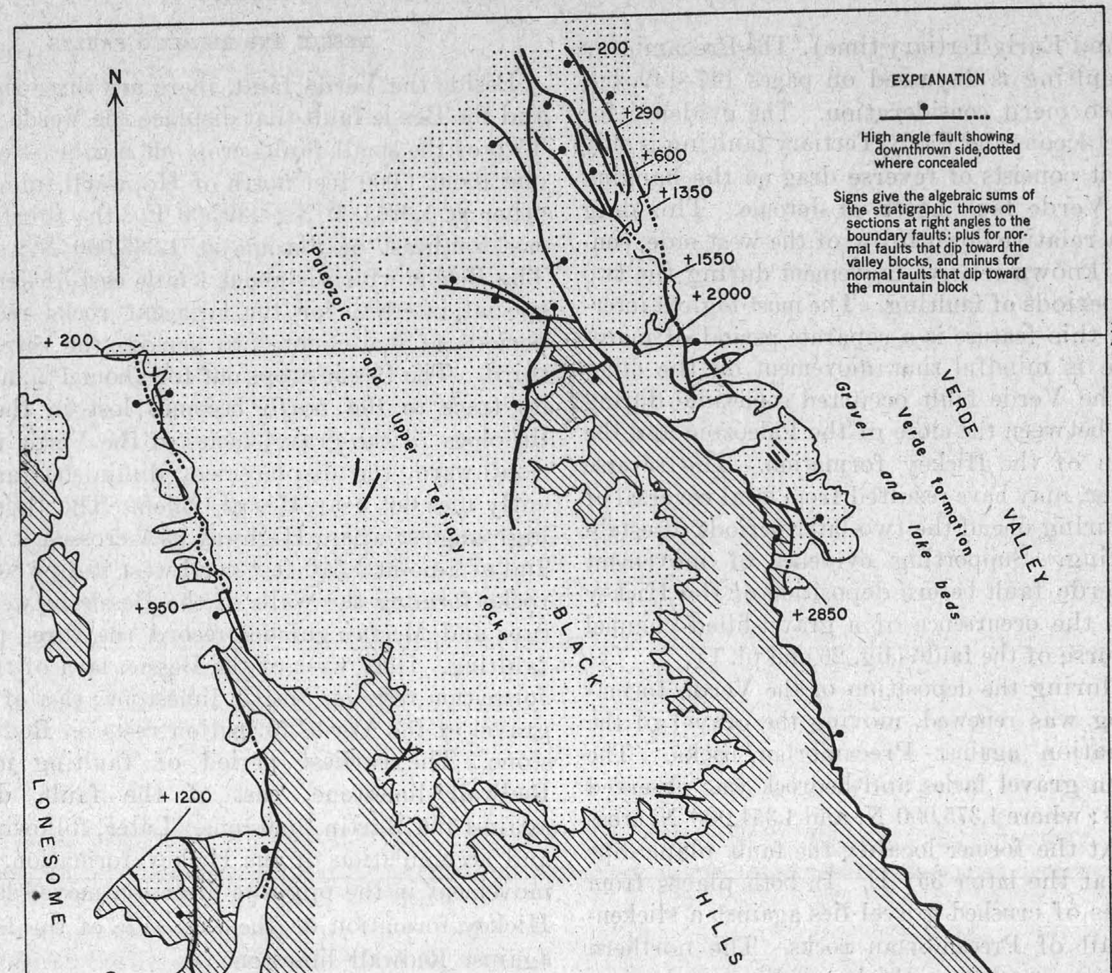

Figure 6.—Structural map showing Tertiary displacement in the Black Hills. Figures indicate sum of vertical separation along sections at right angles to boundary faults. Vertical separation on westward-dipping faults deducted from eastward-dipping faults and the proper sign is given to the difference.

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

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

BESSIE AND RELATED FAULTS

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

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

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

COYOTE AND RELATED FAULTS

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

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

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

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

OTHER FAULTS

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

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

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

PHYSIOGRAPHY

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

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

Drainage in the Black Hills is radial; to the north in the Clarkdale quadrangle, the streams drain northward to the Verde River. In the Jerome area, the eastern drainage flows to the Verde River in the Verde Valley. On the west side of the Black Hills, the drainage flows to Granite Creek in the Prescott quadrangle and to the Agua Fria River. Southward drainage from Mingus Mountain flows to tributaries that eventually join the Agua Fria River. The streams in the area are essentially consequent, in that their courses were determined by the original surface slopes. Tex Canyon, a western tributary to Ash Creek is an exception; its course parallels the structure of weaker beds in the Grapevine Gulch formation. The south wall of Tex Canyon consists of more resistant hornfels in the Grapevine Gulch formation adjacent to the quartz diorite. Grapevine Gulch for about a mile of its course (423,000 E.) follows a fault zone separating quartz diorite from Grapevine Gulch formation. The topography suggests that Grapevine Gulch captured the headwaters of Texas Gulch along the fault zone.

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

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

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

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

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

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

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

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

ORE DEPOSITS

HISTORY [4]

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

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

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

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

In 1595, Juan de Onate was given a contract for the conquest and settlement of New Mexico but it was not until 1598 that he started his expedition. When he arrived in the Hopi villages, he learned of the mines to the west and sent his Captain of the Guard and of the Horses, Marcos Farfan de los Godos, with eight companions and Hopi guides to visit the mines. Farfan arrived at the Jerome deposits on November 24, 1598, and found an old shaft, three estados (16½ feet) deep, and a large dump. Farfan and companions located many claims for themselves and companions who had remained behind in the Hopi villages, as they were impressed with the richness of the deposit.

The Spaniards never worked the deposits and no other record is available concerning the discovery or mining of ore minerals until 1863 when a party of prospectors under the leadership of Joe Walker discovered gold in Hassayampa and Lynx Creeks in the Prescott region (Hamilton, 1883, p. 47).

The oldest mine in the area described in this report is the Silver Belt, located in the Big Bug mining district, along the Silver Belt-McCabe vein, west of the Iron King mine (pl. 1). The Silver Belt mine was located about 1870, and rich silver ore was mined from 1870 to 1880. The ore was freighted to Ehrenberg on the Colorado River and shipped by boat to San Francisco. The mine was reopened in 1906, but no ore was found. The McCabe-Gladstone mine was located on the south end of the Silver Belt-McCabe vein and mining of gold-silver ore started about 1880. The mine was operated continuously from 1898 to 1913 by the Ideal Leasing Co.

The Verde Valley east of Jerome was first settled by white men in 1865 when a group of pioneers from Prescott located land near the Verde River and started farming. It is reported that in 1875, U. S. Army scouts from Fort Whipple at Prescott, discovered mineral showings near Jerome, and in February 1876, John O'Dougherty, Edward O'Dougherty, and John D. Boyd located claims on outcrops that later became the property of the United Verde Copper Co. M. A. Ruffner, one of the pioneers from Verde Valley, did considerable prospecting, and in June 1876 he and Angus McKinnon located the Eureka claim, south of the present pit of the United Verde mine, and in June 1877 they located the Wade Hampton claim which is in the west center of the present United Verde pit. Ruffner, George McKinnon, and Angus McKinnon sank a 45-foot shaft on one of these claims which showed good ore. Other prospectors were attracted to the area and much of the ground near the Eureka and Wade Hampton claims was located, later to be purchased by Charles Lenning, a wealthy Philadelphian.

Transportation was difficult in 1878 as the nearest port for water transportation was on the Colorado River, about 200 miles to the southwest, and the nearest railroad terminal was at Abilene, Kans. In spite of these difficulties, F. A. Tritle, Governor of the Territory of Arizona, who was living in Prescott, then the capitol, became keenly interested in the possibilities of exploiting the Wade Hampton and adjoining claims. He had been manager of the Yellow Jacket mine on the Comstock Lode in Virginia City, Nev., and had much experience in mining. Tritle supported F. F. Thomas in an effort to interest eastern capital in organizing a corporation to develop the claims in the Verde mining district. James A. Macdonald and Eugene Jerome (the town of Jerome was named for him) of New York provided the financial assistance to organize the United Verde Copper Co. in 1882, which purchased 12 claims including those held by Ruffner, the McKinnons, and Lenning. Macdonald was named president and Jerome the secretary-treasurer of the company.

Lenning sold his claims to the corporation under a lease and bond agreement for $75,000, and it is reported that the Ruffner and McKinnon claims were purchased for $15,000, probably through a third party who sold to the United Verde Copper Co.

The Atlantic and Pacific Railroad (later the Atchison, Topeka and Santa Fe Railway Co.) crossed northern Arizona in 1882, and a wagon road was built from Ash Fork to Jerome, 60 miles away. Supplies were freighted to Jerome by mule and ox teams at a cost of $20 per ton. Two small water-jacket blast furnaces were transported to Jerome and production of matte and bullion started in August 1883. Surface oxide ores were mined. These deposits were rich in gold and silver as well as copper, and averaged about $30 per ton.

The following statement as given by H. C. Burchard, Director of the Mint on the production of precious metals in the United States for 1884, is quoted by Lindgren (1926, p. 62):

> The mines owned by the United Verde Copper Co., at Jerome, have proved a series of surprises to the owners. The properties were purchased and worked as copper properties, but as they have been developed they are found to contain silver in large quantities—in fact, so large that the silver is sufficient to pay all the running expenses of the mine, leaving the copper as a profit to the owners.
>
> Superintendent Thomas writes that another rich strike has been made in the Wade Hampton, one of the company's mines, on a drift from the 100-foot level 25 feet north of the last body of ore struck on the same level. The extent of it had not been ascertained, but the first samples taken assayed 20 percent copper and rich in silver. In every direction that drifts or crosscuts have been run ore bodies have been encountered. The furnace has run up to October 1, 1884, 289 days, and Superintendent Thomas gives the product by assays at the mines 4,396,951 pounds of refined copper and 237,951 ounces of silver. Estimating the average price of copper at $250 per ton, the gross yield of copper amounts to $548,500, and the silver at its coining value, $1.29 per ounce, amounts to $307,655.

In this same report (p. 32) it was stated that the United Verde Copper Co. paid a dividend of $60,000 in 1884. By this date, however, costs of mining increased, the rich surface ores were exhausted, and copper prices dropped from 14½ to 7½ cents per pound. Profits were replaced by losses and the mine closed in 1884. Governor Tritle remained optimistic about the property and in 1887 secured a lease and resumed mining, but his losses were heavy.

W. A. Clark, who later became Senator from Montana, visited the New Orleans Exposition in 1885 and saw on exhibit many specimens of ore from the United Verde mine. Clark recorded a note about the specimens but on returning to Butte forgot about the Jerome deposit. Mines owned by Clark at Butte supplied ore to the Port Orford Copper Co. in New Jersey, and when this refinery went into liquidation in 1886, Clark as one of the chief creditors, assumed control of the refinery and operated it for a year or more. Records of many shipments of matte, extremely rich in gold and silver, from the United Verde mine caught Clark's attention and he sent J. L. Giroux to examine the United Verde mine. Giroux found that the controlling interest of the ownership of the mine was under option to Dr. James Douglas of Phelps Dodge & Co. Dr. Douglas decided not to exercise his option because of the inaccessibility and "spotty" character of the ore, and Clark obtained an option through the assistance of James A. Macdonald who became vice-president of the company under Clark's reorganization. Clark and Giroux visited the United Verde property in March 1888, and after inspection, Clark immediately made the first payment required under the option, acquiring 70 percent of the stock. Clark started buying the remainder of the widely scattered stock of the company and at the time of his death in March 1925, he and his family owned about 299,000 of the 300,000 issued shares.

Under Clark's guidance, active mining operations started in 1888, using the original small smelting plant. The first dividend under this regime was paid in 1892. The Bullock railroad from Seligman to Prescott had then been completed, and freighting in and out of Jerome was done by way of Prescott. A branch of the Atchison, Topeka and Santa Fe Railway Co. was built from Ash Fork to Phoenix via Prescott in 1893, and Senator Clark immediately built a narrow-gage railroad, the United Verde and Pacific Railway, connecting Jerome with the Atchison, Topeka and Santa Fe Railway Co. at Jerome Junction, north of Prescott. In 1894, Clark built a new smelter at Jerome. In this same year, the first of the many serious mine fires occurred in the United Verde mine, as the massive pyrite caught fire readily.

Active prospecting started in the Jerome district in 1900 and the Verde Queen smelter treated carbonate copper ore from near the Columbia shaft, later United Verde Extension property. In 1904–05 the Equator Mining and Smelting Co. mined and smelted copper sulfide ore from the Iron King claim, a part of what is now known as the Copper Chief deposit.

The Verde Central group of claims was located as the Verde King in 1904 and four short tunnels were driven (Fearing and Benedict, 1925, p. 609) to explore faults and contacts, as no gossan was present. Some copper-stained rock was discovered, but no ore bodies, and when funds were exhausted in 1907, exploration ceased.

Mining operations in the Yaeger mine, on the west side of the Black Hills in the Black Hills mining district, began about 1904 and produced some copper and silver until 1909, when the mine closed.

In the Cherry district, south of the Jerome mining district, gold-bearing quartz veins were explored by small-scale mining, starting in 1883; mining activity reached a maximum in 1907 when 7 mines and 6 mills were in operation. Activity slowed down and by 1916, only 2 properties produced a little bullion and a shipment of a little copper ore.

During the early history of mining, Jerome was essentially a "one mine" camp and no extensions of the United Verde ore bodies were found outside the property of the United Verde Co. Visitors were not permitted underground at the mine so that little information was available to the "outsider" concerning its geology.

The United Verde ore bodies cropped out conspicuously in the country rock west of the Verde fault. East of the fault, however, the ore-bearing host rocks are covered by a thick blanket of Paleozoic and Tertiary rocks so that the deposits were hidden, and exploration was costly.

The United Verde Extension mine, the second largest producer in the Jerome district, is east of the Verde fault and the ore body is buried by younger rocks. The history of this mine started in 1900 when J. J. Fisher, United States deputy mineral surveyor, was surveying claims for G. W. Hull, a Jerome resident of many years. Fisher was well acquainted with the district, having been employed by many companies, including the United Verde Copper Co., and he was aware of the importance of tracing the United Verde ore zone into outside ground. While surveying Hull's claims, Fisher found a fraction of vacant ground covering less than an acre, between the March claim owned by Hull and the United Verde property. Fisher located this fraction as the Little Daisy (fig. 7).

Earlier in 1899, a company organized by G. W. Hull, was called the United Verde Extension Gold, Silver, and Copper Mining Co. It became known as the U. V. X., and its successors have been known by the same abbreviation. This first company controlled a dozen claims west, south, and north of the United Verde property. L. E. Whicher of the Schofield, Whicher, and Co. of Boston became interested in the promotion of the U. V. X. Whicher acquired 190,000 shares and Hull kept 110,000 shares of the 300,000 shares issued by the company. A shaft was sunk in claim "1888" adjacent to and southwest of the United Verde mine. M. F. Johnson, mining engineer, was employed by Whicher to examine the holdings of the company, and Johnson reported his belief that if any ore was to be found outside the holdings of the United Verde Copper Co., it would be to the east. By this time Fisher had located the Little Daisy fraction, and Whicher obtained an option on the Little Daisy for $50,000 and sank a shaft 300 feet deep that passed through 125 feet of oxidized copper-bearing rock, too low grade for ore, but sufficiently promising that Whicher began negotiations with Hull for four claims adjoining the Little Daisy: the March, Conglomerate, Iron Carbonate, and Bitter Creek (fig. 7). As a result of the transaction, Hull transferred his 110,000 shares of U. V. X. stock to the U. V. X. company and deeded the four claims to U. V. X. in exchange for other claims near the United Verde mine owned by the U. V. X. company. Shortly thereafter, Whicher purchased the Little Daisy fraction from Fisher for 5,000 shares of U. V. X. stock. The U. V. X. company was reorganized in 1902 under the laws of Maine, and 300,000 shares at a par value of $10 were issued. In 1910, another reorganization of the company took place under the laws of Delaware and 400,000 shares of stock of the same par value were issued.

The Daisy shaft was sunk to the 800 level and the search for ore continued under Fisher's direction. In depth, all lateral exploratory openings to the east had to pass through the Verde fault characterized by a wide gouge and brecciated zone. From 1907 to 1911, C. C. Burger, who was confident of a favorable outcome to the exploration program, was the consulting engineer for the U. V. X. In 1911, ore assaying 2.6 to 3.1 percent copper and $2 in precious metals was found in a winze, 65 feet below the 800 level. The copper was present as a sulfide, but showed signs of leaching. In addition, a patch of chalcocite ore 5 feet wide and 15 feet long averaging 18.7 percent copper was found in a crosscut on the 700 level. Along the 800 level, a much larger body of low-grade oxidized rock was exposed, and its downward persistence was indicated by a winze. By this time, about $500,000 had been spent in exploratory work, and as the experts called in for examination of the property reported adversely, the exploration program ended.

In 1908 Major A. J. Pickrell purchased stock in the U. V. X. company from Fisher, and for a time Pickrell served as director of the company. When exploration was discontinued in December 1911, Pickrell wrote to James S. Douglas, son of Dr. James Douglas of Phelps Dodge & Co., and urged Douglas to visit Jerome and examine the exploration workings of the U. V. X. Douglas had been a frequent visitor to Jerome and at one time had examined copper carbonate showings on claims that later became a part of the Jerome Verde Copper Co. Thus Douglas had some familiarity with the local geology.

Douglas visited Jerome in December 1911 and was shown the U. V. X. workings of the 800 level, but because of water in the mine, he was unable to examine the winze or lower workings, which at that time extended to the 1,200 level. Douglas was favorably impressed with the mineralized rock on the 800 and 700 levels, and took an option on the property. This option was offered to Phelps Dodge & Co., whose legal counsel "disapproved of the transaction on account of some fancied defect in the title, in which supposition he proved to be wrong" (Rickard, 1918, p. 13). Thereupon Major Pickrell urged the younger Douglas to take over the option himself, and in the summer of 1912, Douglas and George E. Tener of Pittsburgh took the option and reorganized the company into the United Verde Extension Copper Co. and invited their friends to join in providing $225,000 to explore the property.

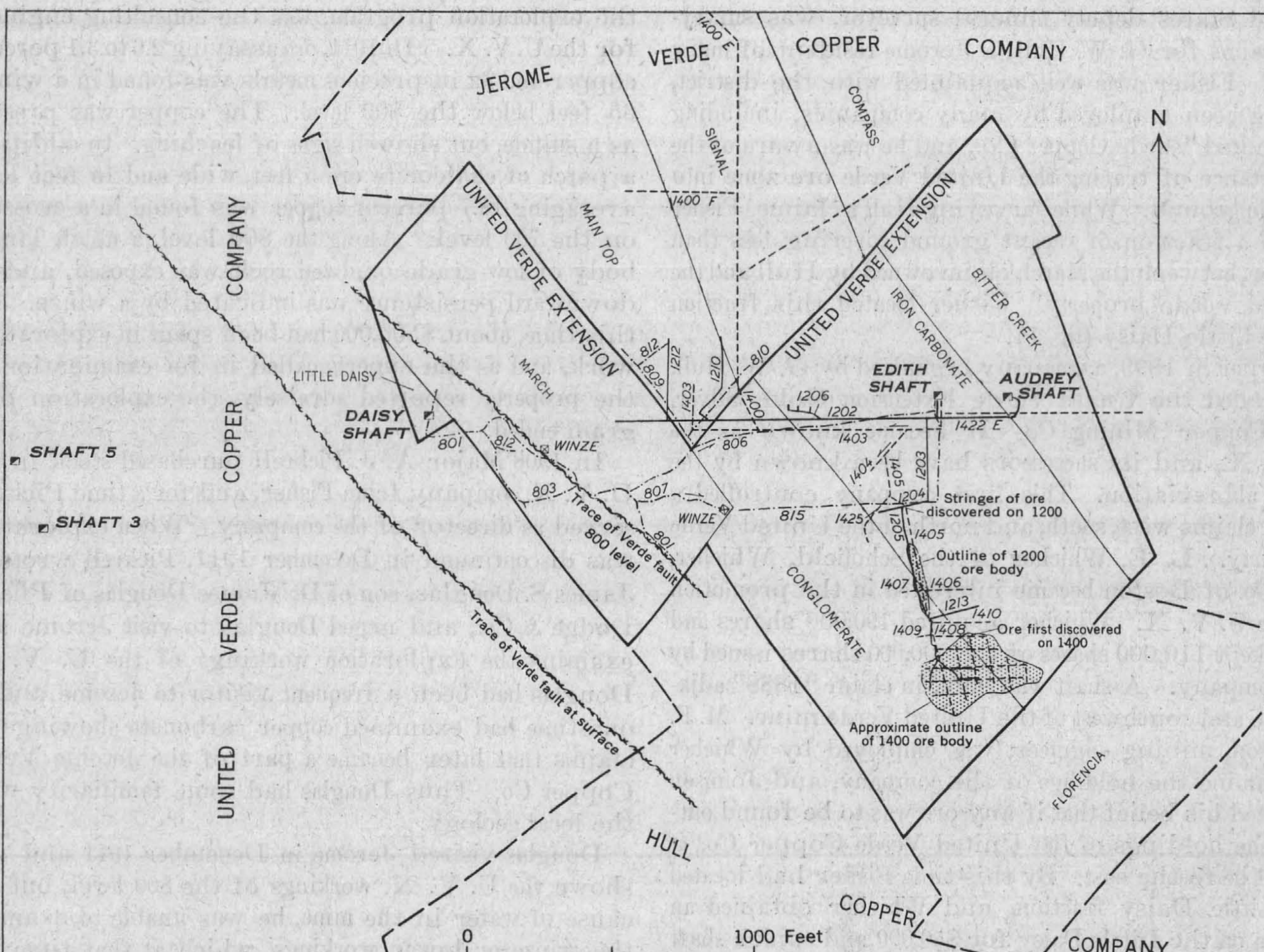

FIGURE 7.—Sketch map showing United Verde Extension property. Map of initial workings of the mine of the United Verde Extension Co. Numbers indicate the order in which the work was done. Modified from T. A. Rickard (1918, p. 15).

The necessary funds were raised by sale of stock; more than a third was bought by residents of Arizona.

Douglas decided to sink a new shaft farther from the Verde fault to avoid the trouble of driving workings through the wide crushed zone of the fault; and the Edith shaft, named after Miss Edith Tener, was started in June 1913—1,740 feet east of the Daisy shaft (fig. 7). The Edith shaft was sunk through 180 feet of lava, 400 feet of limestone, and 90 feet of sandstone, and reached the Precambrian rocks at 678 feet, which was equivalent to the 800 level of the Daisy shaft.

An agreement was made with the Jerome Verde Copper Co. to explore the Main Top, Signal, and Compass claims (fig. 7) to the north of the U. V. X. holdings, and the two properties were to be consolidated if ore was found on the Jerome Verde holdings. Under this agreement, exploratory work was done on the 800 and 1,200 levels without productive results. Likewise, search for ore on the U. V. X. ground was disappointing. The Edith shaft was deepened to 1,400 feet, and additional exploration was made in the Main Top claim of the Jerome Verde Copper Co. but no ore found; and in June 1915, the option on the Jerome Verde Copper Co. holdings was dropped.

In the fall of 1914, the U. V. X. company engaged the services of one of the foremost geologists in the country to examine the mine and show where the management had erred in their search for ore. "After a thorough examination, the distinguished geologist, whose name is purposely withheld, condemned the mine and strongly advised his clients to cease operations" (Rickard, 1918, p. 16). Lindgren (1926, p. 80) wrote, "I am not sure but that sympathy should be extended to this geologist. One may wonder how other 'experts' would have reported on the property." "However, as the Cornishman says, 'Never abandon a drift until you have driven 20 feet farther.' The management persevered a little longer" (Rickard, 1918, p. 16). Although the resources of the U. V. X. company were practically exhausted, in December 1914, a drift was driven toward the center of the U. V. X. holdings on the 1,200 level and a stringer of ore was found (fig. 7) which led to the discovery of a chalcocite body, 5 feet wide and containing 45 percent of copper. This was part of an ore body 120 feet long and reaching a little above the 1,100 level. About $600,000 worth of ore was mined from this body in 1915.

A drift was started on the 1,400 level to intersect the downward continuation of ore present on the 1,200 level, but it went beyond the point where ore was expected without finding ore. No crosscutting was done until January 1916, because of a heavy flow of water south of the ore body on the 1,200 level, which was not drained. Pumps were installed on the 1,400 level station, and the 1,400 exploratory drift continued under the "1,200 ore body," stopping in kaolinized rock 850 feet south of the Edith shaft. Crosscuts were driven both east and west in hopes of finding the high-grade ore discovered on the 1,200 level but these crosscuts failed as the ore body took a westward dip 60 feet below the 1,200 level, and the crosscuts were all too far to the east (fig. 7). The third crosscut to the east (labeled 1408, fig. 7) cut 100 feet of chalcocite ore and the main drift was extended into the ore body. Crosscuts driven 50 feet apart proved that an ore body reached within 30 feet of the Florencia side line (fig. 7). This was a new and larger ore body than the one discovered on the 1,200 level. On the 1,400 level, this mass of chalcocite ore had a maximum width of 260 feet and a length of 440, covering an area of 62,400 square feet. The upward termination of this main ore body was at 1,240 feet. "During 1916 the U. V. X. produced 36,402,972 pounds of copper from 77,461 tons of ore, an average of 23.5 percent copper, besides 2,570 ounces of gold and 128,468 ounces of silver, the total output being worth $9,949,918, of which $7,400,000 was profit. It had proved a glorious bonanza!" (Rickard, 1918, p. 17). In April 1917, the U. V. X. mine produced 6,991,480 pounds of copper, worth at the high prices then prevailing, $2,167,358.

Peak annual production of the United Verde Extension mine in terms of pounds of copper, was reached in 1917 when 63,879, 506 pounds of copper was obtained from 115,064 tons of ore. The ore averaged about 27.5 percent copper. A smelter was completed in 1918 at Clemenceau to handle the ore from the United Verde Extension mine. Peak annual tonnage production from the mine was reached in 1929 when 358,654 tons was mined, yielding 59,125,000 pounds copper, about 8.6 percent copper as the average grade mined for the year. In 1930, a 200-ton flotation mill was added to the smelter at Clemenceau to handle lower grade copper ore but in January 1937, as reserves in the mine were depleted, the smelter closed and in May 1938, the United Verde Extension mine closed; the deposit was mined out.

Naturally the discovery of the buried rich U. V. X. ore body led to many prospecting ventures in the district, but most of these were unsuccessful. An outlying ore body of the U. V. X. zone was found in 1917 in the Main Top claim of the Jerome Verde Copper Co., but this was mined out in 1920 after producing 1,500,000 pounds of copper (Elsing and Heineman, 1936, p. 101). During this period, chrysocolla-bearing gravel was explored by the Dundee-Arizona Copper Co. on two claims east of Jerome. The Copper Chief mine operated from 1916 to 1923, when a 125-ton cyanide mill was in operation and oxidized ore from the upper levels was mined. Copper-silver ore was mined from the Shea mine, south of the Copper Chief mine, during this same period. The Yaeger mine in the Black Hills district was active in 1922, and 800 tons of copper-silver ore was shipped to the Humboldt smelter.

The Arizona National mine in the Big Bug mining district is located on the Silver Belt-McCabe vein, and high-grade silver-lead ore and concentrates were shipped intermittently between 1915 and 1931. Total production during this period was valued at $250,000 to $300,000, judging from the available records.

At the United Verde mine, mine fires and subsequent ground movement required the construction of a new smelter as the old smelter was adjacent to the mine. The new smelter was built at Clarkdale and placed in operation in 1915. A standard gauge railroad from Drake on the Prescott-Ash Fork branch of the Atchison Topeka and Santa Fe Railway Co. along the Verde River to Clarkdale was completed in 1914. Stoping within the fire area of the upper part of the United Verde mine was discontinued in 1916, and open-pit stripping operations were started in 1917 to uncover the ore above the 600 level. Open-pit mining started in 1922. A standard-gage railroad was completed from Clarkdale to Jerome in 1920, and in 1921, a direct highway was constructed from Jerome to Prescott. A flotation concentrator was added to the smelter at Clarkdale in 1927 to handle copper ore associated with silicate gangue minerals. The United Verde mine had its peak annual copper production in 1929, obtaining 142,290,000 pounds of copper from 1,737,000 tons of ore, an average grade of 4.1 percent copper.

The Verde Central mine, south of the United Verde mine at Jerome, attracted the attention of C. T. Joslin of Prescott, in April 1916, and a new exploration program was started, and in March 1918, W. F. Staunton became president and general manager of the Verde Central Mines, Inc. In December 1918, T. H. Collins purchased $50,000 worth of stock from Joslin in order to keep the exploration program active. Calumet and Arizona took an option on a large block of stock in 1921 and provided the funds for continued exploration. In 1924, Calumet and Arizona took complete control of the property, and G. A. Campbell was made president and Staunton remained in charge as vice-president and general manager. A 350-ton flotation mill was placed in operation in 1929; and in 1929 and 1930, 8,600,000 pounds of copper was mined and milled, having a value with the silver content of $1,338,000 (Elsing and Heineman, 1936, p. 101). The mine closed in 1930 and the property was purchased by the United Verde Co. in 1931. No mining has been done since that date. During 1949, R. L. d'Arcy shipped some of the low-grade siliceous copper ore from the dump, and mined some oxide ore from the top of the hill to the south of the mine where the Verde Central ore zone crops out.

The underground operations at the United Verde mine stopped in 1931 owing to the low price of copper and were not resumed until April 1937, but in the open pit, stripping operations continued and mining of ore was resumed in 1934 and continued until 1940 and bottomed on the 630 level. The mine was purchased from the Clark family in 1935 by Phelps Dodge Corp., the present owner (United Verde Branch, Phelps Dodge Corp.). Because of depleted reserves, the smelter at Clarkdale closed in June 1950 and the mine continued operations on copper-zinc ore that was concentrated at Clarkdale whence the zinc concentrates were sold to custom smelters and the copper concentrates were shipped to the Phelps Dodge smelter at Douglas, Ariz. The mine closed in February 1953.

With the depletion of ore reserves at the United Verde mine, the Iron King lead-zinc mine in the Big Bug district (shown on the southwest corner of pl. 1) is the leading productive mine in the area described in this report. Some of the older residents of Prescott believe that the original location was made about 1880 in search for gold and silver, but that a few shallow workings proved disappointing, and there are no records of production. In 1906–7 the Iron King mine was operated by Reverend Ben Blanchard, who mined oxide ore. In 1907 production amounted to 1,253 ounces of gold, 35,491 ounces of silver, and 3,933 pounds of copper (Lindgren, 1926). Some secondary copper ore was reported from near water level. The mine was inactive until World War I when it was reactivated by George Colvocoresses in order to obtain sulfide ore for his smelter at Humboldt. By 1922, it is reported that several thousand tons of ore averaging $8 per ton in gold and silver had been mined. In 1933, Fred Gibbs sampled the mine and purchased the property in 1934 under tax sale for a few hundred dollars. In 1936 Gibbs and associates formed the Iron King Mining Co. to operate the mine, and in 1937 they shipped oxide gold-silver ore. A bulk flotation mill with a daily capacity of 140 tons was built in 1938 to separate the lead-zinc sulfide minerals. In 1939 the mill was converted to differential flotation to produce separate lead and zinc sulfide concentrates, and the mill capacity was increased to 225 tons daily. Shattuck Denn Mining Corp. purchased the property in 1942, and is the present operator. Mill capacity has been increased gradually, and in 1952, was 600 tons daily; gold, silver, lead, zinc, a little copper, and fluxing pyrite are recovered.

During World War II under the incentive of premium prices for copper, oxide copper ore was mined from the Tertiary gravel beds at the Dundee-Arizona mine, a short distance east of Jerome, and from Tertiary gravel and basalt flows on United Verde Extension property, south of the Edith shaft. The demand for siliceous flux for the Clarkdale smelter resulted in some small-scale mining of the oxidized ore at the Copper Chief mine and upper levels of the United Verde Extension mine, and the richer parts of the Verde Central dump were shipped to the smelter.

Early records are incomplete but the available data indicate that to 1952, the following approximate amounts of metal have been recovered from the Jerome area: gold, 1,890,000 ounces; silver, 65,600,000 ounces; copper, 1,868,000 tons; lead, 37,725 tons; and zinc, 127,000 tons. The Verde mining district dominated by the United Verde and United Verde Extension mines accounts for 82 percent of the gold, 87 percent of the silver, 18 percent of the zinc, and more than 99 percent of the copper. The Iron King mine accounts for 11 percent of the gold, 10 percent of the silver, 82 percent of the zinc, and 96 percent of the lead. Records are incomplete concerning the value of this production, but $700,000,000 is a fair estimate.

MINERALOGY

Minerals related directly to the ore deposits are described briefly in this section. The common rock-forming minerals of no economic importance are not mentioned. The ore and associated minerals are grouped according to their origin; those formed by ascending ore-forming solutions (hypogene minerals) and those formed by the action of descending waters (supergene minerals). The supergene group includes sulfide, oxide, carbonate, and sulfate minerals. Some of the so-called ore minerals, such as arsenopyrite, have no economic importance, but because of their genetic relationship to the ore minerals are described under that heading.

HYPOGENE MINERALS

ORE MINERALS

Pyrite (FeS_2).—This is one of the most abundant hypogene minerals in the area. It is a major constituent in the massive sulfide bodies of the United Verde, United Verde Extension, and Copper Chief mines at Jerome, and the Iron King mine at Humboldt. Pyrite is common in all the fissure veins. Scattered pyrite crystals are widespread in all parts of the area where there has been any appreciable hydrothermal alteration.

Pyrite is recovered in the concentrator at the Iron King mine, largely because of the gold content, but in part because of its use as a smelter flux. The large body of massive pyritic sulfide at the United Verde mine represents an important natural resource of sulfur that undoubtedly will be mined in the future when economic conditions are favorable.

Chalcopyrite ($CuFeS_2$).—This primary sulfide of copper and iron is the chief ore mineral at the United Verde and Verde Central mines and on the lower levels of the United Verde Extension mine at Jerome. Small quantities are present in many fissure vein deposits throughout the area. Chalcopyrite is present in the "copper vein" a short distance west of the Iron King mine.

Bornite ($CuFeS_4$).—Beautiful hand specimens of bornite ore are found locally in the United Verde mine, and much of the ore in the Yaeger mine was massive bornite. Small blebs of bornite are found in fissure veins near the Yaeger mine and in the south-central part of the Mingus Mountain quadrangle.

Galena (PbS).—Very fine to microscopic grains of galena are important economic constituents of the Iron King massive sulfide ore. Small grains of galena are present also in the United Verde massive sulfide, but not in recoverable amount. Scattered medium to large crystals of galena are common in most of the fissure veins.

Sphalerite (ZnS).—Sphalerite is the most abundant ore mineral in the Iron King mine, where it occurs as minute grains forming streaks and aggregates in the pyritic massive sulfide, and as grains interstitial to the pyrite. In the United Verde mine, sphalerite, in small quantities, is present throughout much of the massive sulfide body. Locally, there is some concentration of sphalerite, particularly where the copper content is low, but not in sufficient tonnage to be mined under past economic conditions. During recent years, sphalerite has been selectively floated in the concentrator from zinc-copper ore from the north ore body on lower levels of the United Verde mine. Scattered crystals of sphalerite are common in many of the fissure veins.

Tetrahedrite ($(Cu, Fe)_{12}Sb_4S_{13}$).— Argentiferous tetrahedrite was the chief ore mineral at the Shea mine. The chemical analysis (table 15) shows the low arsenic content in this mineral, proving that it is tetrahedrite.

TABLE 15.—*Chemical analysis of tetrahedrite, Shea mine*

[Analysis furnished by the courtesy of Phelps Dodge Corp.]

Cu	percent	31.7
Fe	do	3.3
Sb	do	23.4
As	do	2.1
S	do	22.8
Ag	ounces per ton	583.57
Au	do	.21

According to Lindgren (1926, p. 100) mercurial tetrahedrite was found at the Shylock mine, and upon de-composition, secondary cinnabar was deposited as coatings and filled cracks.

Tennantite ($(Cu, Fe)_{12}As_4S_{13}$).—Tennantite, the arsenical variety of gray copper, is present in both the United Verde and Iron King mines, where it adds to the silver content of the ore. The chemical analyses (table 16) are of tennantite from the United Verde mine.

TABLE 16.—*Chemical analyses of tennantite, United Verde mine*

[Analyses furnished by the courtesy of Phelps Dodge Corp.]

	1	2
Cu percent	41.6	36.95
Fe do	.3	
Zn do	10.9	
Sb do	Tr.	2.2
As do	17.4	16.0
S do	27.3	24.4
Ag ounces per ton	38.08	31.76
Au do	.075	.14

1. 2,700 level; 2. 1,350 level.

Tennantite high in zinc content is common although the 10.9 percent zinc here reported is higher than any reported by Palache and others (1944, p. 376). The high-zinc sample (analysis 1) occurred as massive homogeneous tennantite in a quartz-carbonate vein in the United Verde mine, in which there is no visible contamination by sphalerite.

Tennantite has been found in the Yaeger mine.

Pyrrhotite ($Fe_{1-x}S$).—Pyrrhotite is rare in the Jerome area, and has been recognized only in the Haynes massive sulfide body, west of the United Verde ore body. The pyrrhotite is associated with pyrite and chalcopyrite.

Arsenopyrite (FeAsS).—Small crystals of arsenopyrite are scattered throughout the massive sulfide ore bodies at the United Verde and Iron King mines. A cobaltiferous variety is probably present in a prospect southeast of the Shylock mine (1,315,000 N.; 422,000 E.), as cobalt bloom is present at the surface.

Molybdenite (MoS_2).—Molybdenite has been observed in only one locality in the Jerome area; a small prospect on Burnt Canyon (1,327,800 N.; 438,700 E.) reveals scattered molybdenite crystals in a quartz vein.

Gold (Au).—Gold is economically important in all the mines of the area, and is intimately associated with the sulfide minerals. At the Iron King mine, metallurgical tests and assays show that the gold is free, and occurs in galena, sphalerite, and pyrite. Pyrite carries most of the gold, chiefly because pyrite is more abundant. Visible gold has been reported from mines in the Cherry Creek district. Much of the gossan overlying the important sulfide deposits contains gold; it was this gold that attracted early attention to mining in the Jerome area.

Magnetite (Fe_3O_4).—Magnetite is rare in the United Verde massive sulfide body, but is more conspicuous in the Haynes ore body to the west where some concentration of magnetite was seen in chlorite associated with the sulfide minerals.

Magnetite forms in distinct crystals and masses in altered facies of the Spud Mountain rhyolite, 2½ miles east of the Iron King mine. Although of no present economic importance, magnetite is a major constituent in sedimentary jasper-magnetite interbeds in the Grapevine Gulch formation.

Hematite (Fe_2O_3).—The specular variety of hematite is present in late veinlets cutting the massive sulfide at the United Verde mine. Specular hematite is common as veinlets in all the Precambrian rocks south of Jerome. Hematite is an important constituent in many of the jasper-magnetite interbeds in the Grapevine Gulch formation.

GANGUE MINERALS

Quartz (SiO_2).—Quartz is one of the most important gangue minerals in the ore deposits in the Jerome area. Pure masses of quartz are closely associated with the massive sulfide deposits at Jerome and at the Iron King mine. Quartz is also common in the massive sulfide, interstitial to the sulfide grains. Many bodies of quartz and jasper south of Jerome are associated with minor sulfide minerals. Quartz is the dominant gangue mineral in the fissure veins.

Muscovite ($H_2KAl_3(SiO_4)_3$).—The fine-grained variety of this mineral, sericite, is common in the alteration zone associated with the Iron King massive sulfide ore bodies. Some of the rocks in the United Verde mine contain sericite, particularly the quartz porphyry and fine tuffaceous sedimentary rocks of the Grapevine Gulch formation, but it is debatable whether this sericite is of metamorphic or hydrothermal origin. We believe that much of it is hydrothermal.

Sericite is present in many of the alteration zones associated with the fissure veins.

Carbonate minerals.—Carbonate minerals recognized in the ore deposits of the Jerome area includes: calcite ($CaCO_3$); dolomite $(Ca,Mg)CO_3$; ankerite $CaCO_3 \cdot (Mg,Fe,Mn)CO_3$. The distinction of these varieties is not always simple, for commonly they occur in microscopic aggregates, and are recognizable only in thin sections under the microscope. Enough indices of refraction have been measured by microscopic methods to demonstrate the occurrence of calcite, dolomite, and ankerite, in many places closely associated.

Dolomite and ankerite are common in the massive sulfide bodies, interstitial to the sulfide granules. In the United Verde mine, dolomite and calcite are common in the late quartz-carbonate veins. Most of the fissure veins in the area contain variable amounts of these carbonate minerals.

Siderite ($FeCO_3$) has been reported by Lindgren (1926, p. 25) in the Shea and Yaeger mines, but no siderite was positively identified in our study.

Barite ($BaSO_4$.—Barite is reported (Reber, 1938, p. 59) in the Shea mine, not associated with the sulfide minerals, which may be older. Lindgren (1926, p. 25) reported barite in the Silver Belt vein, to the west of the Iron King mine.

Tourmaline (complex silicate of B, Al, Fe, Mg).—Minute needles of black tourmaline have been seen in the quartz of some fissure veins in the Cherry Creek district, and quartz-tourmaline rock occurs west of the Iron King mine.

Chlorite (complex silicate of Fe, Mg, Al_2O_3).—Chlorite is an important alteration mineral at the United Verde mine, for much of the copper is spatially related to the introduced chlorite that replaced different rocks to form nearly pure chlorite aggregates, referred to as black schist in the United Verde mine. The term "black schist" is used in this report for all chlorite-rich aggregates that have replaced any rocks in the area. Chlorite has been introduced in many of the Precambrian rocks south of Jerome, although few areas of black schist are large.

Microscopic studies revealed considerable variation in optical properties, and some chemical analyses of the chlorite were made by Phelps Dodge Corp. to check the optical data. The intermediate index of refraction, optical sign, and birefringence are the optical properties most easily determined. Much of the chlorite in the alteration zones is extremely fine grained, and the color of the aggregate is black. The rare large flakes of chlorite are pale green. In all samples tested the optical properties of the pale green coarse chlorite and of adjacent microscopic chlorite are identical.

Much of the chlorite measured has birefringence between 0.004 and 0.010, positive sign, negative elongation, and with beta between 1.585 and 1.622. The chlorite of lower index has 26.12 percent MgO and 5.48 percent FeO. Another sample of chlorite in this group, with beta index 1.610, has 18.98 percent MgO, and 20.04 percent FeO. The higher indices of refraction reflect a higher FeO:MgO ratio. Using Winchell's chart (1936, p. 649), the chlorite of lower index would be called clinochlore; the chlorite of higher index would be prochlorite. Much of the chlorite associated with the United Verde ore body belongs to the clinochlore-prochlorite group.

Chlorite with birefringence less than 0.004, and optically positive, is common in the andesitic dikes that cut the United Verde ore body and in some of the partly chloritized tuffaceous sediments of the Grapevine Gulch formation near the United Verde ore body. The intermediate index of refraction ranges from 1.622 to 1.635 in this group. The chlorite of lower index (beta, 1.622) has about 10 percent MgO and about 15 percent FeO. The higher index chlorite (beta, 1.635) has 9.9 percent MgO and 25.9 percent FeO. The chlorite of lower index would be called rumpfite and that of higher index ripidolite, following Winchell (1936, p. 648). One sample of ripidolite forms black schist associated with ore at the United Verde mine.

All chlorite studied from the chloritized Deception rhyolite near Mescal and Deception gulches, south of Jerome, has a birefringence less than 0.004, and is optically negative, with beta index between 1.618 and 1.635. Some chlorite in the black schist at the United Verde mine also belongs to this low birefringent, optically negative group, as does some chlorite in the chloritized Grapevine Gulch and gabbro in the United Verde mine. A sample from Deception Gulch, having an intermediate index of 1.624, contains MgO, 12.50 percent, and FeO, 22.83 percent. In chloritized gabbro from the Haynes ore body, on the west side of the gabbro, United Verde mine, the chlorite has an intermediate index of refraction of 1.643, and the MgO content is about 7 percent and the FeO content is about 40 percent. According to Winchell (1936, p. 648), the chlorite having an intermediate index below 1.630 would be called diabantite, and above 1.630 the chlorite would be called aphrosiderite. Following this classification, most of the low birefringent, optically negative chlorite in the Jerome area is diabantite.

Very little chlorite was found to be optically negative and with a birefringence between 0.004 and 0.010. The few specimens from this group have beta index above 1.640. Only one specimen of black schist associated with ore in the United Verde mine belongs to this group; its beta index is 1.642. Southwest of the Copper Chief mine along the Pipe Line road (1,344,000 N.; 445,000 E.), a 3-inch band of chloritized Shea basalt is composed of chlorite with beta index 1.646, 5.8 percent MgO, and 31.4 percent FeO. According to Winchell (1936, p. 648) this chlorite would be termed brunsvigite. East of the Shea mine in the Buzzard rhyolite (1,344,000 N.; 450,000 E.), small areas of black schist contain chlorite with beta index 1.660, 4.8 percent MgO, and 34.2 percent FeO. Winchell (1936, p. 648) would classify this chlorite as thuringite.

Plotting the molecular ratios of MgO; FeO, Fe_2O_3; and Al_2O_3 of 6 chlorite minerals for which data are suitable, in a triangular diagram, shows that the chlorite minerals in the Jerome area have a considerable range in the MgO-FeO, Fe_2O_3 ratio, but the Al_2O_3 ratio to the

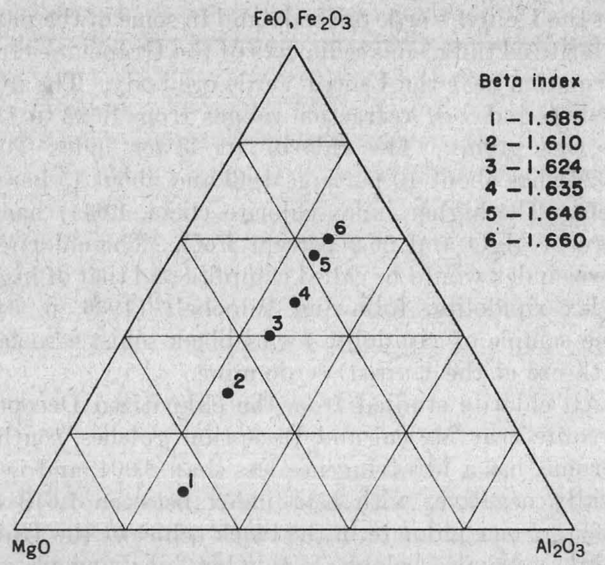

FIGURE 8.—Triangular diagram showing molecular ratio of FeO, Fe_2O_3; MgO; and Al_2O_3 in chlorite minerals.

MgO; FeO, Fe_2O_3 is nearly constant (fig. 8). The intermediate index of refraction is given, showing how the increased iron content is reflected in the higher index of refraction.

The data indicate that most chlorite forming black schist associated with ore in the United Verde mine belongs to the clinochlore-prochlorite varieties, though most other varieties of chlorite recognized here may locally form black schist also. Rumpfite-ripidolite is more common in the andesitic dikes that cut the United Verde ore body, and in the weakly chloritized Grapevine Gulch formation where the chlorite content is too low to form "black schist." Diabantite appears to be the common chlorite in the local chloritized parts of the Deception rhyolite south of Jerome. Aphrosiderite is less common among the chlorite minerals and is limited largely to chloritized gabbro in the United Verde mine and to Shea basalt, well to the south of Jerome. Brunsvigite and thuringite, the most iron-rich of the chlorite minerals, have been found only 4 to 5 miles south of Jerome except for one specimen of black schist in the United Verde mine.

Gavelin (1939, p. 70) determined the optical properties of the chlorite minerals associated with the massive sulfide deposits in the Malenas district, Sweden, and found most of the varieties to be optically positive, and to belong to the clinochlore-prochlorite-rumpfite groups of Winchell (1936).

SUPERGENE MINERALS

ORE MINERALS

Chalcocite (Cu_2S).—The rich copper ore at the United Verde Extension mine was composed of massive hard gray and soft, sooty chalcocite, formed by supergene enrichment in Precambrian time. Chalcocite was also mined from the upper levels of the United Verde mine, a product of supergene enrichment largely related to the present erosion surface. Chalcocite is present in the eastern part of the Copper Chief mine, where the grade was high enough in copper to warrant mining. Some secondary chalcocite was present in the upper part of the Yaeger mine. Small quantities of secondary chalcocite are found in all the fissure veins that contain primary chalcopyrite.

Silver (Ag).—Native silver was found in the upper levels of the United Verde mine in gossan directly overlying sulfide minerals. Enrichment in silver was noted at the United Verde Extension mine in a similar environment but Lindgren (1926, p. 86) does not state whether or not native silver was observed.

Copper (Cu).—Native copper is common in the United Verde Extension mine where it occurs with chalcocite and cuprite, and in the soft gossan overlying the chalcocite ore. Native copper was also mined from the oxidized upper part of the United Verde ore body.

Cuprite (Cu_2O).—This red oxide of copper was a common associate with native copper in the United Verde and United Verde Extension mines.

Copper carbonate.—Green malachite ($CuCO_3 \cdot Cu(OH)_2$) and blue azurite ($2CuCO_3 \cdot Cu(OH)_2$) are common in all the oxidized parts of all the copper deposits in the area. Scattered films of these two minerals are common at the surface of many of the fissure veins.

Chrysocolla ($CuSiO_3 \cdot 2H_2O$).—Chrysocolla was the chief copper mineral mined from the gravel beds in the Dundee-Arizona mine at Jerome. Surface films of chrysocolla are common throughout the Jerome area where copper is found.

Anglesite ($PbSO_4$) *and cerussite* ($PbCO_3$).—Anglesite and cerussite are common oxidation products of galena and are present in the oxidized parts of the Iron King ore body. They have also been reported from the Copper Chief mine.

Limonite (hydrated iron oxide).—This term is a desirable field term for the ferric oxide that contains varying amounts of water and occurs in yellow and brown colors. Much of the powdery aggregate includes jarosite, a hydrous sulfate of iron and potash. Limonite forms much of the gossan over all the sulfide deposits in the area. Limonite (gossan) has been mined at many of the deposits in the area because of the gold-silver enrichment. Much of the early mining at some deposits was due entirely to the precious-metal content.

MINERALS RELATED TO MINE FIRES

During the open-pit mining of the United Verde deposit, stripping operations uncovered hot rocks impregnated with various hydrous sulfate minerals formed as a result of the burning sulfide. Several methods were used to extinguish the fires, such as flooding with water, adding carbon dioxide, and later steam, but none was successful. Finally air was forced into the fire area under a pressure of 2 to 5 pounds and this method forced the hot gases back and cooled the rocks (Tally, 1917).

Lausen (1928) believed that the hydrous sulfate minerals were probably formed when water was first used to put out the fires, but as crystals of one composition rest upon others, conditions must have changed from time to time. The gases must have contained a large percentage of water vapor which reacted with oxidized sulfides of iron and copper and carried these constituents to higher levels. Presumably the constituents were transported upward in a gaseous phase.

In 1924, E. D. Gardner and G. W. Jones of the U. S. Bureau of Mines collected 5 gas samples from cracks and drill holes in the pit. These samples were analyzed by G. W. Jones of the Bureau of Mines and the results are given by Lausen (1928, p. 206). Two samples contained chiefly atmospheric gases, and in only two samples did the SO_2 content equal 6 percent. One sample contained 4.67 percent CO_2 which may have been derived from burning mine timber.

The hydrous sulfate minerals were deposited chiefly in cracks in the quartz on the west and north sides of the massive sulfide, although some were found along cracks in the massive pyritic sulfide.

Altogether, nine hydrous sulfate minerals were found, five of which are new (Lausen, 1928). The new minerals are as follows: butlerite, $(Fe,Al)_2O_3 \cdot 2SO_3 \cdot 5H_2O$; guildite, $3(Cu,Fe)O \cdot 2(Fe,Al)_2O_3 \cdot 7SO_3 \cdot 17H_2O$; louderbackite, $2FeO \cdot 3(Fe,Al)_2O_3 \cdot 10SO_3 \cdot 35H_2O$; ransomite, $CuO(Fe,Al)_2O_3 \cdot 4SO_4 \cdot 7H_2O$; rogersite, hydrous sulfate of ferric iron. The other hydrous sulfate minerals that formed are alunogen, $Al_2O_3 \cdot 3SO_3 \cdot 18H_2O$; copiapite, $2Fe_2O_3 \cdot 5SO_3 \cdot 18H_2O$; coquimbite, $2(Fe,Al)_2O_3 \cdot 7SO_3 \cdot 22H_2O$; voltaite, hydrous potash iron sulfate (Anderson, 1927). In addition, some sulfur was deposited and a black globular coating on fragments of rock beneath the iron hoods placed over the vents from which the heated gases were escaping. Lausen (1928, p. 227) named this material jeromite, a sulfide of arsenic containing some selenium.

GENERAL CHARACTERS

The ore deposits of Precambrian age in the Jerome area may be classified into two types: (1) massive sulfide deposits and (2) fissure veins. Most of the metal production in the area has come from the massive sulfide deposits, represented by the United Verde and United Verde Extension copper mines at Jerome and by the Iron King lead-zinc mine at Humboldt. The fissure veins have been mined chiefly for their precious-metal content, such as the gold-quartz veins in the Cherry Creek district and silver-bearing vein of the Shea mine south of Jerome. Copper has been the chief metal from the Verde Central and Yaeger veins, although much gold and silver were recovered from the Yaeger ore. The McCabe-Gladstone vein, west of the Iron King mine, was an important producer of gold and silver, but it cannot be proven that this deposit is of Precambrian age.

At Jerome, some copper has been mined from Tertiary lava flows and gravel beds impregnated with supergene chrysocolla.

The distribution of the principal mines in the Bradshaw Mountains and Black Hills is shown in figure 9;

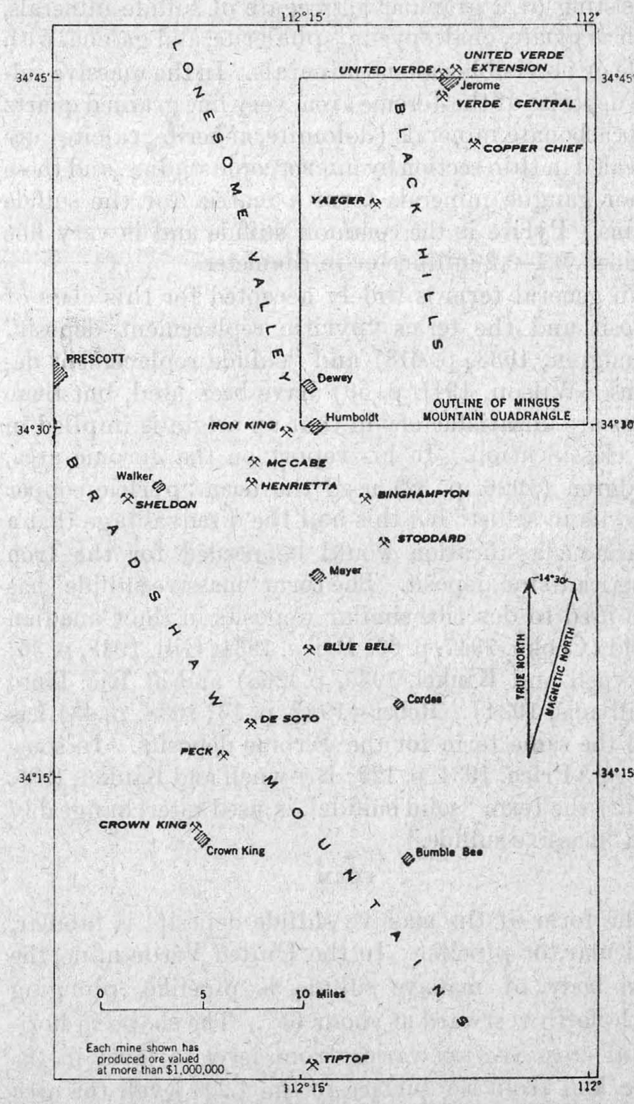

FIGURE 9.—Sketch map of Bradshaw Mountains and Black Hills showing location of principal mines.

only the mines north of the McCabe mine, southwest of Humboldt, are described in this report. Figure 9 shows that except for the Sheldon, southwest of Walker, the mines of significant production lie within two northeastward-trending zones, one extending from the Henrietta to the United Verde and United Verde Extension mines at Jerome, and the other extending from the Crown King to the Binghampton. Part of these deposits are massive sulfide and part are fissure veins. This northeast trend in part is reflected by the northeastward-trending structures in the Alder group, but not by any known structure in the Black Hills. Possibly this distribution of the principal mines is fortuitous; no satisfactory explanation can be advanced with our present information.

MASSIVE SULFIDE DEPOSITS

The term "massive sulfide" is used for the deposits consisting of a granular aggregate of sulfide minerals, such as pyrite, chalcopyrite, sphalerite, and galena, with little or no visible gangue minerals. In the massive sulfide deposits of the Jerome area, very fine grained quartz and carbonate minerals (dolomite, ankerite, calcite) are revealed in thin section by microscopic studies, and these minor gangue minerals form a matrix for the sulfide grains. Pyrite is the common sulfide and is very fine grained, 0.1–0.2 millimeter in diameter.

No general term is widely accepted for this class of deposit and the terms "pyritic replacement deposit" (Lindgren, 1933, p. 618) and "sulfide replacement deposit" (Wilson, 1941, p. 59) have been used, but these terms are unsatisfactory in that an origin is implied in the classification. In his report on the Jerome area, Lindgren (1926, p. 32) used the term "pyritic copper deposits in schist" but this had the disadvantage that a separate classification would be needed for the Iron King lead-zinc deposit. The term "massive sulfide" has been used to describe similar deposits in the Canadian shield (Cooke, 1947, p. 66; Price, 1934; Gill, 1948, p. 36; Brownell and Kinkel, 1935, p. 268) and at Rio Tinto (Williams, 1934). Reber (1922, p. 16; 1938, p. 43) has used the same term for the Jerome deposits. In some reports (Price, 1934, p. 129; Brownell and Kinkel, 1935, p. 272) the term "solid sulfide" is used interchangeably with "massive sulfide."

FORM

The form of the massive sulfide deposits is tabular, lenticular, or pipelike. In the United Verde mine, the main body of massive sulfide is pipelike, plunging north-northwestward at about 65°. The shape in horizontal cross section varies from level to level in the mine, and from the surface to the 1,200 level, the area of the massive sulfide deposit averages about 250,000 square feet, but on the 1,500 level, it is almost 500,000 square feet. On lower levels, the horizontal area of massive sulfide decreases so that on the 3,300 level it is only 50,000 square feet. The North ore body in the United Verde mine extends from above the 3,750 to below the 4,500 levels and is a northward-trending, steeply dipping, elongate lens, arcuate at the south end. The average width of this North ore body ranges from 10 to 30 feet, swelling locally to 60 feet.

The massive sulfide body in the United Verde Extension mine is a lens trending eastward; on the 1,300 level, the horizontal cross section area is about 100,000 square feet. The body tapers in depth and is in fault contact with the Verde fault.

The Copper Chief deposit is also lenticular, elongated in an easterly direction, but it is only about 300 feet deep. The ore body is about 800 feet long and 60 feet wide.

The Iron King massive sulfide deposit forms a set of en echelon veins, striking N. 22° E. and dipping 71° W. The north ends of the veins plunge northward at nearly 60°. Each vein on the west extends farther to the north than adjacent veins to the east. The width of the vein ranges from 1 to 14 feet, and the length measures in hundreds of feet.

The regional structure apparently has considerable control on the form of the massive sulfide bodies, as pointed out by Gavelin (1939). In areas characterized by pronounced foliation (*S*-tectonites) the massive sulfide minerals form veins or disseminated deposits parallel to foliation, as exemplified by the Iron King deposit. However, in a metamorphic terrane dominated by folds (*B*-tectonites) a close relationship may exist between fold axes, linear structures and form of the deposit, which results in lenticular and pipelike bodies. The United Verde pipelike deposits are apparently localized in the nose of a north-northwestward-plunging anticline, and the plunge of the pipe is parallel to the plunge of the minor folds and lineation marked by cleavage-bedding intersections.

MINERALOGY AND STRUCTURE

The massive sulfide deposits are dominantly pyritic and contain variable amounts of chalcopyrite, tennantite, sphalerite, and galena, as well as minor amounts of arsenopyrite and locally, pyrrhotite and bornite. The pyrite-rich facies is generally nonbanded, but where appreciable amounts of sphalerite, chalcopyrite, or galena are associated with the pyrite, banded facies are present.

The margins of the massive sulfide bodies are generally in sharp contact to the adjacent rocks, but in places the massive sulfide bodies grade into disseminated deposits where the sulfide minerals are in scattered crystals or aggregates or in veins in the host rocks. In the United Verde mine, some of the country rock containing veins and disseminated grains of chalcopyrite constitutes ore, but at the Iron King mine, the disseminated sulfide is pyrite that is not ore.

Veins of quartz or quartz-carbonate cut the massive sulfide bodies and related disseminated mineralized rocks, and are late in the sequence of mineralization. These veins locally contain the sulfide minerals found in the massive sulfide bodies except bornite and pyrrhotite. Argentiferous tennantite is common in these veins.

Table 17 shows the mineralogic similarity of several massive sulfide deposits.

TABLE 17.—*Comparison of the mineralogy of the Iron King and United Verde deposits with other massive sulfide deposits*

Iron King deposit, Arizona	United Verde deposit, Arizona	Huelva, Rio Tinto district, Spain (Bateman, 1927)	Shasta County deposits, California (Graton, 1909)	Sullivan deposit, Canada (Swanson and Gunning, 1945)	Rammelsberg deposit, Germany [1] (Lindgren and Irving, 1911)	Mandy deposit, Manitoba (Hanson, 1920)	Deposits at Kyshtim, Russia (Stickney, 1915)
Sulfide minerals							
Pyrite Arsenopyrite Sphalerite Galena Chalcopyrite Tennantite	Pyrite Arsenopyrite Pyrrhotite Sphalerite Tennantite Bornite Galena Chalcopyrite	Pyrite Arsenopyrite Sphalerite Galena Tetrahedrite Enargite Luzonite Famatinite Chalcostibite Whitneyite Umangite Hauchecornite Ullmanite Berthierite(?)	Pyrite Chalcopyrite Sphalerite Galena Bornite	Pyrite Arsenopyrite Sphalerite Chalcopyrite Galena Bornite Chalcocite Tennantite	Sphalerite Chalcopyrite Galena Pyrite Arsenopyrite	Pyrite Arsenopyrite Sphalerite Chalcopyrite Galena	Pyrite Chalcopyrite Sphalerite Tennantite Galena Bornite (rare)
Gangue minerals and hydrothermal alteration products							
Quartz Ankerite Sericite Apatite(?)	Quartz Ankerite-dolomite Chlorite Magnetite Specular hematite	Sericite	Quartz Calcite Barite	Chlorite Quartz Specularite Dolomite	Barite	Sericite Carbonates Quartz Chlorite Rutile	Quartz Barite Sericite Rutile

[1] Listed in order of abundance.

ALTERATION

The host rock is commonly silicified at some margins of the massive sulfide deposits. At the United Verde mine, silicified rock is largely restricted to the north contact where flinty quartz forms a narrow screen between the hanging-wall gabbro or Grapevine Gulch formation and the massive sulfide bodies. At the Iron King mine, fine-grained quartz is present at the north ends of the veins.

Sericitization of the host rock is common adjacent to the massive sulfide deposits. At the United Verde mine the host rocks are largely composed of sericite, quartz, and a minor amount of chlorite. At the Iron King mine, sericitization and minor silicification are in the zone of veins and in the zone of hydrothermal alteration lying to the west.

Widespread chloritization of the host rock is common on the south (footwall) side of the United Verde pipe of massive sulfide, and tuffaceous sedimentary rocks of the Grapevine Gulch formation and the quartz porphyry are converted in part or entirely to chlorite rock that forms appreciable masses. The chlorite rock is weakly schistose, mimetic after the foliation in the quartz porphyry and tuffaceous sediments. This rock has been called "black schist" by the geologists of the United Verde mine, and this term is used in this report for the chlorite-rich alteration product.

Chloritization probably followed the formation of the massive sulfide pipe because the black schist is localized on the footwall side of the pipe, and nonchloritized inclusions of Grapevine Gulch formation and quartz porphyry lie within the pipe on upper levels of the mine.

Quartz-carbonate veins and carbonate nodules and augen are common, and indicate formation of carbonate minerals throughout the period of mineralization.

ORE SHOOTS

It should be emphasized that the massive sulfide bodies in the United Verde and Iron King mines have not constituted ore unless enriched in base or precious metals. Much of the enrichment was primary and caused by the addition of metals in the fracture zones in the pyrite-rich facies resulting in replacement of some or much of the pyrite and associated quartz and carbonate gangue.

At the United Verde mine, the bulk of the chalcopyrite was not deposited until after the large bodies of black schist (chlorite rock) on the footwall side of the massive sulfide pipe were formed, and on the lower levels in particular, fractured black schist contains much chalcopyrite, resulting in large and important shoots of "schist ore." Locally, fracturing and introduction of chalcopyrite extended into the quartz porphyry on the footwall (south) of the black schist and some of the smaller ore shoots consist of chalcopyrite-bearing quartz porphyry. The largest and richest ore shoots in the United Verde mine occur in the massive sulfide in the intermediate and upper levels.

Sphalerite is locally abundant in the United Verde mine, appearing in the pyrite facies of the massive sulfide body, but no zinc ore has been mined except in the high-zinc ore shoot in the north ore body on the lower levels. Small amounts of sphalerite and galena are present in all the copper ore shoots, whether in the massive sulfide pipe, black schist, or quartz porphyry.

The ore shoots at the Iron King mine are near the north ends of the massive sulfide veins, and locally include parts of the quartz noses where the precious-metal content may be higher than average. The ore shoots formed by the introduction of galena and sphalerite along fracture zones in the earlier pyrite, replacing pyrite and carbonate.

Although primary enrichment in chalcopyrite was important in forming the ore shoots of the United Verde Extension mine, secondary enrichment in Precambrian time formed the rich chalcocite ore bodies. On the lower levels, the primary enrichment was similar to that of the United Verde mine.

Secondary enrichment was important in the upper levels of the United Verde mine, possibly enriching the chalcopyrite ore as deep as the 500 level, but detailed records of early mining activity that might have information on this matter are lacking.

The precious metals were residually enriched during oxidation of the massive sulfide deposits of the United Verde mine and high concentrations of native silver were found overlying the sulfide minerals in the open pit. Practically all the soft gossan at the United Verde mine was minable; the parts relatively low in gold and silver had sufficiently high silica for the ore to be used for converted flux. Much of the oxidized top of the United Verde Extension ore body was mined for copper and precious metals, yielding about one-eighth of the production of the mine. At the Copper Chief mine the soft limonitic gossan was like the gossan at the United Verde mine, except that the lead content was higher. The Copper Chief gossan was mined chiefly for the precious-metal content, like the gossan at the Iron King mine.

ORIGIN

All previous investigators (Reber, 1922, 1938; Fearing and Benedict, 1925; Lindgren, 1926) of the massive sulfide deposits in the Jerome area are unanimous in concluding that these deposits are of the replacement or metasomatic type and that schistose rock is the replaced material. We agree with this conclusion, although we place other interpretations on some features they considered as evidence of replacement.

In the United Verde mine, tuffaceous and cherty sedimentary rocks of the Grapevine Gulch formation and quartz porphyry, both foliated, are the host rocks that have been replaced by massive sulfide deposits. The distribution of the sedimentary rocks in relation to the massive sulfide deposits indicates replacement rather than displacement of the sedimentary rocks. The Grapevine Gulch formation on most of the levels, if not all, intertongues with, and abuts against the massive sulfide. On certain levels, such as the southwestern part of the 1,650 level (pl. 7), the bedding in the Grapevine Gulch formation strikes into the massive sulfide, indicating replacement. The extent of replacement that we ascribe to is shown on figure 10 by the assumed trace of the former Deception rhyolite-Grapevine Gulch formation contact through the massive sulfide.

Evidence of replacement origin is also given in the upper levels of the United Verde mine by the concordant structural orientation of unsupported rock masses within the massive sulfide pipe with the host rocks adjacent to the sulfide pipe. Near the margins of the massive sulfide, many small masses and streaks of unreplaced rock retain their original orientation, and locally layering of the sulfide and gangue minerals in the massive sulfide reflects the orientation of the foliation in the contiguous wall rocks. Relict bedding and drag folds in the Grapevine Gulch formation were recognized on the hanging wall (north side) of the massive sulfide pipe on the upper levels, proving replacement of the sedimentary rocks by the massive sulfide. Much of the banding in the massive sulfide is not relict schistosity, as has been assumed by earlier investigators, but represents closely spaced parallel fractures along which sphalerite, galena, or chalcopyrite have replaced earlier pyrite. In places the banding in the massive sulfide parallels schistosity in adjacent rocks, but in many others it is discordant.

The contacts of the massive sulfide bodies with the host rock are usually sharp but in the United Verde mine, gradational contacts between black schist and massive sulfide are common. The pyrite crystals are coarse in the gradational facies, and seem to represent pyrite introduced into the black schist after the massive sulfide was formed. The contact of the massive sulfide pipe shows all the embayments and irregularities characteristic of the walls of clear-cut sulfide replacement bodies in limestone.

Evidence of the replacement origin of the Iron King veins is less satisfactory than for the United Verde pipe. The absence of crustification and notable quantities of gangue minerals indicate that the veins are not fissure fillings. In some narrow veins, banding in the massive sulfide parallels the foliation in the chloritic schistose host rock (Spud Mountain volcanics) suggesting that the banding is perhaps relict foliation. Much banded ore, however, resulted from fracturing of early pyrite, quartz, and ankerite, and later introduction of sulfide that replaced the pyrite and ankerite. The gradation southward from massive sulfide to disseminated pyritic schist suggests that the massive sulfide veins formed by replacement. But in the final analysis, the theory of replacement origin is based largely by analogy to the United Verde deposit where the evidence is compelling.

The mineralizing solutions, presumably, were hydrothermal, as shown by the associated sericite at the Iron King mine and widespread chlorite at Jerome. Several chemical analyses of the Jerome chlorite and black schist contain 9 to 10 percent of water of crystallization (+105°C), indicating that the temperatures during mineralization were not high, possibly belonging to Lindgren's mesothermal zone.

The source of the mineralizing solutions and metals is debatable. Lindgren (1926, p. 35) concluded that no evidence is available to connect the solutions definitely with the Bradshaw granite, but they probably came from great depths, perhaps from granitic magma, and from levels where mineralizers such as fluorine, boron, and phosphorus had not separated to any great degree. Reber (1938, p. 59) stated that our general knowledge of metalliferous ore deposits makes it almost certain that the solutions were related to igneous rocks or magma reservoirs, but that the particular igneous rock or inferred magma reservoir to which the ore solutions are related is speculative. Reber suggested that the proximity of the massive sulfide deposits to the Bradshaw granite makes this granite a plausible deep-seated source for solutions, and that the ore solutions probably had a common deep-seated origin in the granitic magma of the Bradshaw.

The quartz diorite in the southern part of the Mingus Mountain quadrangle is part of the Bradshaw granite terrane of the Prescott-Jerome area. At the Copper Chief ore deposit, granodiorite porphyry dikes cut the massive sulfide lens and are younger (Reber, oral communication). These dikes appear to be satellites of the quartz diorite to the south, suggesting that the massive sulfide lens at the Copper Chief deposit is older than the quartz diorites; perhaps it can be presumed that all the massive sulfide deposits in the Jerome area are older than the quartz diorite.

At Jerome, the massive sulfide deposits are clearly younger than the quartz porphyry and gabbro, but older than eastward-trending andesitic dikes that cut both the gabbro and massive sulfide. These dikes only occur near the United Verde and United Verde Extension mines, and they are only distinguished on the mine maps (pl. 7). The general similarity in chemical composition of the andesitic dikes and gabbro as well as their close spatial relationship may be used as evidence that the dikes are genetically related to the gabbro and were intruded shortly after the emplacement of the gabbro. The massive sulfide deposits therefore would be closely related in time to the intrusion of the gabbro, and possibly, genetically related. Gabbro is widespread in the Jerome area, and any one massive sulfide deposit could be related to the nearest gabbro body.

The gabbro may be part of the early igneous activity culminating in the emplacement of the composite batholith of central Arizona (Bradshaw granite), and it may be academic to debate as to the relative importance of the gabbro and quartz diorite in relation to the ore-forming solutions. Granting that these solutions are of magmatic source does not answer the question as to the source of the copper, iron, magnesia, and sulfur introduced into the host rock to form the massive sulfide and black chloritic schist. These elements could either have been derived from deep magma chambers or picked up from country rocks and transported by the solutions for deposition above. If our structural interpretation is correct, an appreciable thickness of Shea and Gaddes basaltic flows should be present in depth below the massive sulfide deposits at Jerome, which are in the youngest rocks of the Ash Creek group. These basaltic flows could provide an ample source of iron and magnesia. It does not appear likely, however, that the copper in the Jerome deposit was derived from them.

The copper content was determined for the Gaddes basalt from Black Canyon, about 7 miles south of Jerome, away from the zones of pronounced alteration and metallization (table 18).

TABLE 18.—*Copper content of Gaddes basalt, in percent*

[Gaddes basalt from Black Canyon. Spectographic determinations by K. J. Murata]

Sample no.	*Percent Cu*
1	0.0023
2	.0012
3	.0012
4	.0030
Average	.0019

The average copper content of the earth's crust was estimated to be 0.01 percent by Clarke and Washington

(1924, p. 10), an estimate that has been widely accepted. Later, Sandell and Goldich (1943, p. 181) found 0.007 percent copper as the average, and that mafic igneous rocks contain about 10 times more copper than silicic rocks.

The copper content of 24 samples of Keweenawan lava averaged 0.012 percent, and 15 other subsilicic rocks averaged 0.0149 percent (Sandell and Goldich, 1943).

The average content of copper in the Gaddes basalt, 0.0019 (table 18), is below that of the average of the earth's crust and far below that normally present in basalts. This low-copper content does not substantiate the theory that the Gaddes basalt was the source.

FISSURE-VEIN DEPOSITS

Production from the fissure-vein deposits in the Jerome area has been sporadic, and none of the mines located in veins were active during the period of this study. Most of the available information concerning these veins is obtained from Lindgren (1926) and Reber (1938).

The gangue minerals of the veins are quartz, calcite, ankerite, and barite. The sulfide minerals are pyrite, arsenopyrite, chalcopyrite, galena, sphalerite, tetrahedrite (or tennantite), and bornite. Some veins formed through open-space filling and show local banding and crustification, and the sulfide minerals fill their centers or collect in local aggregates scattered through the gangue. Although replacement may have been locally important, in forming these veins, their general characters are commonly those of fissure veins.

The gold-quartz veins in the Cherry Creek district are chiefly north-striking shear zones locally mineralized with lenses composed of quartz, subordinate sulfide and ankerite, and minor tourmaline. Free gold was common in the massive quartz, but some gold was derived from oxidized sulfide deposits. The ore formed pockets or irregular small ore shoots and mining was largely limited to the oxidized zone which was locally as deep as 300 feet. Small quantities of chalcopyrite, bornite, sphalerite, and galena have been observed in the primary parts of the veins, which are discussed in more detail in the section on the Cherry Creek mining district.

Copper-silver veins with general easterly strikes have produced some ore from the Shea and Yaeger mines. The Shea vein, on the east side of Mingus Mountain, consists of coarsely crystalline quartz containing bunches of ankerite and pyrite with some chalcopyrite. Locally, a little arsenopyrite and argentiferous tetrahedrite are present, and where sulfide deposits are abundant, the vein has a banded structure. A little galena is present near the ore shoot that produced high-grade silver-copper ore from the abundant argentiferous tetrahedrite.

The Yaeger mine on the west side of Mingus Mountain has been the most productive of any of the fissure-type veins in the district. It contains some high-grade bornite ore in an east-striking calcite-quartz vein. Tennantite and a little pyrite are present, as well as some secondary chalcocite in the upper part of the vein. Some gold and silver were recovered. The Shylock vein, south of the Yaeger mine, contains tetrahedrite, galena, and sphalerite in massive quartz. No production has been recorded.

The quartz-copper vein in the Verde Central mine, south of Jerome, may not belong to the fissure-type vein, for although it contains considerable white quartz, most of it is greenish-gray, fine-grained quartz that may represent silicified Deception rhyolite (Fearing and Benedict, 1925). Pyrite and chalcopyrite are irregularly distributed in the quartz. In addition to the quartz-chalcopyrite vein, chalcopyrite and pyrite were found in black chloritic schist formed along the contact of the Deception rhyolite and quartz porphyry, resembling the "black schist ore" on the footwall side of the United Verde massive sulfide pipe.

West of the Iron King mine, the Silver Belt-McCabe vein strikes northeast; it was mined chiefly for the gold-silver content. Ankerite or siderite are present next to the vein walls, and the sulfide minerals fill the centers of the veins. Argentiferous galena and argentiferous tetrahedrite are the important sulfide minerals, but some sphalerite and pyrite are present.

Most of the veins in the area studied are clearly of Precambrian age; in the Cherry district and east and west of Mingus Mountain, nonmineralized Paleozoic sedimentary rocks overlie or crop out near the veins. West of Humboldt where the Silver Belt-McCabe crops out, however, there are no Paleozoic sedimentary rocks to date the mineralization. Lindgren (1926, p. 127) suggested that these veins may be related to unmetamorphosed rhyolitic dikes presumably younger than Precambrian and possibly of Late Cretaceous or Early Tertiary age.

VERDE MINING DISTRICT

The Verde mining district is located on the eastern slope of the Black Hills, parallel to the northwest front of the range. The district is about 7 miles long and about 3 miles wide. Two mines, the United Verde and the United Verde Extension, located at Jerome, have contributed more than 99 percent of the total production of the district, and commonly the district is called the "Jerome district."

Annual production data on the Verde mining district for 1908–51 are given in table 19. Earlier annual production data are not available. From production data assembled by Elsing and Heineman (1936, p. 101) and combined with the data in table 19, it appears that from 1883 to 1951 the dollar value of recoverable metals from the Verde district has been about $650,000,000. For the same period, the following amounts of metals have been produced: gold, about 1,565,000 ounces; silver, more than 57,000,000 ounces; copper, about 1,860,000 tons.

UNITED VERDE MINE

PRODUCTION

Underground production at the United Verde mine totaled 22,519,846 tons to January 1, 1952.[5] From 1900 to 1910, the annual production averaged 245,000 tons, and except for a shutdown in 1921–22, when the mine

[5] The following information on production and exploration at the United Verde mine was obtained from a mimeographed pamphlet prepared by C. E. Mills, general superintendent, United Verde Branch, Phelps Dodge Corp., and distributed to members of the Underground Mining Branch, Arizona section, Am. Inst. Min. Metall. Eng., March 1947, and from subsequent annual reports of Phelps Dodge Corp.

TABLE 19.—*Production of gold, silver, copper, lead, and zinc from Verde (Jerome) district in terms of recoverable metals* [1]

Year	Number of mines producing	Ore treated or sold (short tons)	Gold (ounces)	Silver (ounces)	Copper (pounds)	Lead (pounds)	Zinc (pounds)	Total value
1908	1		20,335	494,574	36,183,089			$5,460,818
1909	1		17,021	495,479	36,695,259			5,370,766
1910	2		19,316	398,247	38,663,880			5,624,449
1911	4		15,240	461,145	33,167,987			4,705,395
1912	2		15,083	484,222	31,565,539			5,817,855
1913	2	393,866	20,667	641,074	35,334,694			6,291,252
1914	1		21,401	646,572	32,448,170			5,115,499
1915	5		28,405	922,273	50,266,821			9,851,397
1916	3		37,725	[2] 1,250,000	[2] 103,000,000			[2] 26,992,986
1917	6	949,094	36,443	1,884,673	145,933,703	205,817		42,177,535
1918	7	1,011,966	37,327	1,986,845	133,780,865	7,962		35,802,862
1919	4	566,445	21,148	949,436	72,332,704			14,954,395
1920	4	870,129	24,363	1,275,323	107,084,032			21,597,163
1921	3	189,474	4,671	319,646	26,567,602			3,845,288
1922	3	572,246	20,345	843,151	74,793,982			11,360,899
1923	4	1,465,006	84,418	2,480,692	140,425,401			24,421,705
1924	3		52,030	1,839,844	142,238,947			20,941,417
1925	4		56,740	2,097,951	151,920,324			24,201,438
1926	2		55,538	1,971,914	150,269,756			23,416,169
1927	3		53,513	1,920,901	141,261,930			20,700,536
1928	3		61,677	2,195,595	163,312,041			26,076,159
1929	4		75,404	2,749,290	211,542,611			40,255,410
1930	3		[2] 50,796	[2] 1,530,000	[2] 117,600,000			[2] 16,925,000
1931	2		[2] 18,141	[2] 1,050,000	[2] 44,100,000			[2] 4,525,000
1932	3	241,113	10,086	361,273	35,790,460			2,565,165
1933	3	243,245	12,288	338,860	33,201,565			2,497,543
1934	1	200,754	8,065	242,632	26,147,463			2,530,505
1935	4	851,993	30,850	1,049,934	75,535,518			8,103,822
1936	5	1,312,012	64,405	1,911,774	99,157,152			12,857,309
1937	2	1,326,940	62,748	1,840,150	84,742,620			13,873,393
1938	3	799,564	45,541	1,144,652	57,863,745			8,004,559
1939	4	999,023	40,312	1,327,472	75,430,241			10,156,737
1940	6	896,727	24,652	1,144,028	74,459,646			10,090,291
1941	6	1,121,004	32,047	1,544,317	84,484,800			12,189,032
1942	6	1,242,133	28,429	1,532,108	87,874,000			12,717,268
1943	5	898,699	18,117	1,036,194	68,467,100			10,271,667
1944	5	525,263	8,620	589,538	52,429,000			7,798,842
1945	5	392,295	8,602	475,290	40,224,000			6,069,294
1946	4	350,465	8,132	418,578	32,351,000			5,863,693
1947	3	350,645	6,931	367,778	29,205,600			6,708,600
1948	4	358,491	11,374	408,669	29,087,800		917,000	7,201,970
1949	2	410,607	10,790	509,828	34,429,800	71,600	8,700,000	8,711,854
1950	2	361,320	9,412	456,254	26,581,400	100,000	15,600,000	8,500,299
1951	1	299,729	7,325	408,891	19,484,000	205,000	20,310,000	9,073,455
Total			1,296,473	47,997,067	3,287,436,338	590,379	45,527,000	$572,216,691

[1] Figures for 1908–31 are from Elsing and Heineman (1936, p. 86). Figures for 1932–51 are from Minerals Yearbooks, U. S. Bureau of Mines.
[2] Estimated.

was closed, underground production increased to a maximum of 1,062,500 tons in 1924. Underground mining stopped in 1931 owing to the low price of copper and did not start again until April 1937. Production then increased to a maximum tonnage of 1,015,200 tons in 1942. Since then, the rate of production has decreased steadily. In 1949, copper-zinc ore was mined for selective flotation and by January 1, 1952, 489,988 tons of copper-zinc ore had been mined, compared with 570,753 tons of copper ore for the same period. In 1951, copper-zinc ore comprised much of the production, 211,578 tons as compared with 87,858 tons of copper ore. The mine and concentrating mill closed in February 1953.

Approximately 15,542,000 cubic yards or 32,000,000 tons of rock was stripped from the open pit to produce 8,153,000 tons of direct shipping ore. The stripping operation included oxidized ore placed on surface dumps and later used for siliceous flux; by January 1, 1947, the smelter had used 1,688,200 tons of this dump ore. The open-pit mining was completed by April 1, 1940, and bottomed at the 630 level.

DEVELOPMENT AND MINING

According to C. E. Mills, the development throughout the period that the property was in operation totaled 81 miles of underground workings: 304,470 feet of drifts, 123,800 feet of raises, and 14,100 feet of shafts. This represents about 1 ton of development for each 46 tons of ore mined underground.

The upper levels are spaced 100 feet apart, except for the 1,200 level which is 200 feet below the 1,000 level. The remaining levels to the 3,450 level are 150 feet apart. The lower levels are the 3,750, 4,050, 4,200, and 4,500 levels. The 4,050 and 4,200 levels are not complete and extend only from shaft 8 to the North ore body.

Eight shafts were sunk since mining operations started. Shaft 1, which was started in 1883, was sunk to the 500 level and used until 1894. In 1900 shaft 2 was sunk to the 700 level, deepened to the 1,000 level in 1909 and abandoned in 1912. Shaft 3 is a three-compartment timbered shaft, started in 1901 and sunk to the 1,950 level by 1917. This shaft was abandoned after completion of shaft 5 and was used later as a mine ventilation raise. In 1910 shaft 4, a timbered shaft, was started; in 1914 it was extended to the 1,000 level and was abandoned in 1921.

Shaft 5, which has three compartments and is of concrete construction, served for hoisting. It was begun in 1915 and completed to the 3,150 level in 1927. A timbered section below the 3,150 level was sunk to the 3,515 in 1931. A double-drum hoist (5 by 10 feet) with an 800-horsepower motor handled 7.5-ton ore skips. The hoist room is on the 1,000 level and the ore pocket at the 800 level.

Service shaft 6, of concrete construction, was started in 1919 and completed to the 3,000 level in 1929. A single-drum hoist (6 by 12 feet) with a 500-horsepower motor handled a double-deck cage in an 8 by 13-foot service compartment. The hoist room is on the 500 level, and the sheave room is on the 400 level.

Shaft 7, which has five compartments of concrete construction, was started in 1929 and completed to the 3,000 level in 1933. A double-drum hoist (7 by 12 feet) equipped with a 2,700-horsepower motor to handle 10-ton ore skips was installed in the hoist room at the surface (300 level). An ore pocket was built at the 800 level. This shaft was sunk to replace shafts 5 and 6 when it was feared that mine subsidence might make these shafts inaccessible. However, both shafts have been kept in operation and it has not been necessary to use shaft 7.

Shaft 8 is a two-compartment timbered shaft from the 2,850 to the 3,300 level and has three compartments from the 3,300 to the 4,630 level. This shaft was started in 1940 and completed in 1942. The hoist room is on the 3,000 level and is equipped with a double drum ($3\frac{1}{3}$ by 7 feet) driven by a 450-horsepower motor. Skips in tandem with cages handle a $3\frac{1}{2}$-ton load. The dump pocket is below the 2,850 level, and small loading pockets are below the 3,450, 3,750, 4,050, 4,200, and 4,500 levels. All ore below the 3,000 level was hoisted in shaft 8 to the 3,000 level and hauled to shaft 5.

Ore was transported from shaft 5 ore bins on the 1,000 level through the Hopewell tunnel, 6,600 feet long, to ore bins on the surface. The Hopewell tunnel is 9 by 11 feet in cross section. Ore was transported by 25-ton trolley locomotives hauling 40-ton cars on standard-gage track.

Men and supplies were transported through the 500-foot level adit connecting the 500-level surface plant with shaft 6. This adit is 1,600 feet long and 8 by 9 feet in cross section.

Early underground mining was done chiefly by horizontal cut and fill stoping, leaving vertical pillars where ore bodies are large (Cowley and Quayle, 1930). The vertical and floor pillars on the middle levels were later mined by square-set method. Starting in 1937 when the mine was reopened, scrapers were introduced and the mining method was changed to inclined cut and fill stoping. For the pillars and remnants of ore in the upper levels, a modified Mitchell slice system was adopted (Pullen, 1941).

The zinc-copper stope in the ore body 800 feet north of the main sulfide pipe was mined from the 4,500 to the 3,750 levels, using diamond-drill long-blast holes and drawing ore without filling the stope with waste (Little, 1950).

SMELTING OPERATIONS

The smelter at Clarkdale had a daily ore capacity of 5,000 tons and was operated until June 1950 when it was partly dismantled. A concentrating mill for handling black schist ore started operations in 1927 (Barker, 1930), and late in 1948 one section was converted to differential flotation for the separation of sphalerite and chalcopyrite from the North ore body. The history of smelting (Lanning, 1930) and descriptions of the crushing plane (Keefe, 1930), roasting plant (Parsons, 1930), Cottrell plant (Denny, 1930), blast-furnace smelting (Kuzell, 1930), reverbatory smelting (Mooney, 1930), and converters (Williams, 1930) are available.

GEOLOGY

OLDER PRECAMBRIAN ROCKS

Two formations of the Ash Creek group, Deception rhyolite and Grapevine Gulch formations are exposed in the underground workings of the United Verde mine and on the surface adjacent to the mine. Quartz porphyry, gabbro (United Verde diorite of Reber, 1922), and andesitic dikes are associated intrusive rocks.

Deception rhyolite

The Deception rhyolite crops out in discontinuous exposures from the Hull fault to the south to the 500 level surface workings to the north (pl. 5). The intrusive quartz porphyry has cut out much of the Deception rhyolite south of the mine. A small patch of Deception rhyolite, partly covered by Paleozoic sedimentary rocks, crops out in the Verde fault zone at the east margin of the gabbro. This patch of rhyolite must have been down-faulted as shown by the capping of Paleozoic rocks, and before faulting was not closely associated with the Deception rhyolite now exposed to the south.

The Deception rhyolite occurs west of the Verde fault near the portal of the 500 level adit in the northern part of the United Verde mine area (pl. 5). Some andesitic (or basaltic) rock is intercalated with the rhyolite, like other mafic flows in the Deception rhyolite in Mescal Gulch.

In the underground workings of the United Verde mine, rhyolite is exposed on the 1,650, 1,000, and the 500 levels (pl. 7) and in the open pit. These workings expose three separate masses of rhyolite. One is in the haulage adit on the 500 level and on the surface near the portal. The second mass extends from the surface in the pit downward through the 500 level, where it is exposed 500 feet east and 600 feet northeast of the massive sulfide, and through the 1,000 level where it is cut 800 feet east of the massive sulfide. It is a thin wedge along the east side of a larger mass of Grapevine Gulch formation; both trend from northward to north-northeastward, and dip very steeply eastward. The third mass presumably extends from the surface in the pit downward through the 1,000 and 1,650 levels, southeast of the massive sulfide (not shown on pl. 7). On these levels, the north margin of the rhyolite is near 800 S., whereas on the surface, it is at 500 N. (pl. 5), indicating a southward plunge. This mass is bounded on the east by the Verde fault, and on the other sides by quartz porphyry.

Breccia structure locally occurs in the rhyolite on the 1,000 and 1,650 levels. Elsewhere the rhyolite is massive, similar to most of the Deception rhyolite near Jerome that has been hydrothermally silicified and sericitized.

Grapevine Gulch formation

The surface exposures of the Grapevine Gulch formation are west of the Deception rhyolite. The Grapevine Gulch formation is (1) in lenticular masses associated with Deception rhyolite surrounded mostly by quartz porphyry, (2) in a screen separating quartz porphyry from the gabbro (United Verde diorite of Reber, 1922), and (3) in an irregular mass that intertongues with the quartz porphyry east of the massive sulfide (pl. 5). Although now largely covered by dump, the Grapevine Gulch formation west and north of the gabbro occurs on both sides of the Haynes fault.

Most of the Grapevine Gulch formation in and near the United Verde mine consists of the fine-grained tuffaceous sedimentary facies with conspicuous bedding. Crystal tuff, chert, cherty shale, and some slate are the dominant rock types. Locally, a little limestone is present. The lithic tuffaceous facies is exposed in a few places underground, but absent from most surface exposures. Reber (oral communication) noted conglomeratic bands in the area now covered by dump.

Rhyolitic breccia intercalated in the Grapevine Gulch formation is exposed west of the Warrior fault and south of the Haynes fault (pl. 5). Flow-banded fragments 3–4 inches across are conspicuous, and no bedding is recognizable. In the western exposure, some of the rhyolite is strongly foliated and the clastic structure is not recognizable. Possibly some of these rhyolitic rocks represent lava flows. Diamond drilling shows that the rhyolitic rocks disappear at depth, only to reappear at greater depths. Two interpretations are possible: (1) the rhyolite is the Deception rhyolite exposed in the crest of an anticline, or (2) the rhyolite is intercalated between tuffaceous sedimentary rocks of the Grapevine Gulch formation, and represents a return of volcanism after the start of deposition of the rocks of the Grapevine Gulch formation. The second interpretation is favored because of the probable lenticular nature of the rhyolitic rocks revealed by drilling.

Quartz porphyry

The quartz porphyry in the United Verde mine area occurs in three masses, southeast, north, and west of the main body of gabbro. The southeast mass is the northern part of the large quartz porphyry body that extends southward beyond Mescal Gulch. This large mass forms the footwall of the main massive sulfide pipe, and is present from the surface to the lowest level in the mine.

About 400 square feet of quartz porphyry is exposed northwest of the portal of the 500 level adit. This quartz porphyry and associated Grapevine Gulch formation and Deception rhyolite occupy a trough in the gabbro, which is concealed by the Paleozoic rocks to the north. On the lower levels north of the gabbro near the north ore body, a much larger mass of quartz porphyry is present. It was probably connected to the quartz porphyry northwest of the 500 level adit before the intrusion of the gabbro. On the 3,300 level, diamond drills cut through 350 feet of quartz porphyry north of the gabbro (pl. 7, 2,500 N.; 500–1,000 W.). On the 3,750 level, quartz porphyry is exposed for 800 feet north of the gabbro, and on the 4,500 level (pl. 7) a similar amount is cut by drill holes and drifts. On the 4,500 level, part of the Grapevine Gulch formation separates the gabbro from the quartz porphyry. The exact size and shape of this northern mass of quartz porphyry is not known, but the horizontal area at depth is certainly greater than exposed at the surface.

The small mass of quartz porphyry west of the gabbro (pl. 5, 320 N.; 1,150 W.) is only 30 feet wide and 100 feet long, appearing as a tongue in the Grapevine Gulch formation. This small tongue probably was formerly connected with the larger mass of quartz porphyry exposed southeast of the gabbro, and that the intrusion of the gabbro separated them.

The quartz porphyry is well foliated south of the massive sulfide pipe, but is more massive to the east (pl. 5). Locally, feldspathic facies are present; the feldspar phenocrysts are now composed of a felt of sericite. The quartz porphyry from near the mine has the same mineralogy as the Deception rhyolite, that is quartz, sericite, and some chlorite, but the large quartz phenocrysts in the quartz porphyry aid in the distinction of the two rocks.

On many levels, a highly foliated purplish rock contains large quartz crystals and elongated sericitic aggregates probably pseudomorphs after feldspar. This purplish rock is present only as narrow locally persistent bands. In most places, the purple rock is at the margin of quartz porphyry and Grapevine Gulch formation. Its origin is somewhat in doubt; it may be a marginal facies of the quartz porphyry impregnated with hematite derived from the adjacent tuffaceous rocks, or perhaps a crystal tuff in the Grapevine Gulch formation, whose hematite pigment may be a contact metamorphic effect of the quartz porphyry.

At the surface in Hull Canyon, tuffaceous sediments of the Grapevine Gulch formation are locally purplish, where they are in contact with or form inclusions in the quartz porphyry, and locally, the quartz porphyry is purplish at the margin. If these rocks were highly foliated, the original character would be uncertain. Because of this doubt concerning their origin, the purplish rocks found in the United Verde mine are of dubious value for stratigraphic purposes.

Gabbro (United Verde diorite of Reber, 1922)

The term "United Verde diorite" was given by Reber (1922, p. 6) to designate the mafic plutonic rock exposed in the United Verde mine, but as was pointed out earlier in this report, the chemical composition compels designating the rock as gabbro. Because the gabbro is widespread in the Prescott and Mingus Mountain quadrangles, this variance with local nomenclature seems justified for this report.

At the United Verde mine, the gabbro forms a large northward-trending semiconcordant mass. To the north, the gabbro at the surface is cut off by the Haynes fault (pl. 5). Southward, the gabbro extends to Deception Gulch. At the surface, the eastern contact of the gabbro is essentially concordant with the bedding of the Grapevine Gulch formation; with depth, it dips west to the 3,000 level, and east below that level.

At the surface (pl. 5) 1,200 N.; 2,000 W. a small irregular-shaped mass of gabbro is separated from the main mass of gabbro by a belt of Grapevine Gulch formation and the Warrior fault. At depth, the two gabbro bodies probably are connected, for on the 1,500-foot level, no Grapevine Gulch formation is in the crosscut that cuts the gabbro to 2,050 W.; 1,700 N. (pl. 5).

The gabbro is, in general, very massive and well jointed, but along the margin of the gabbro, foliated facies are present; the foliation plane essentially parallels the contact.

Andesitic dikes

Many dikes, ranging from 1 inch to 50 feet in width, cut the Grapevine Gulch formation, quartz porphyry, and gabbro, as well as the massive sulfide pipe. The dikes do not intrude the overlying Paleozoic sedimentary rocks indicating that they are also of Precambrian age.

These dikes are called andesite following Reber (1922, p. 19). Lindgren (1926, p. 58) suggested that a more suitable term would be diorite porphyry, but the original texture and mineral composition cannot be determined owing to the altered nature of the dikes, and there is little merit in debating between andesite and diorite porphyry. The original dikes apparently had a fine-grained texture and were rich in mafic minerals and plagioclase.

The dikes trend eastward and dip almost vertically, except for a few larger ones on the lower levels that trend northeastward and are only on the northwest side of the massive sulfide.

The dike rocks are intensely altered so that the original mineralogic character is largely masked. In most of the thin sections examined, the rock consists of abundant sericite, dolomite, and chlorite and a few tiny quartz grains. Apatite prisms and leucoxene are accessory minerals. In two thin sections, polysynthetic twinned albite was recognized in randomly oriented euhedral crystals.

The chlorite in the dikes is pale green in thin section, with low birefringence. The optic angle is small (15°–20°), positive. The intermediate index of refraction ranges from 1.620 to 1.628; following Winchell (1936), this chlorite mineral is either rumpfite or ripidolite.

Weak foliation in the dikes is best shown in the older mine workings, where it is parallel to the stronger foliation in the adjacent quartz porphyry; it is recognizable in thin sections by a subparallel arrangement of the chlorite and sericite.

TABLE 20.—*Chemical analyses of andesitic dikes and related materials in United Verde mine*

[In columns 2, 3, and 4, the location in the mine is not known. Analyses furnished by courtesy of Phelps Dodge Corp.]

	1	2	3	4	5
SiO_2	41.03	52.00	45.50	48.40	2.88
Al_2O_3	17.20	19.75	18.41	27.12	39.33
Fe_2O_3	1.52	3.43	1.28	1.83	1.12
FeO	8.09	7.86	19.99	8.64	------
MnO	.22	------	------	------	------
MgO	5.75	3.62	6.30	3.55	.72
CaO	10.00	9.40	Tr.	Tr.	.60
K_2O	------	1.97	1.16	.89	8.42
Na_2O	------	1.46	2.04	1.04	------
TiO	.80	------	------	------	------
H_2O-	.07	.10	.20	.40	.50
H_2O+	------	.10	5.10	6.75	11.20
Loss on ignition	11.16	------	------	------	------
S	.02	.66	.40	------	------
ZnO	.81	------	------	------	------
Cu	.02	------	------	------	------
CO_2	------	Tr.	.25	1.95	------
SO_3	------	------	------	------	35.41

1. Andesite dike, 3,000 level.
2. Andesite dike, fresh material.
3. Partly kaolinized dike.
4. Kaolinized dike.
5. Alunitelike material from altered dike, 1,650 level.

In the upper levels of the United Verde mine, the andesitic dikes have been extensively altered to clay containing gypsum, and form "water courses." Chemical analyses of partly altered and altered dike rock from the upper levels are given in table 20. Presumably sulfuric acid solutions derived from the weathering of pyrite caused the clay alteration, as this type of alteration is not found in the lower levels. The acid solutions must have had a low-copper content, as no chalcocite was found in the adjacent sulfide-bearing rocks.

PALEOZOIC AND TERTIARY ROCKS

The Tapeats sandstone(?) crops out in a nearly continuous band west and north of the United Verde mine (pl. 5). It is an important marker for determining displacements along the many minor faults. One small patch is exposed in the Verde fault zone, northeast of the pit area.

The Martin and Redwall limestone beds overlie the Tapeats sandstone(?) west and north of the United Verde mine and also appear in the Verde fault zone to the northeast.

The Tertiary rocks include gravel overlain by basalt (Hickey formation) and are exposed only on the east (hanging-wall) side of the Verde fault (pl. 5).

MAJOR STRUCTURES

Folds

Evidence indicates that a major control for the location of the United Verde deposit is a small north-northwestward-plunging anticline (fig. 10) on the western flank of the north-northwestward-plunging Mingus anticline (fig. 3). The evidence, however, is not clear cut; a major difficulty is the obscuring effect of the intrusive bodies of quartz porphyry and gabbro.

Regional studies indicate the existence of the Mingus anticline, as discussed on page 64. The existence of the smaller fold localizing the mineralizing activity is indicated by the trace of the Deception rhyolite-Grapevine Gulch formation-contact, the configuration of the southeast margin of the gabbro, and the minor structural elements such as the minor folds, lineation, and foliation.

The trace of the Grapevine Gulch formation-Deception rhyolite contact provides an important link in the chain of evidence, if the interpretation of the course of the contact through the quartz porphyry (fig. 10) is correct. Southwest of the mine, the contact is well exposed in Deception Gulch (pl. 1); and here, excellent evidence shows that beds face west, indicating an anticlinal structure to the east. The contact strikes slightly east of north and is buried farther north by the Paleozoic sedimentary rocks. North of the Hull fault, tongues of quartz porphyry intrude the Grapevine

Gulch formation and Deception rhyolite, but the contact can be traced with assurance to 700 S. (pl. 5). Northward, quartz porphyry has removed all of the Deception rhyolite, but patches of Grapevine Gulch formation remain between the gabbro and quartz porphyry or massive sulfide.

East of the mine on the quarry bench (300 level of the mine), the contact between Grapevine Gulch formation

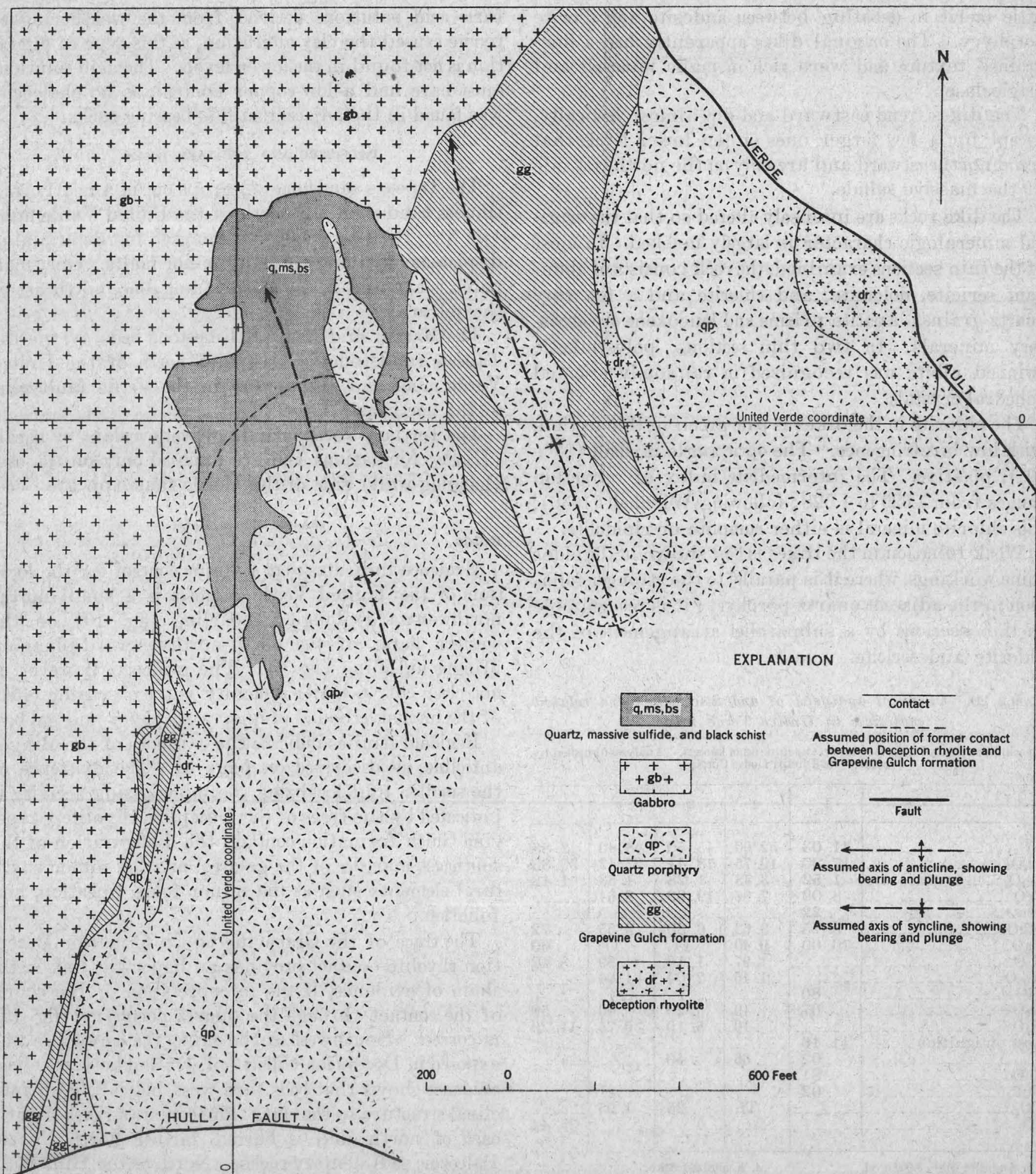

FIGURE 10.—Suggested structural interpretation at United Verde mine.

and Deception rhyolite is exposed for 950 feet (pl. 5, 200 S.–750 N.; 900 E.); the trend of the contact swings from north-northwest to north-northeast to within 200 feet of the Verde fault where it is cut out by quartz porphyry. The south end of the contact, which is at 200 S. (pl. 5), has shifted eastward from 100 W. to 1,000 E. The offset of this contact can be explained by faulting or by a north-northwestward-plunging anticline and parallel syncline lying to the east (fig. 10). No evidence of such a fault has been found, and it can be considered only as a remote possibility.

Minor folds are common in the Grapevine Gulch formation, and the abundant tongues of quartz porphyry cutting through the formation may in part be guided by them. They are best exposed in the southern walls of the open pit (pl. 5, 100 N.; 800–900 E.). The fold axes trend N. 20° W. and the folds plunge north-northwestward 40°–55° (pl. 4*B*). Foliation in the Grapevine Gulch formation and the quartz porphyry parallels the axial planes of the folds (pl. 4*A*), and the plunge of the intersection of bedding and foliation (lineation) (pl. 3*F*) is essentially the same as the plunge of the minor folds.

Northeast of the massive sulfide pipe (pl. 5, 650 N.; 500–700 E.) many minor folds in the Grapevine Gulch formation crop out. The foliation, lineation, and axes of the folds strike more nearly north in contrast to the N. 20° W. strike on the south wall of the pit. Most of the foliation in the quartz porphyry south of the United Verde mine strikes N. 20° W., indicating that this is the prevailing trend of foliation and bearing of lineation and fold axes. Northeast of the massive sulfide the northerly strikes are adjacent to the gabbro and their deviation in bearing is interpreted as the result of eastward "push" by the gabbro as it was emplaced. Local foliation along the gabbro parallel to its margin, also indicates movement of the gabbro after emplacement.

The axis of the massive sulfide pipe in the United Verde mine trends N. 20° W. and plunges 65° northward, coinciding very closely with the general bearing of the foliation, lineation, and minor folds and plunge of lineation and fold axes.

The gabbro is largely concordant wherever its contact with the Grapevine Gulch formation can be observed. The north-northeast trend of the gabbro mass is essentially that of the Grapevine Gulch formation. In detail, the gabbro contact parallels adjacent bedding in the Grapevine Gulch. The configuration of the gabbro contact in the United Verde mine has been determined from many diamond-drill holes and some underground workings. Northwest of the massive sulfide pipe the gabbro contact forms an inverted trough striking N. 20° W. and plunging 60°–65° to the north to the 3,000 level where it becomes straighter and strikes N. 45° E. (fig. 11).

The conformity of the bearing and plunge of the "inverted trough" in the margin of the gabbro with the bearing and plunge of the foliation, lineation, and minor fold axes appears to be more than fortuitous, particularly as the gabbro is essentially concordant wherever the contact can be observed.

The possibility that the gabbro has been folded to form the inverted trough is disproved by the lack of all but the local marginal foliation, which is discordant to the north-northwest regional foliation. In the Grapevine Gulch formation, the foliation which has an axial plane relationship to the folds, must have formed during the period of folding. The appearance of strongly foliated quartz porphyry at 400 N.–1,150 W. (pl. 5), west of the gabbro, and strongly foliated quartz porphyry southeast of the gabbro denies the possibility that the gabbro was a buttress and not susceptible to foliation. Instead, it shows conclusively that the nonfoliated gabbro separated the foliated quartz porphyry into two segments. Furthermore the gabbro is susceptible to foliation in the proper environment as shown by the foliated gabbro northwest of the Iron King mine (pl. 1).

Relict bedding in the Grapevine Gulch formation in the zone of mineral deposition is not abundant, but attitudes that can be measured are in harmony with the concept of a plunging north-northwest anticlinal fold where the massive sulfide pipe formed. On the 600 level, well within but near the west side of the massive sulfide pipe, bedding in unreplaced rocks of the Grapevine Gulch formation strikes N. 10° E. and dips 70° W. On the 900-foot level at the northeast corner of the massive sulfide, bedding within the sulfide pipe strikes N. 70° E. and dips 60° N. On the 1,000 level in the northern part of the massive sulfide pipe, some of the bedding in the tuffaceous rocks strikes nearly east and northeast (pl. 7). These more easterly strikes conform with the concept of a plunging anticline.

These data, though not conclusive, suggest that a drag fold on the flank of the major Mingus anticline controlled the form of the gabbro contact and, in turn, controlled the path of the later mineralizing solutions. The massive nonfoliated gabbro was undoubtedly less permeable to the mineralizing solutions than the underlying foliated quartz porphyry and Grapevine Gulch formation, and the "inverted trough" of the overlying gabbro channeled the solutions along the nose of the northward-plunging anticline.

An anticlinal structure localizing the massive sulfide pipe requires a complementary syncline to the east as

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the Grapevine Gulch formation and Deception rhyolite are in normal contact east of the mine. Furthermore, if the two formations form an anticline coinciding with the massive sulfide pipe, a bend of this contact to the southeast is needed to join the contact exposed on the 300-foot level bench (pl. 5). Confirmation of a small syncline to the east of the anticline is obtained on the 1,000 level (pl. 7, east of 500 E.) where sufficient under-

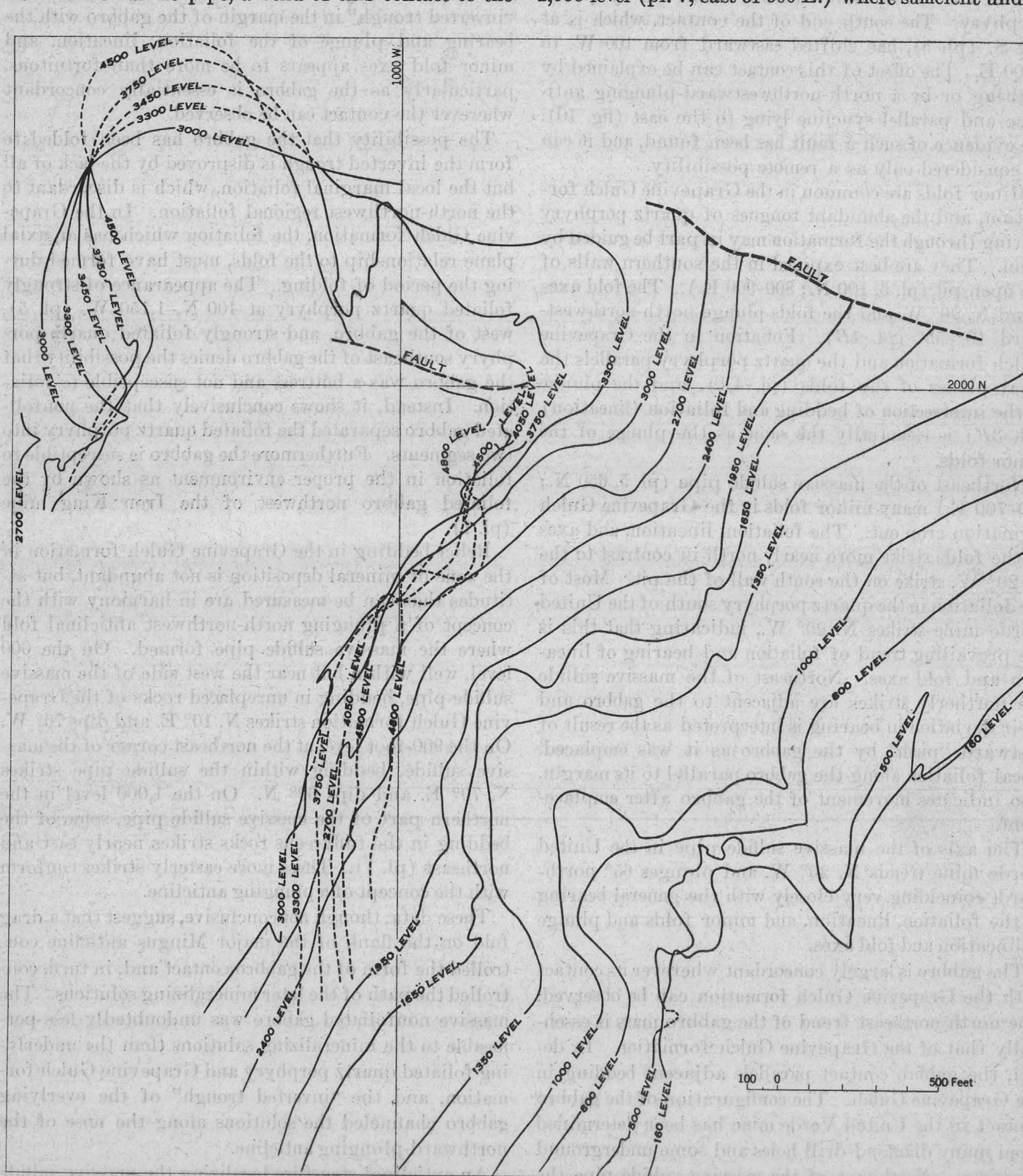
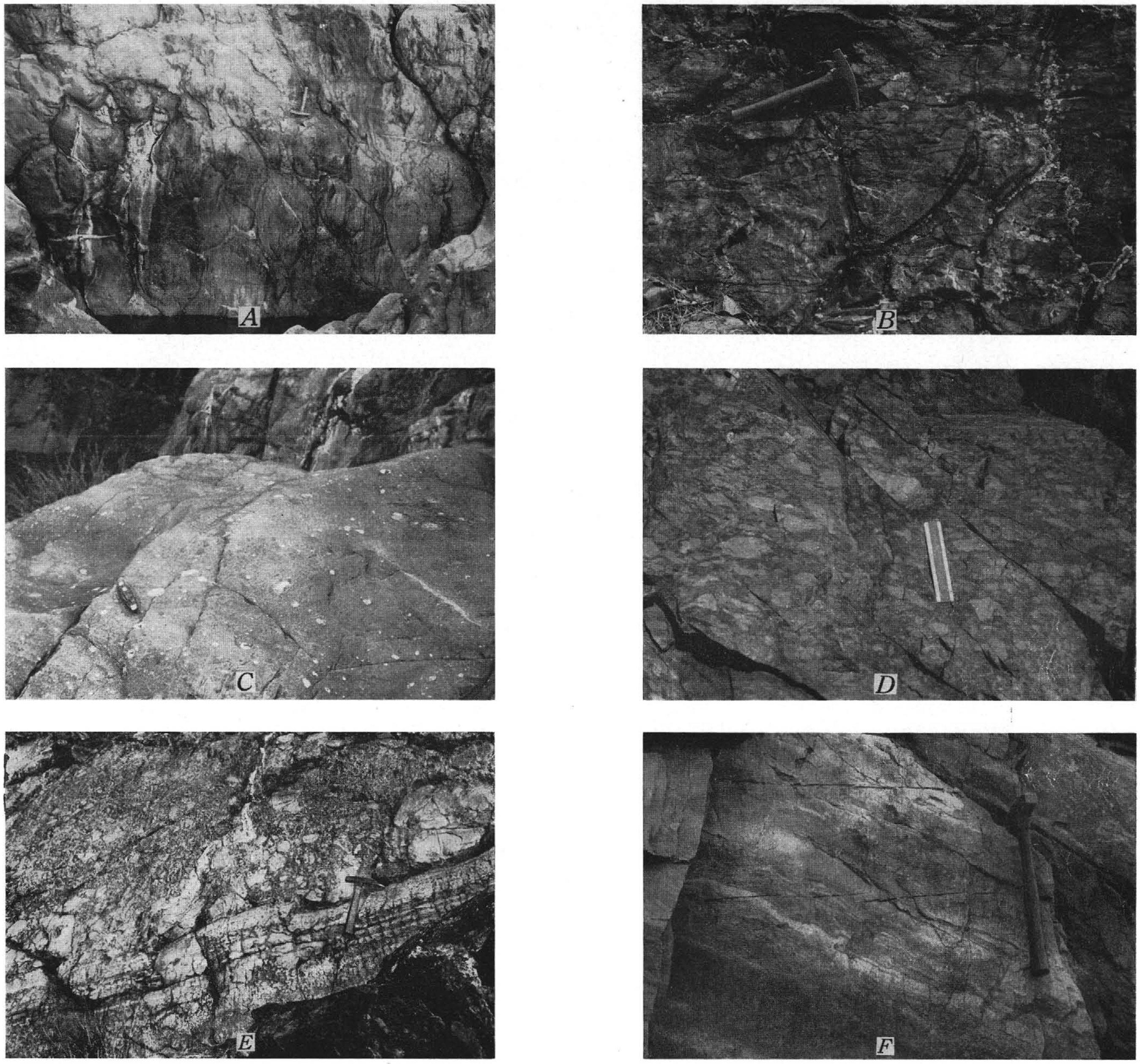
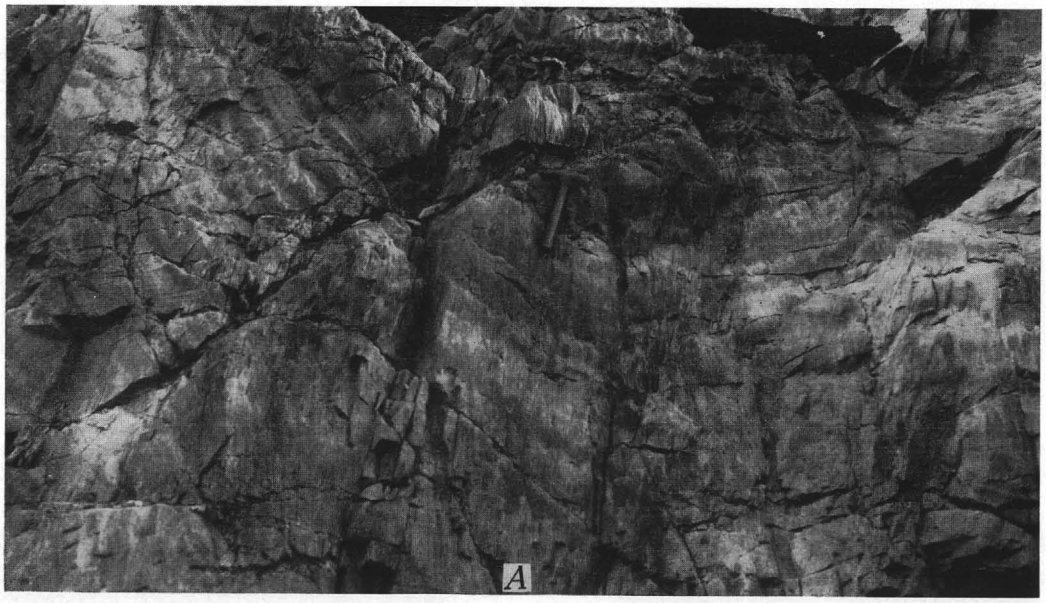


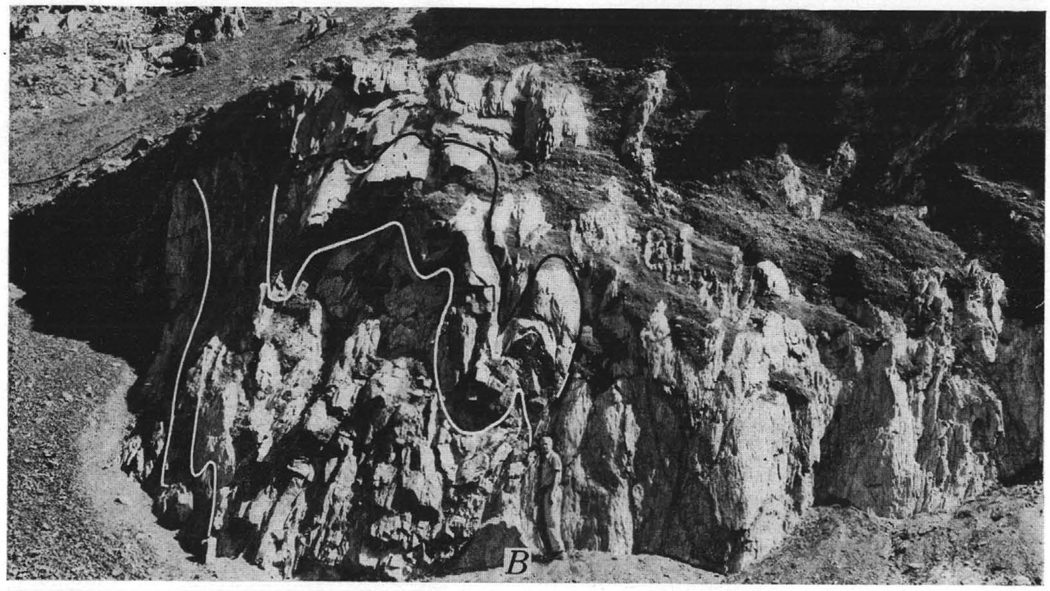
Figure 11.—Contour map of the contact between gabbro (above) and massive sulfide, quartz porphyry, and Grapevine Gulch formation (below), United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.



PHOTOGRAPHS OF ROCKS OF ASH CREEK GROUP

A. Pillow structure in Gaddes basalt. *B*. Pillow structures in Gaddes basalt showing fine-grained selvage on pillows (black). *C*. Quartz amygdules in Gaddes basalt. *D*. Rhyolitic breccia, Buzzard rhyolite. *E*. Coarse breccia resting on finely bedded tuffaceous rock, Deception rhyolite. *F*. Bedding exposed on foliation plane, Grapevine Gulch formation. Lineation dipping at low angle to right produced by intersection of bedding and foliation.





FOLDS IN GRAPEVINE GULCH FORMATION

A. Syncline in Grapevine Gulch formation showing foliation dipping steeply to the left (east) essentially parallel to the axial plane of the fold. Exposed in south wall of pit, United Verde mine. *B.* Minor folds in Grapevine Gulch formation plunging 40°–55° NNW. Exposed in south wall of pit, United Verde mine. Man at base of exposure gives scale.

ground minor structures reveal the major structure. In the absence of good marker beds, the syncline cannot be recognized at the surface owing to the presence of many minor folds.

The structure of the Grapevine Gulch formation west and north of the gabbro can only be conjectured because the formation is largely covered by dump derived from stripping for the open pit. The information for this area, shown on plate 5, is chiefly from maps prepared by L. E. Reber, Jr., formerly chief geologist for the United Verde mine. The rocks south of Haynes fault and west of the gabbro trend northward. The regional pattern indicates that the beds face west, and that probably the overturned beds or minor folds account for the local east dips. North of the Haynes fault, the beds trend northwestward, and probably face northeast, to form an anticline that may be the northwesterly extension of the one at the United Verde mine, or possibly another.

Foliation

Reference has been made earlier to the regional foliation, striking N. 20° W. and dipping 75°–80° E. Near the United Verde mine, foliation is well developed in the quartz porphyry and Grapevine Gulch formation, and is less pronounced but recognizable in the Deception rhyolite. Foliation appears to decrease in intensity to the east and is weak in the Grapevine Gulch formation north of the Haynes fault in contrast to the strong foliation in the same rocks south of the fault.

The intensity of this foliation is strongest in a northwestward-trending zone south of the open pit; the massive sulfide body lies at the north end of this zone, and the partly chloritized quartz porphyry coincides with it (pl. 5). The axis of the massive sulfide pipe plunges 65° N. along a bearing of N. 20° W. so that in depth, this eastward-dipping zone of intense foliation passes to the east of the center of the massive sulfide; and on the 2,250 level (pl. 7), the massive sulfide forms a south-southeastward-trending prong in the strongly foliated zone, indicating that the foliation was an important local control during mineralization. A minor control exerted by the foliation in zones of disseminated sulfide is revealed by the concentration of pyrite cubes in certain foliation planes. Where the disseminated pyrite can be observed on foliation planes, it forms a lineation whose northward plunge is about equal to that of the massive sulfide pipe.

A second and younger foliation, striking N. 15°–20° E. and dipping chiefly 75°–80° E., occurs along the footwall (south) side of the massive sulfide, and near the gabbro (pl. 7). Where both structures occur together, the northeastward-trending one cuts the northwest. Despite the southeast dip of the younger foliation, it does not extend downdip to lower levels; on each successive lower level, the zone of northeast foliation is close to the northwestward-plunging footwall of the massive sulfide pipe.

The origin of the northeast foliation is uncertain, but it may have some connection with the gabbro which is foliated weakly along and parallel to its east margin. A northeastward push exerted by the gabbro magma in the late stages of emplacement after partial crystallization may have produced the northeast foliation in the adjacent rocks, as well as bending the fold axes and foliation as observed on the surface.

Weak northwestward-trending foliation in some of the andesitic dikes parallels the stronger northwest foliation in the quartz porphyry host rock, which is probably caused by recurrent movement along these planes.

Faults

The United Verde mine is in a horst bounded by four normal faults: the Verde to the east, the Warrior to the west, the Hull to the south, and the Haynes to the north (pl. 5). The Hull and Verde faults are discussed in greater detail on page 70. All four faults displaced Paleozoic sedimentary rocks, and near the United Verde mine the Verde fault displaced the Tertiary Hickey formation.

The Haynes fault strikes east and dips steeply north, and extends across the northern part of the area to the Verde fault. The Warrior fault strikes north-northeast; and extends from the Haynes fault southwest for several miles. Both faults offset Paleozoic rocks; whether they were active during the Precambrian is uncertain, but the distribution of the Precambrian rocks on each side of the Hull fault indicates an earlier period of movement.

North of the Haynes fault, the divergent structural trend in the Grapevine Gulch formation from that to the south indicates greater displacement than the 200 feet of vertical separation shown by the Tapeats sandstone (?).

Many small faults trending westward or northwestward cut Paleozoic rocks in the northwestern part of the United Verde area (pl. 5), but none of these are known to have moved during the Precambrian.

SEQUENCE OF MINERALIZATION

Ore deposition at the United Verde mine was connected closely with special types of alteration, so that it is difficult to draw a clear-cut distinction between alteration and deposition of sulfide minerals. The evidence is convincing that in and near the United Verde mine, tuffaceous rocks of the Grapevine Gulch formation were silicified to masses of fine-grained quartz (fig. 12, *2*). This early silicification was followed by the replacement of Grapevine Gulch formation and quartz porphyry by massive pyrite forming a pipelike body (fig. 12, *3*). Carbonate minerals and quartz are part of the pyritic pipe; the carbonate minerals, as well as the pyrite, clearly represent partial alteration of the host rocks. The quartz may be residual in part from

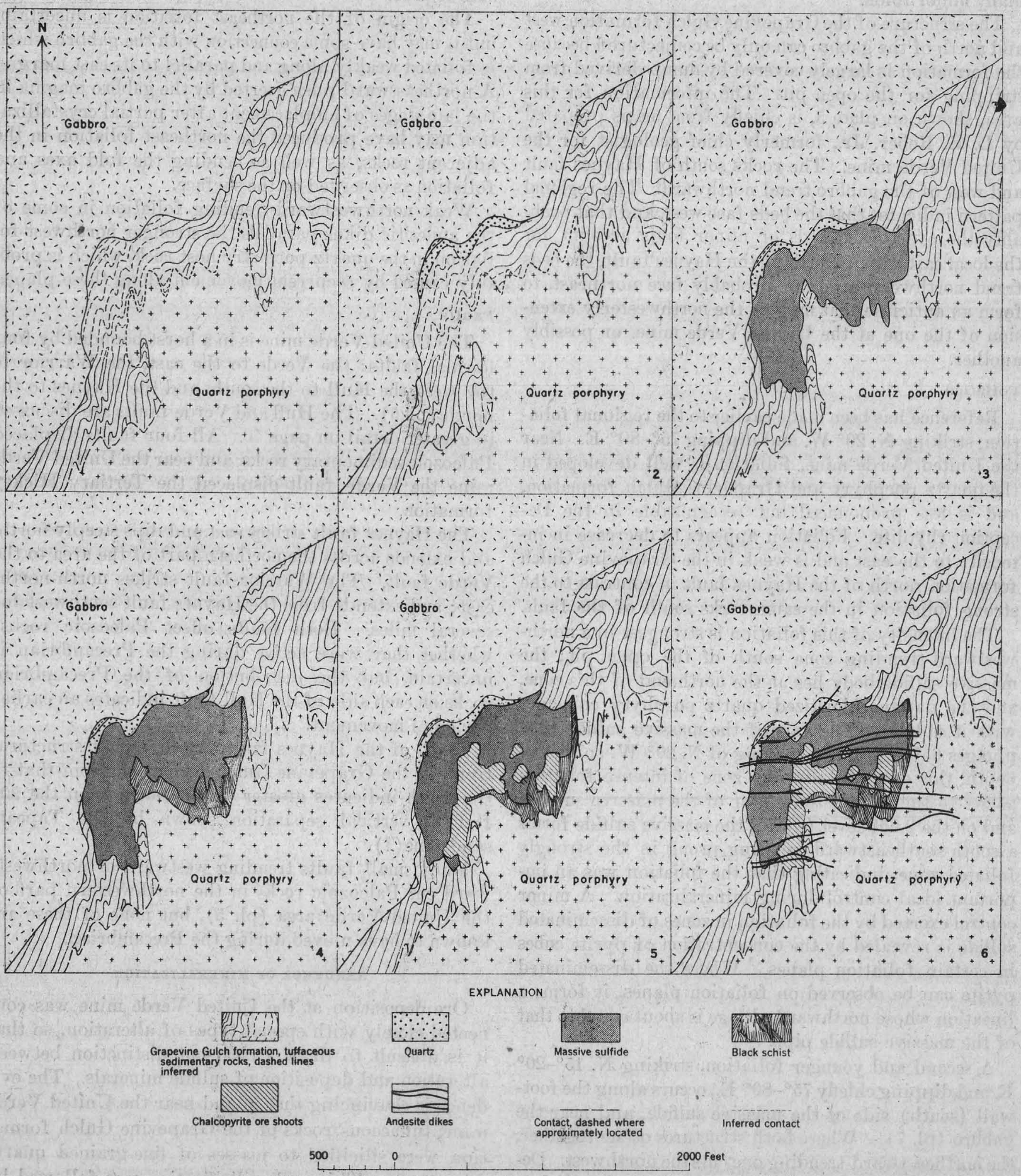


FIGURE 12.—Series of sketch maps to show sequence of mineralization, United Verde mine.

the host rock and redistributed during replacement. Extensive chloritization of the rocks on the footwall (south) side of the massive sulfide pipe was next in the sequence, overlapping somewhat with the main period of pyritic deposition (fig. 12, *4*). Quartz porphyry was the dominant rock chloritized to form black schist, but part of Grapevine Gulch formation was likewise chloritized. The main period of copper deposition followed the extensive chloritization (fig. 12, *5*). Quartz-carbonate veins and nodules were the final phase of mineralizing activity, containing some sulfide minerals. The andesitic dikes were intruded before the end of the quartz-carbonate veining and after the main period of copper deposition (fig. 12, *6*).

The separation of the products of dynamic metamorphism and widespread hydrothermal alteration is not easy, for each process resulted in the formation of sericite from the feldspar in the silicic rocks. Near Jerome the alkali content drops appreciably in the Deception rhyolite and the K_2O and MgO content increases in the quartz porphyry. To the south, the Na_2O content in the quartz porphyry is higher. These chemical data seem to indicate widespread hydrothermal activity, which near Jerome is revealed in the quartz porphyry and Deception rhyolite by a simple mineralogy of quartz and sericite, locally modified by clots and streaks of chlorite.

The writers believe that from south of the United Verde mine to Mescal Gulch, the quartz porphyry and Deception rhyolite have been hydrothermally altered after the regional metamorphism. In places the foliation is destroyed or so weakened, that it can be recognized only in favorable exposures. Near the United Verde mine, the Grapevine Gulch formation locally shows the same type of alteration.

Eastward, beyond the Verde fault, alteration is recognizable in the underground workings of the United Verde Extension mine near the ore body, but near the Audrey shaft (pl. 10) bright feldspar crystals indicate an absence of sericitic alteration.

The chloritization and copper mineralization in the United Verde ore body are clearly younger than the sericite in the sericitized and silicified quartz porphyry, but we are uncertain as to the relation between (1) the local early silicification and main period of pyritic mineralization and (2) the widespread sericitization.

SILICIFICATION

The process of silicification resulted in the formation of large and small masses of nearly pure quartz, limited largely to the hanging-wall side (northwest) of the massive sulfide pipe (pl. 7). The quartz masses in places form a screen between the massive sulfide and overhanging gabbro. Information on the distribution of the quartz has been obtained chiefly from diamond-drill holes, for it is only in the upper levels of the mine that drifts and crosscuts regularly reach into the hanging-wall side of the massive sulfide pipe. Some of these workings in the upper levels were accessible at the time of our study.

The quartz is fine grained, almost flinty, and is red, black, or white. Color variations are not sharp. The quartz bodies formed partly by replacement of tuffaceous sedimentary rocks of the Grapevine Gulch formation, and in favorable exposures, relict bedding and minor folds are sufficiently preserved to prove the replacement origin of the quartz. In other exposures, the quartz bodies grade into quartz porphyry in such a manner to suggest that the igneous rock has been replaced. Although the data are incomplete it appears that the tuffaceous rocks were more commonly replaced by quartz than the quartz porphyry.

The form of the quartz bodies is lenticular to irregular. In horizontal outline, some of the quartz bodies are 500 feet long and as much as 150 feet wide; others are only 10 feet wide but 200 feet long. Contacts between quartz and massive sulfide are irregular, but between quartz and gabbro, they are even. The large quartz masses formed adjacent to the gabbro in local structural terraces in the gabbro contact (fig. 13).

Quartz bodies occur well within the main body of massive sulfide; and local veins, nodules, and streaks of quartz are common. Some of the quartz veins appear to cut the massive sulfide, but many of the nodules and streaks and larger bodies of quartz are cut by veins of pyrite that project into the quartz from the massive sulfide. The larger masses of quartz on the hanging wall of the massive sulfide are also veined by fine-grained pyrite; and locally, brecciated quartz was cemented by pyrite. The quartz therefore was deposited before the pyrite that forms the massive sulfide pipe.

In the footwall (southeast) side of the massive sulfide, particularly on the lower levels, jasper nodules are common in the quartz porphyry. These nodules range from 1 inch to 2 feet in length, and range in shape from rounded to elliptical or irregular. Some truncate foliation in the quartz porphyry whereas others parallel it, or the foliation wraps around the nodules. The interior of some contains pyrite in disseminated grains and in aggregates. The jasper nodules may be either contemporaneous with the quartz masses in the hanging-wall side of the massive sulfide, or they may be older, or younger. The nodules definitely formed before the end of mineralization, as a few are cut by quartz-carbonate veins.

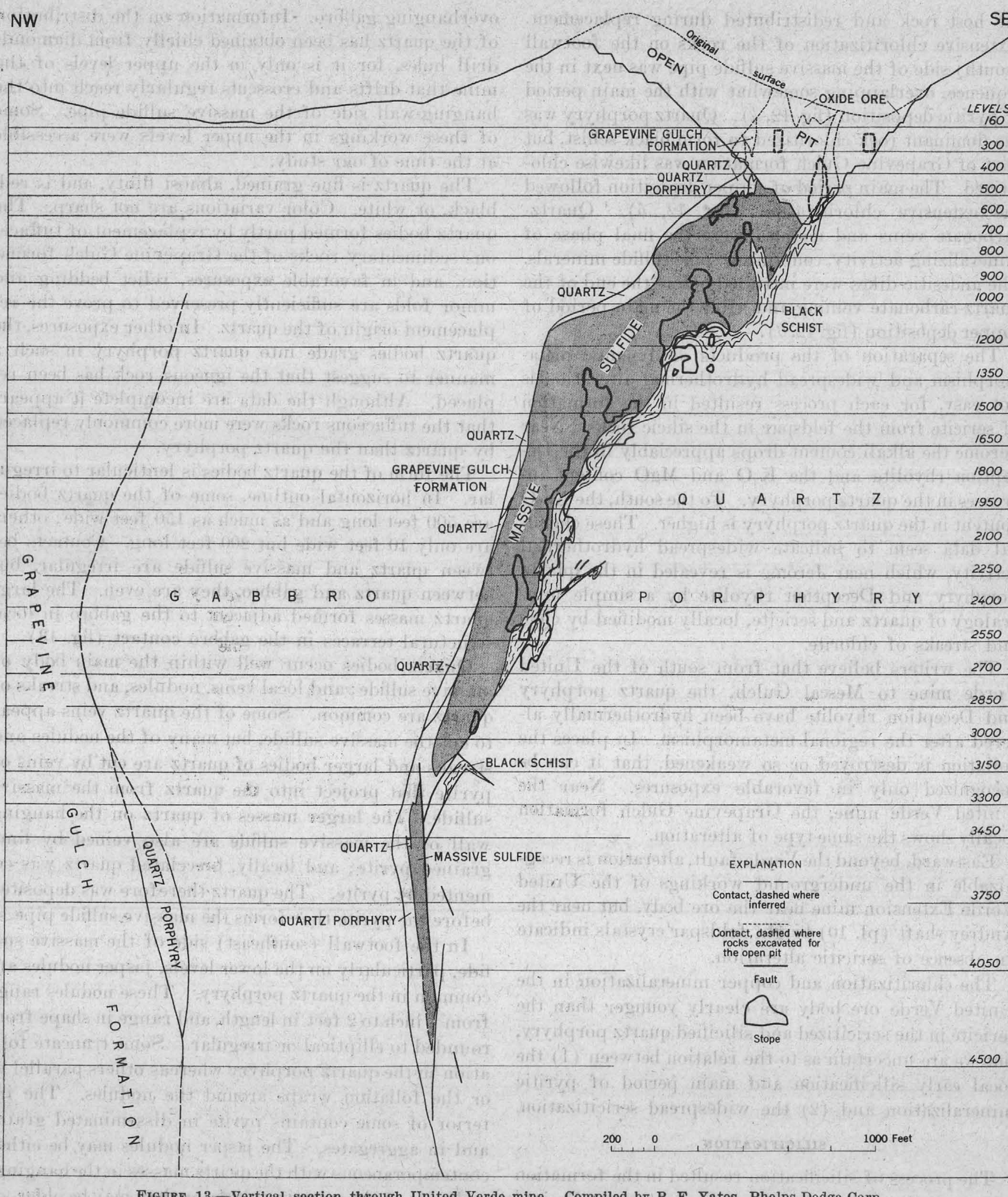


Figure 13.—Vertical section through United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.

MASSIVE SULFIDE PIPE

SIZE AND SHAPE

The large pipelike body of massive sulfide formed during the second period of mineralization by replacement of quartz porphyry and Grapevine Gulch formation by pyrite, quartz, and carbonate minerals. Sphalerite and chalcopyrite are present in minor amounts, and locally, sphalerite in significant quantity. The bearing of the axis of the massive sulfide pipe is about N. 20° W. and the plunge is about 65° N.

The variations in area of massive sulfide for each level are shown graphically by figure 14. From the surface to the 900 level, the horizontal area ranges from 220,000 to 260,000 square feet. Below the 900 level, it increases to about 500,000 square feet at the 1,500 level. Below this level, the horizontal area diminishes regularly to the 3,300 level, where it is about 60,000 square feet. Below the 3,300 level the horizontal area decreases at a smaller rate per level; and for all practical purposes, the massive sulfide bottoms a short distance below the 4,500 level.

The horizontal outline of the massive sulfide pipe varies on each level (fig. 15), and the variations are controlled by the gabbro contact on the hanging-wall (northwest) side of the massive sulfide and by a zone of intense foliation that controlled in part the eastern limb. The gabbro contact to the 3,000 level is essentially an inverted trough, plunging 65° N. and has a bearing of N. 20° W. The zone of intense foliation strikes N. 20° W. and dips 75°–80° E. Where these two structural elements converge on the 900 level and higher levels, the sulfide pipe is nearly equidimensional. At the 1,200 level the eastward-dipping zone of intense foliation diverges from the axis of inverted trough in the gabbro and the eastern part of the massive sulfide pipe is elongated in a south-southeast direction along the zone of intense foliation. This eastern limb dips east, parallel to the foliation (fig. 16) and away from the western part of the sulfide pipe adjacent to the gabbro. The western part of this pipe is also elongated in a south-southeast direction because of this trend in the gabbro contact. Along the west margin of the massive sulfide at successively lower levels, the gabbro contact trends more and more southward; on the 2,550 level, it trends southwestward and the western part of the massive sulfide has a similar trend. Because the eastern limb follows the zone of intense foliation, at this level the massive sulfide is crescent shaped horizontally. At greater depth, the two limbs diverge farther, and the massive sulfide narrows in the axial region of the sulfide pipe. On the 3,000 level, it is divided near the axis into two segments, and below the 3,000 level each limb of the arcuate pipe further divides into several segments. On the 3,450 level, the gabbro contact has a straight northeast trend, and the arcuate form of the massive sulfide has disappeared. Instead two small bodies of massive sulfide next to the gabbro are all that appear on the 3,450 level. Below this level, they break into segments and decrease in size so that on the 4,500 level, massive sulfide occurs only in 6 rootlike masses, the largest of which is 15 by 130 feet.

The gabbro, perhaps because it was more massive than the adjacent rocks, resisted mineralization and in the middle and upper levels channeled the solutions below the inverted trough into the adjacent quartz porphyry and Grapevine Gulch formation. Most of the massive sulfide formed in the zone above the 3,000 level where the gabbro dips west, and the inverted trough is present. Below the 3,000 level, the gabbro contact dips east, and massive sulfide abruptly diminishes. Thus on the intermediate levels where the western limb of the crescent body of massive sulfide is linear, the gabbro contact is the structural element to which the massive sulfide conforms.

In summary, the shape of the massive sulfide pipe is chiefly the result of three controls: (1) the overhang of the gabbro above the 3,000 level, (2) the contact between gabbro and the Grapevine Gulch formation or quartz porphyry, and (3) the zone of pronounced foliation. The shape on any given level is a function of the spatial relationship of these structural elements to each other.

RELATIONSHIP TO ADJACENT ROCKS

The contact of the massive sulfide with the adjacent rocks is generally sharp, but in many exposures it is gradational with disseminated pyritic rock. The massive sulfide is in contact with tuffaceous rocks of the Grapevine Gulch formation, quartz porphyry, early quartz masses, and black schist. Most of the black schist probably formed later than the pyritic massive sulfide.

The massive sulfide-early quartz contacts are best exposed in the upper levels of the mine on the hanging-wall side of the sulfide pipe; the relationship is discussed above.

The contact between the pyritic massive sulfide and tuffaceous rocks is best exposed also on the hanging-wall side of the sulfide pipe on the upper levels. Disseminated pyrite is present locally in some of the tuffaceous rocks. Lenses and tongues of massive sulfide as much as a foot wide are in the tuffaceous rocks, parallel to the stratification. In one locality, minor folds in the tuffaceous rocks are duplicated in the adjacent massive sulfide, clearly indicating replacement; at many others, banding in the margin of the massive sulfide parallels the stratification in the adjacent tuffaceous rocks, and in a few places, banding well within the

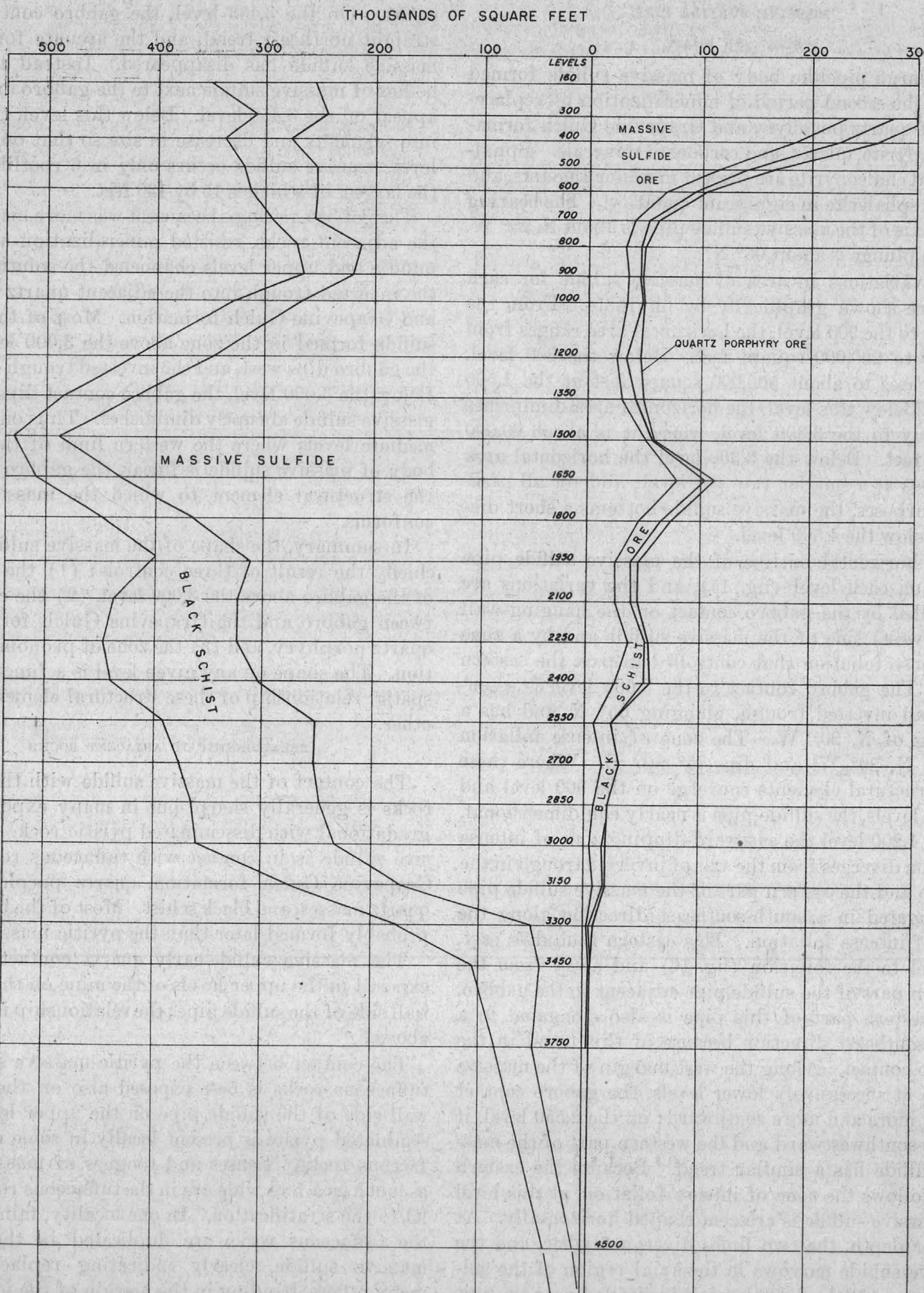


FIGURE 14.—Graphs showing relations of massive sulfide to black schist and types of ore in main ore body, United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.

massive sulfide parallels the bedding in the nearest (100 feet) exposure of tuffaceous rocks. This banding is interpreted as relict bedding following replacement of the sedimentary rocks by the pyritic massive sulfide.

Quartz porphyry generally is foliated where it is in contact with the massive sulfide. In many exposures, the massive sulfide contact is conformable with the adjacent foliation, but in many places, it cuts across the foliated quartz porphyry. Tongues of massive sulfide in part parallel the foliation, but some tongues crosscut it. On the 3,150 level, the quartz porphyry is brecciated, locally, at the contact with the massive sulfide, and pyrite is present between the quartz porphyry fragments. In a few exposures, the massive sulfide grades into pyritic disseminations in the quartz porphyry.

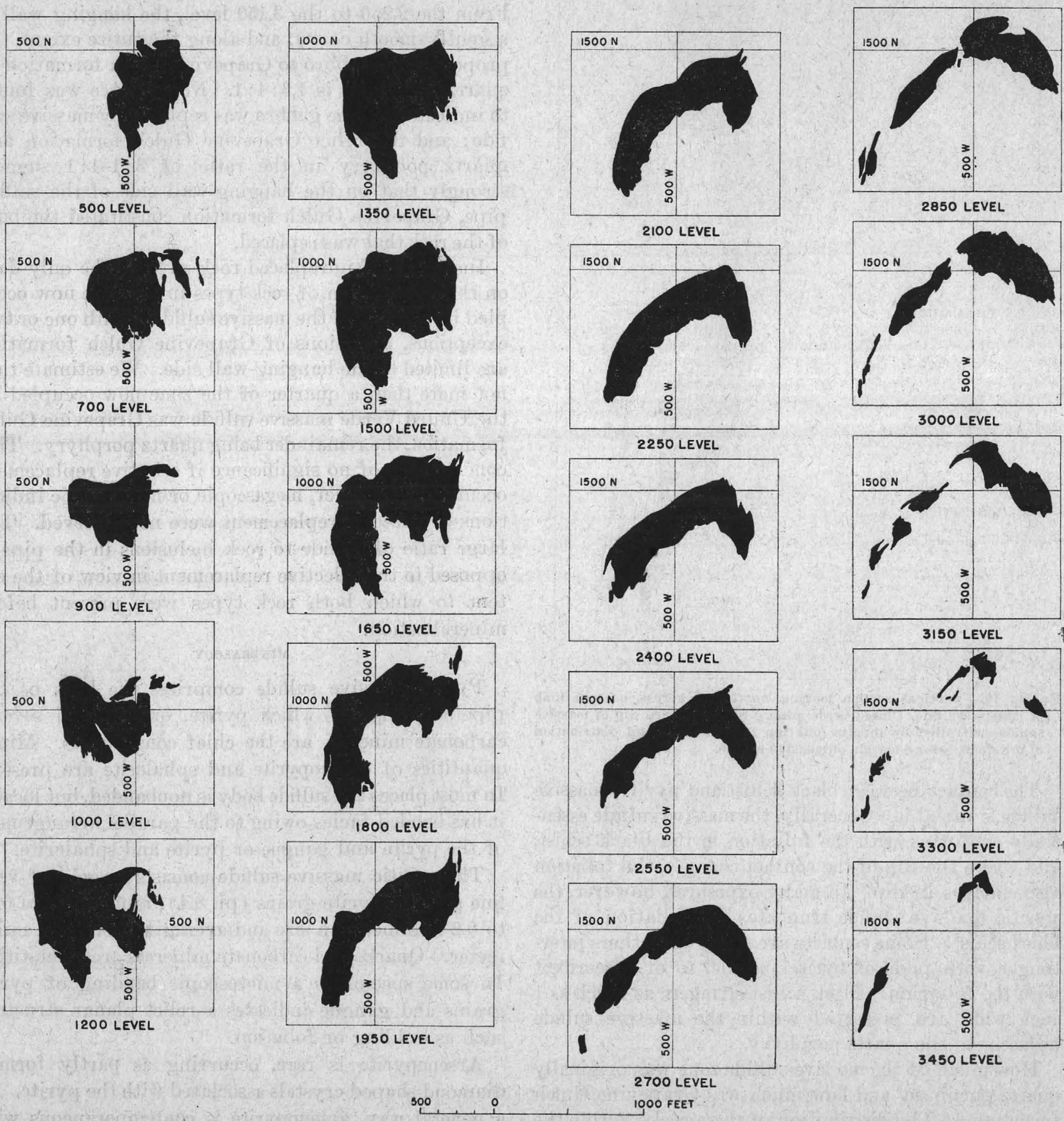


FIGURE 15.—Sketch map showing outline of massive sulfide and associated quartz on various levels, United Verde mine.

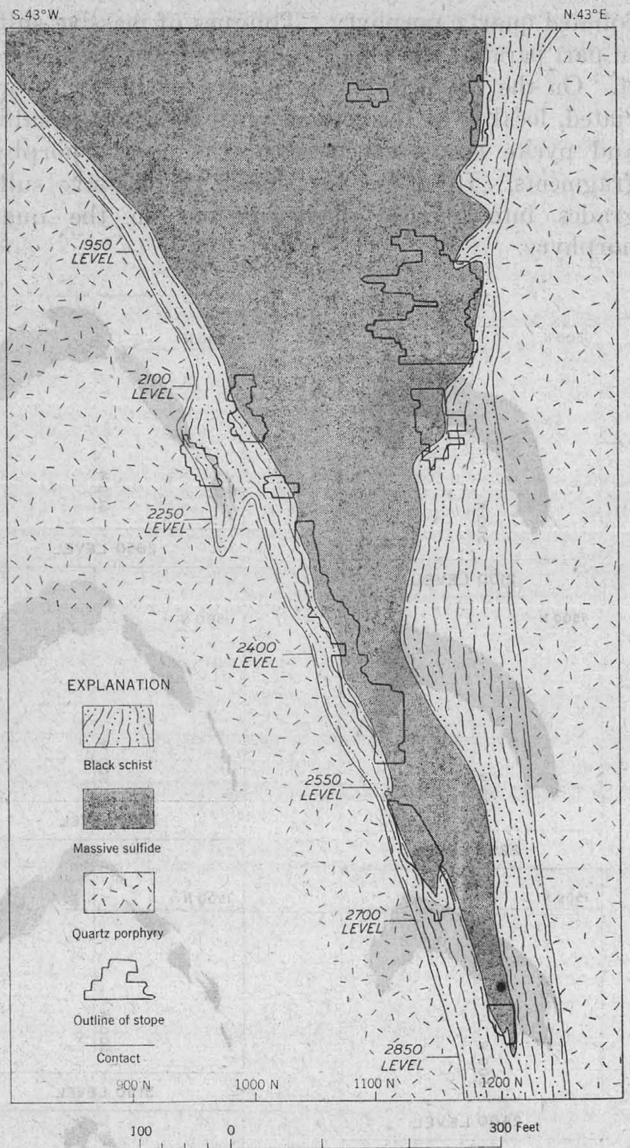


FIGURE 16.—Vertical section looking northward across eastern limb of massive sulfide, United Verde mine. Shows eastern dip of massive sulfide controlled by intense foliation in host rock, and distribution of ore shoots along margin of massive sulfide.

The contact between black schist and pyritic massive sulfide is variable. Generally, the massive sulfide essentially conforms with the foliation in the black schist, and where the dip of the contact changes, the foliation also changes in dip. In many exposures, however, the pyritic massive sulfide truncates the foliation of the black schist. Some contacts are sharp and others intertongue with pods of pyrite parallel to or discordant with the foliation. Black schist stringers as much as 1 inch wide are preserved within the massive sulfide adjacent to the quartz porphyry.

How much of the massive sulfide zone was originally quartz porphyry and how much was Grapevine Gulch formation? The distribution of these rocks within the massive sulfide pipe and adjacent to it provides the only evidence. Quartz porphyry forms the footwall, except for a little of the Grapevine Gulch formation along the southwestern part of the sulfide pipe on the 1,650 level (pl. 7). On the hanging-wall (north) side of this pipe, some approximations can be made. From the 600 to 2,100 level, the north side is fairly well defined; and gabbro, Grapevine Gulch formation, and quartz porphyry are in contact with it in the proportion, 5:3:1. From the 2,250 to the 3,150 level, the hanging wall is a gentle smooth curve; and along the entire extent, the proportion of gabbro to Grapevine Gulch formation to quartz porphyry, is 1.3:4:1. No evidence was found to indicate that the gabbro was replaced by massive sulfide; and the relict Grapevine Gulch formation and quartz porphyry in the ratio of 3:1–4:1 suggest strongly that on the hanging-wall side of the sulfide pipe, Grapevine Gulch formation constituted the bulk of the rock that was replaced.

Inclusions of unreplaced rock provide the only data on the distribution of rock types in the zone now occupied by the core of the massive sulfide. With one or two exceptions, inclusions of Grapevine Gulch formation are limited to the hanging-wall side. We estimate that not more than a quarter of the zone now occupied by the United Verde massive sulfide was Grapevine Gulch formation, the remainder being quartz porphyry. This conclusion is of no significance if selective replacement occurred. However, megascopic or microscopic indications of selective replacement were not observed. The large ratio of sulfide to rock inclusions in the pipe is opposed to the selective replacement in view of the extent to which both rock types were present before mineralization.

MINERALOGY

Pyritic massive sulfide comprises the bulk of the pipelike body, of which pyrite, quartz, and several carbonate minerals are the chief constituents. Minor quantities of chalcopyrite and sphalerite are present. In most places the sulfide body is nonbanded, but locally it has banded facies owing to the parallel arrangement of the pyrite and gangue or pyrite and sphalerite.

The pyritic massive sulfide consists largely of very fine grained pyrite grains (pl. 8*A*), ranging from 0.04 to 0.2 millimeter in size and averaging about 0.1 millimeter. Quartz and carbonate minerals are interstitial. In some specimens a microscopic banding of pyrite grains and gangue indicates a relict planar structure such as bedding or foliation.

Arsenopyrite is rare, occurring as partly formed diamond-shaped crystals associated with the pyrite. In a general way, arsenopyrite is contemporaneous with the pyrite, although individual grains of arsenopyrite may be slightly younger than contiguous pyrite.

The ratio of quartz gangue to the carbonate gangue varies, and in some specimens, the gangue is largely quartz whereas in others it is carbonate minerals. The quartz forms grains about the same size as the pyrite grains, and in part is in single irregular grains between pyrite or carbonate, but locally the quartz forms aggregates. In a few specimens, particularly where the pyrite has been fractured and healed by later quartz, the quartz consists of finely granular aggregates or finely columnar or fibrous crystals. Some of the columnar crystals are normal to faces of single pyritic crystals, but some are subparallel for short distances and show no relationship to individual pyritic crystals. The quartz may be residual from the replaced rock, except for veinlets that cut the pyrite, but the vein quartz may be redistributed original quartz.

The carbonate minerals form single crystals and aggregates, either pure or intergrown with quartz. The composition of carbonate minerals range from dolomite to ankerite, according to the range of the omega index of refraction, from 1.68 to 1.72.

Specimens of the pyritic massive sulfide taken from near the black schist contact on the footwall of the sulfide pipe contain streaks and nests of chlorite, associated with quartz and dolomite. The chlorite crystals are larger than those in typical black schist and in part are in subparallel arrangement but commonly are in random orientation and individual chlorite crystals penetrate pyrite. This relationship gives no clue to relative age: the pyrite and chlorite may be contemporaneous; earlier pyrite may be partly replaced by chlorite, or older chlorite may have been inert to the pyritic solutions.

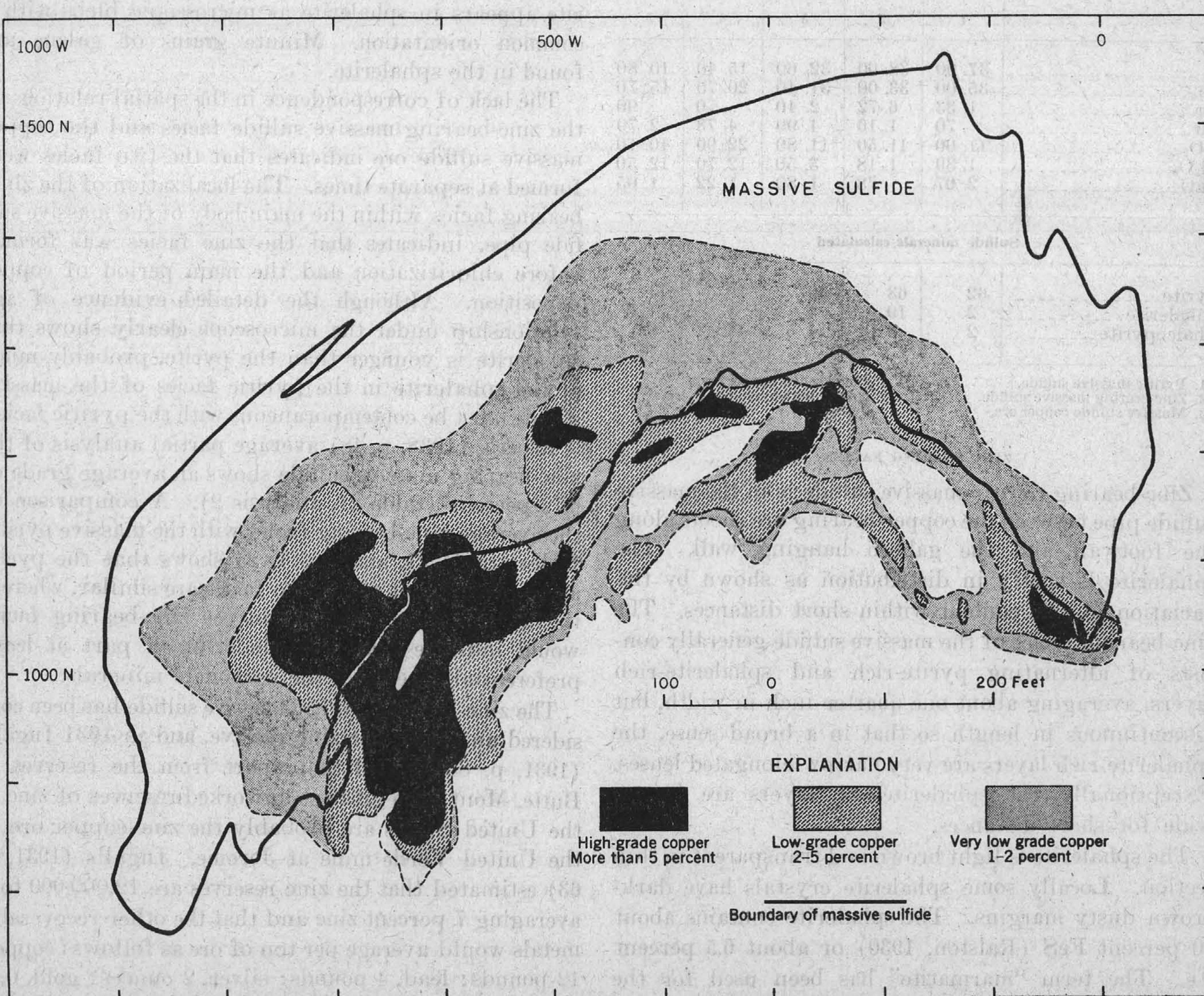


FIGURE 17.—Sketch map showing distribution of copper on a typical level, United Verde mine. Compiled by P. F. Yates, Phelps Dodge Corp.

Most polished surfaces of the pyritic massive sulfide show small amounts of sphalerite and chalcopyrite interstitial to the pyrite. Possibly some copper and zinc were deposited with the pyritic massive sulfide, but the distribution of copper (fig. 17) is spatially related to the footwall side of the pipe, and the copper content decreases toward the hanging-wall (north) side, indicating that the chalcopyrite in the pyritic facies may be younger than the formation of the pyritic massive sulfide pipe.

Reber (1938, p. 48) has given an average partial analysis of the pyritic massive sulfide (table 21, analysis 1), which shows that sulfide minerals average only 66 percent, the remainder consists of quartz, dolomite and ankerite, and locally, chlorite.

TABLE 21.—*Average partial analyses of United Verde massive sulfide and types of copper ores, in percent, from Reber (1938)*

	1	2	3	4	5
S	37.00	38.00	32.60	15.40	10.80
Fe	35.00	33.00	31.40	20.70	15.70
Zn	1.33	6.72	2.40	.70	.90
Cu	.70	1.16	4.99	4.73	2.79
SiO_2	11.00	11.50	11.80	22.90	40.10
Al_2O_3	1.39	1.18	3.50	12.70	12.50
CaO	2.07	.78	1.80	1.22	1.05
Sulfide minerals calculated					
Pyrite	62	63	50	19	14
Sphalerite	2	10	4	1	1.3
Chalcopyrite	2	3	14.5	14	8

1. Pyritic massive sulfide.
2. Zinc-bearing massive sulfide.
3. Massive sulfide copper ore.
4. Black schist copper ore.
5. Quartz porphyry copper ore.

ZINC-BEARING FACIES

Zinc-bearing pyritic massive sulfide is in the massive sulfide pipe between the copper-bearing ore shoots along the footwall and the gabbro hanging wall. The sphalerite is spotty in distribution as shown by the variation in zinc content within short distances. The zinc-bearing facies of the massive sulfide generally consists of alternating pyrite-rich and sphalerite-rich layers, averaging about one quarter inch in width, but discontinuous in length so that in a broad sense, the sphalerite-rich layers are very narrow elongated lenses. Exceptionally, the sphalerite-rich layers are 1 inch wide for short distances.

The sphalerite is light brown and transparent in thin section. Locally some sphalerite crystals have dark-brown dusty margins. The sphalerite contains about 10 percent FeS (Ralston, 1930) or about 6.5 percent Fe. The term "marmatite" has been used for the United Verde zinc sulfide, but Palache, Berman, and Frondell (1944, p. 2122) have defined marmatite as containing more than 10 percent of Fe.

Microscopic studies of polished surfaces show that generally the pyrite is older than the sphalerite, for sphalerite veinlets cut fractured pyrite. On some polished surfaces, the fine-grained pyrite crystals are preferentially oriented, arranged in chainlike trains, and separated by quartz-carbonate gangue, also arranged parallel to the pyrite. Presumably this preferred orientation of pyrite crystals is a relict structure after foliation or bedding. The sphalerite veinlets truncate this older structure at several angles. In places, the sphalerite veinlets contain a gangue of quartz and carbonate. In other specimens, sphalerite is interstitial to pyrite, indicating that the original gangue minerals were replaced.

Chalcopyrite generally forms irregular grains surrounded by sphalerite, but in a few places the chalcopyrite appears in sphalerite as microscopic blebs with a common orientation. Minute grains of galena are found in the sphalerite.

The lack of correspondence in the spatial relations of the zinc-bearing massive sulfide facies and the copper massive sulfide ore indicates that the two facies were formed at separate times. The localization of the zinc-bearing facies within the main body of the massive sulfide pipe, indicates that the zinc facies was formed before chloritization and the main period of copper deposition. Although the detailed evidence of age relationship under the microscope clearly shows that sphalerite is younger than the pyrite, probably much of the sphalerite in the pyritic facies of the massive sulfide must be contemporaneous with the pyritic facies.

Reber's (1938, p. 48) average partial analysis of the zinc-bearing massive sulfide shows an average grade of 6.72 percent (table 21, analysis 2). A comparison of the average zinc-bearing facies with the massive pyritic facies (table 21, analyses 1, 2) shows that the pyrite and silica content of the two facies are similar, whereas the lower content of CaO in the zinc-bearing facies would indicate that the sphalerite, in part at least, preferentially replaces the carbonate minerals.

The zinc-bearing pyritic massive sulfide has been considered an important zinc reserve, and in 1931 Ingalls (1931, p. 83) stated that apart from the reserves at Butte, Mont., the greatest unworked reserves of zinc in the United States are probably the zinc-copper ore of the United Verde mine at Jerome. Ingalls (1931, p. 63) estimated that the zinc reserves are 12,000,000 tons averaging 7 percent zinc and that the other recoverable metals would average per ton of ore as follows: copper, 12 pounds; lead, 4 pounds; silver, 2 ounces; gold, 0.02 ounce.

In 1930, serious attention was given to the recovery of the zinc from the massive sulfide (Ralston, 1930a) but the small size of the pyrite, sphalerite, and chalcopyrite grains requires extremely fine grinding for separation; on the average the ore must be ground to pass 800-mesh to obtain more than 90 percent liberation of copper and zinc minerals from the pyrite. As 300-mesh is about the commercial limit for grinding, the grains in part consist of aggregates of sphalerite and chalcopyrite. Tests made by Ralston produced zinc concentrates containing 45 percent zinc and 2 percent copper. The copper concentrate from the zinc-bearing massive sulfide contained about 15 percent copper and from 8 to 10 percent zinc.

According to Yates [6] the development of the zinc-bearing massive sulfide disclosed that it was irregular and discontinuous and considerable zinc-bearing massive sulfide was removed in the open-pit operation. The remainder of the best grade of zinc in the massive sulfide is between the gabbro hanging wall and filled stopes and along the footwall side. Much of the zinc-bearing massive sulfide is now slowly subsiding owing to compaction of the fill in the stopes, and any plans for future mining of the zinc must take into account this mine subsidence.

Probably the estimate of grade and tonnage of the zinc by Ingall is optimistic, but no accurate figure can be determined because of the scarcity of data to block out irregular and discontinuous zinc ore shoots. But it is true that the reserves of zinc-bearing massive sulfide are large. When it is economic to mine 4–5 percent zinc in massive sulfide, considering the fine intergrowth of zinc, copper, and pyrite, and the mine subsidence, the zinc-bearing massive sulfide in the United Verde mine will receive serious consideration.

BLACK SCHIST

ORIGIN

Large masses of nearly pure chloritic rock are on the footwall and locally east of the massive sulfide pipe. The chloritic rock is greenish black to black and has been called "black schist" by the mine operators. The black schist largely formed from quartz porphyry, but on the upper levels on the east side of the massive sulfide pipe, tuffaceous rocks north, east, and south of black schist indicate that considerable Grapevine Gulch formation was replaced by chlorite.

In the large masses of black schist where all relict minerals and textures are gone, it is impossible to determine the original rock with certainty, but the unreplaced rock at the margins provides some information. On the fringes of the masses of the black schist, unreplaced rock is cut by fractures containing chlorite, and at fracture intersections, small masses of black schist are present. Where quartz porphyry is the replaced rock, relict quartz phenocrysts in a base of chlorite are present several inches from the unreplaced quartz porphyry. Thin sections reveal that the quartz phenocrysts have ragged edges and are partly veined by chlorite. Near the margins of the black schist, the groundmass of the quartz porphyry consists of quartz granules and chlorite indicating that sericite and feldspar are the first minerals in the porphyry to be replaced by chlorite.

[6] Yates, P. F., 1946, Bottoming of the United Verde sulphide pipe: Mimeographed paper distributed at meeting of Arizona section, Am. Inst. Min. Metall. Eng., Tucson, Ariz., October 1946.

STRUCTURE AND MINERALOGY

The black schist, generally, is poorly foliated and breaks with irregular surfaces parallel to the foliation. Some blocks are massive and others splintery. Microscopic studies show that the black schist consists of a very fine grained aggregate of pale-green chlorite, in places having a preferred orientation and in other places, forming a mat. The diameter of the individual crystal ranges from 0.002 to 0.01 millimeter. Locally, chlorite crystals from 0.02 to 0.5 millimeter in diameter form aggregates, commonly divergent from the foliation.

Zircon crystals, ranging from 0.001 to 0.002 millimeter in length, are commonly surrounded by pleochroic haloes as large as 0.006 millimeter across. A few larger zircon crystals, 0.07 millimeter long, have pleochroic haloes 0.01 millimeter across. Rutile crystals are common in the black schist, 0.01 millimeter long, and arranged in clusters or in trains of disconnected granules. Blue tourmaline was observed in one thin section. Ankerite in individual crystals and in aggregates is in some thin sections. Quartz grains are rare in the solid masses of black schist except as younger crosscutting veinlets.

CHEMICAL COMPOSITION

The high chlorite content of copper ore associated with black schist added to smelter difficulties which necessitated many complete and partial chemical analyses by the United Verde smelter analysts. These chemical data have been made available for our study (table 22). Three composite samples of black schist were carefully collected, two from the 2,400 level and one from the 3,000 level of the mine. A sample of quartz porphyry was collected on the 2,100 level, 1,600 feet south of the massive sulfide pipe. Using bulk specific gravity determinations, it is possible to calculate the amount of milligrams per cubic centimeter of the main analyzed constituents, and these are plotted graphically (fig. 18) to show the appreciable loss in SiO_2 and CaO and marked gain in Al_2O_3, FeO, MgO, and H_2O in the conversion of quartz porphyry to black schist. This graph

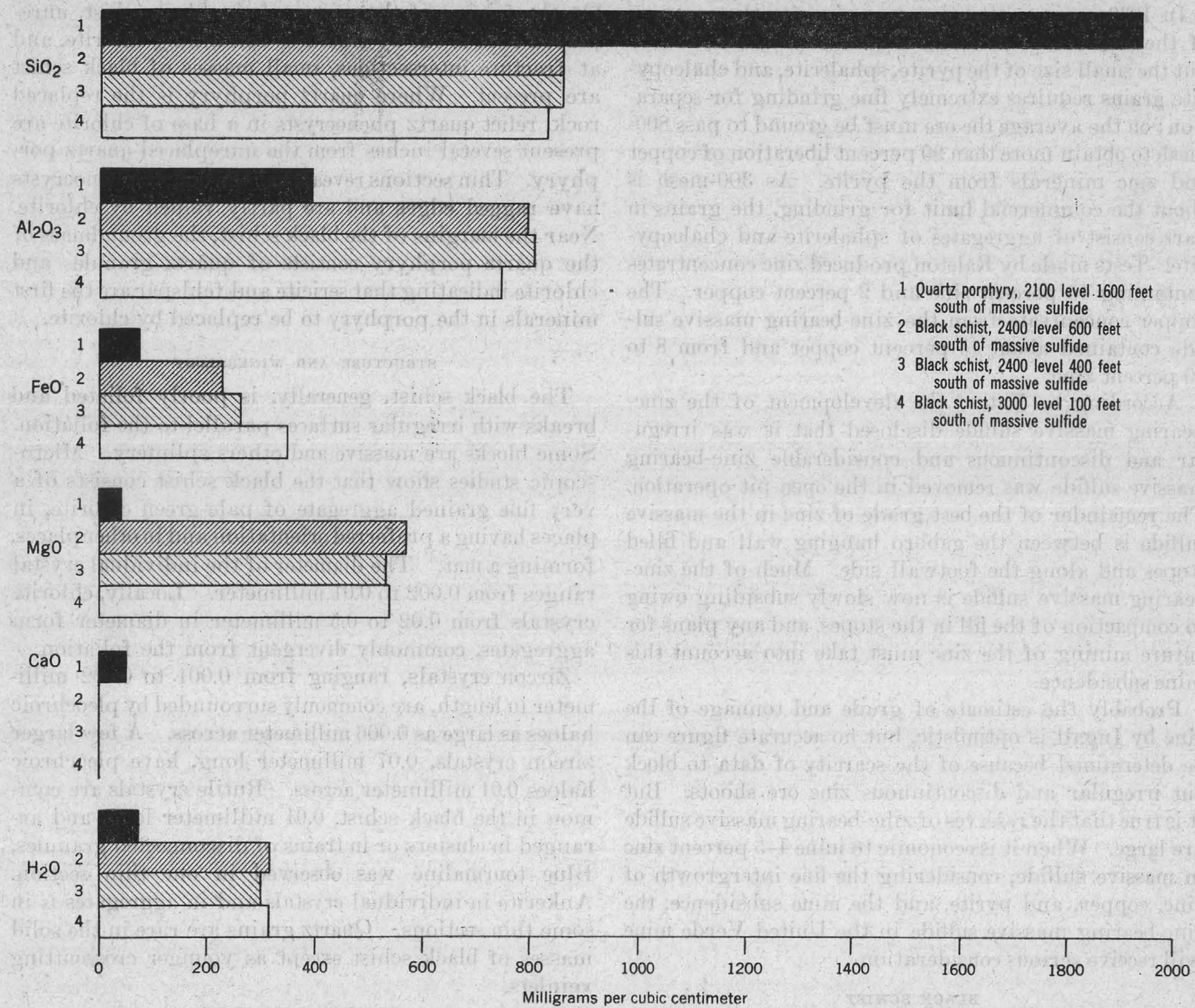


FIGURE 18.—Graph showing losses and gains of principal constituents in the formation of black schist from quartz porphyry.

also indicates that the iron content in the black schist increases toward the massive sulfide.

There is some variation in the ratio of SiO_2, Al_2O_3, FeO, and MgO in the black schist, and these variations are shown graphically in figure 19 by means of Niggli's *al* and *si* values, and *mg* ratios (Grubenmann and Niggli, 1924, p. 29). These values are obtained by calculating the molecular ratios of SiO_2, Al_2O_3, FeO, and MgO from the weight percentage of the chemical analyses. The molecular ratios of Al_2O_3, FeO, and MgO are recalculated on the basis of 100 percent so that an *al* value of 26 indicates that the Al_2O_3 content is 26 percent of the total Al_2O_3, FeO, and MgO molecules. The *si* value is the percent of the molecular ratio of SiO_2 to the total Al_2O_3, FeO, and MgO molecular ratios. The *mg* value is the molecular ratio of MgO to total FeO and MgO, which means that for *mg* in excess of 0.5, MgO is in excess of FeO.

The *al* and *mg* values are plotted against the *si* value (fig. 19) and they show, in general, a decreased *al* value with decreased *si* value, and in most of the analyzed black schist from the United Verde mine, *mg* is in excess of 0.5, indicating that MgO is in excess of FeO.

CHLORITE MINERALS

The variations in the chemical composition of the black schist are due in part to the variety of chlorite forming the rock, and in part, to variations in amounts of constituents such as quartz and carbonate. The differences in composition of the chlorite are most significant, because small amounts of accessory minerals are normally present. As discussed in the section on

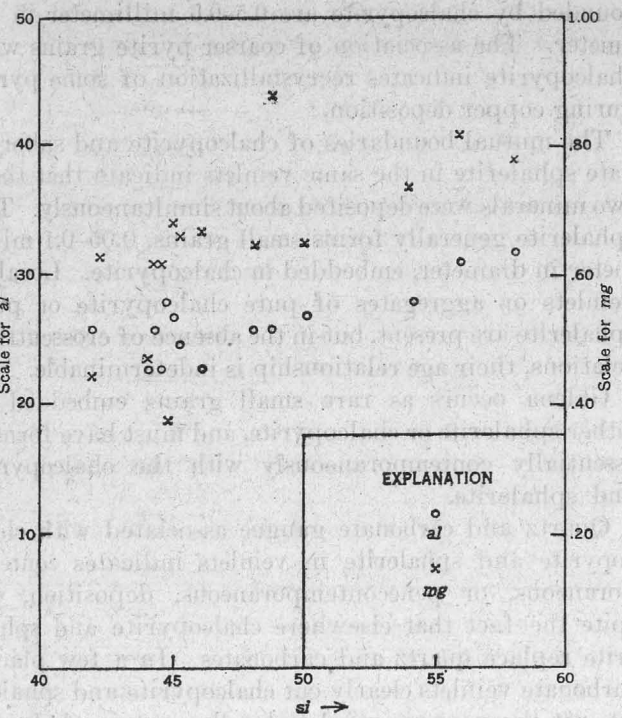


FIGURE 19.—Diagram showing Niggli values for analyzed black schist, United Verde mine.

TABLE 22.—*Chemical analyses of quartz porphyry and black schist, United Verde mine, in percent*

[Analyses furnished by Phelps Dodge Corp.]

	1	2	3	4
SiO_2	72.08	30.88	30.92	29.71
Al_2O_3	14.57	28.77	29.30	26.66
FeO [1]	2.86	8.42	9.44	12.45
MgO	1.48	20.64	19.36	19.42
CaO	1.98	Tr.	.04	.03
H_2O	2.67	11.44	10.06	11.28
CO_2	1.35	----	----	----
S	.06	.02	.01	.55
Cu	.02	Tr.	Tr.	.04
Zn	.04	.14	.12	.27
Total	97.11	100.31	99.25	100.41
Specific gravity	[2] 2.70	2.78	2.78	2.82

[1] Iron determined as Fe, calculated as FeO.
[2] Average of 4 samples of quartz porphyry collected in the mine.

1. Quartz porphyry, 2,100 level, 1,600 feet south of massive sulfide.
2. Black schist, 2,400 level, 600 feet south of massive sulfide.
3. Black schist, 2,400 level, 400 feet south of massive sulfide.
4. Black schist, 3,000 level, 100 feet south of massive sulfide.

mineralogy, p. 93, the varieties of chlorite that have been recognized in the black schist in the United Verde mine are: clinochlore, prochlorite, diabantite, ripidolite, and brunsvigite. Most of the black schist is composed of clinochlore-prochlorite, but even with these varieties, the ratio of MgO to FeO varies appreciably.

Where veinlets of quartz, carbonate, or sulfide, or several of these minerals, cut the black schist, chlorite flakes adjacent to the veinlets are from 10 to 20 times larger than the fine-grained chlorite in the black schist. The optical properties of the coarser crystals are identical with those of the adjacent chlorite, indicating similar composition and mineralogy. Along some of the quartz-carbonate veins, pale-green chlorite crystals 1 millimeter or more in diameter are arranged normal to the vein walls. These flakes are of the same mineralogical variety as the adjacent host, and in most places are clinochlore or prochlorite.

RELATIONSHIP TO MASSIVE SULFIDE

The age relationship of the black schist to the massive sulfide presents a problem that cannot be answered easily. Reber (1922, p. 16) in his first report on the Jerome district, suggested that the black schist was present before the pyritic sulfide was deposited, but later (Reber, 1938, p. 45), he stated that the period of chlorite deposition marked a definite break in sulfide deposition, and that chlorite was deposited after the massive sulfide pipe. Lindgren (1926, p. 33) recognized that the chlorite preceded the introduction of chalcopyrite, but he concluded that the chlorite accompanied the sulfide in many places.

That the pyritic massive sulfide pipe formed before the black schist is shown by the unreplaced quartz porphyry and tuffaceous rocks (see 600 level, pl. 7) within the massive sulfide body and between the north side of the massive sulfide pipe and the gabbro. These unreplaced rocks within the massive sulfide are more common in the upper levels where the pipe is nearly circular.

Limitation of black schist to the footwall side of the massive sulfide pipe indicates that the solutions that added FeO, MgO, and Al_2O_3 and removed SiO_2 were restricted to the footwall. The distribution of the black schist clearly reveals replacement of quartz porphyry and tuffaceous rocks along favorable structures, chiefly foliated zones and fractures.

Along the margin of massive sulfide and black schist, pyrite has been deposited on foliation planes and fractures in the black schist. Wisps of unreplaced black schist are present within the outer several feet of massive sulfide. However, no chlorite is present toward the center of the massive sulfide pipe. In a few specimens from the margin of the massive sulfide, the transition from a chloritic gangue to a normal quartz-dolomite gangue occurs within the field of a single thin section. The chlorite in the massive sulfide is a little coarser than in most of the black schist, and is randomly oriented. This lends support to the belief that there may have been a slight overlap between the periods of pyritic and chloritic replacement.

Chlorite was observed in a few exposures underground on the hanging-wall side of the massive sulfide pipe, but no masses of black schist were found. On the 600 level on the north side of the massive sulfide, chlorite was seen in two places: narrow seams of chlorite line fractures in the tuffaceous rocks, and at the contact of the massive sulfide, folded tuffaceous rocks were selectively replaced by pyrite and chlorite. Beds from 1 to 2 inches wide were replaced by pyrite and the intervening one-quarter inch beds were replaced by chlorite. The folded structure of the tuffaceous rocks is clearly preserved. The local chlorite on the hanging wall of the sulfide pipe shows either that some chlorite formed early, and perhaps contemporaneously with the early pyrite of the masive sulfide pipe, or that there was some leakage through the massive pyrite to the hanging-wall side during the main period of chlorite replacement.

COPPER MINERALIZATION

The introduction of copper and some zinc into the pyritic massive sulfide, black schist, and quartz porphyry followed the main period of chloritization, and most of the commercial copper ore formed during this period of mineralization. Intramineralization fracturing is indicated by the appearance of chalcopyrite in numerous intersecting veinlets. The spacing and distribution of these intersecting fractures at the time of the copper mineralization were fundamental in the formation of the copper ore shoots.

Three types of copper ore are present in the mine: massive sulfide, black schist, and quartz porphyry. The massive sulfide ore formed the largest and highest grade copper ore shoots and the quartz porphyry the smallest and lowest grade. Figure 14 graphically shows the ratio of these three types of ore. Table 21, analyses 3, 4, and 5, shows the average analyses.

MASSIVE SULFIDE ORE

Microscopic studies of the massive sulfide copper ore demonstrate that much of the chalcopyrite and associated sphalerite were introduced along fractures in the pyritic sulfide (pl. 8 *B*), a conclusion reached by Lindgren (1926, p. 73) and Rice (1920, p. 64). Irregular corroded margins of pyritic crystals indicate replacement of pyrite in the chalcopyrite-rich specimens, for the wider chalcopyrite bands contain very little pyrite. In part, however, chalcopyrite and associated sphalerite are interstitial to the small grains of pyrite, indicating that the replacement of the quartz-carbonate gangue was important. Recrystallization of pyrite is indicated because pyrite crystals lining chalcopyrite veinlets are as large as 0.6 millimeter in contrast to an average size of 0.1 millimeter in the main pyritic facies. In places, these coarser pyrite grains form a comblike structure, with chalcopyrite filling the spaces between. In a few of the chalcopyrite veinlets, the pyrite crystals surrounded by chalcopyrite are 0.5–0.6 millimeter in diameter. The association of coarser pyrite grains with chalcopyrite indicates recrystallization of some pyrite during copper deposition.

The mutual boundaries of chalcopyrite and subordinate sphalerite in the same veinlets indicate that these two minerals were deposited about simultaneously. The sphalerite generally forms small grains, 0.05–0.1 millimeter in diameter, embedded in chalcopyrite. Locally, veinlets or aggregates of pure chalcopyrite or pure sphalerite are present, but in the absence of crosscutting relations, their age relationship is indeterminable.

Galena occurs as rare small grains embedded in either sphalerite or chalcopyrite, and must have formed essentially contemporaneously with the chalcopyrite and sphalerite.

Quartz and carbonate gangue associated with chalcopyrite and sphalerite in veinlets indicates contemporaneous, or penecontemporaneous, deposition, despite the fact that elsewhere chalcopyrite and sphalerite replace quartz and carbonates. In a few places, carbonate veinlets clearly cut chalcopyrite and sphalerite, yet the younger veinlets locally contain chalcopyrite and a little sphalerite.

Bornite is present locally in the footwall side of the massive sulfide, associated with quartz. Bornite, chalcopyrite, and coarse-grained quartz form veinlets in fine-grained quartz. Early pyrite is fractured and partly replaced by chalcopyrite and bornite. Sphalerite is rare and appears older than the bornite.

Specular hematite is present sparingly in the massive sulfide ore, generally as narrow veinlets that contain very little carbonate gangue. Lindgren (1926, p. 73) stated that the specular hematite is younger than the pyrite but older than the other sulfide minerals. Examination of the few specimens of massive sulfide ore in which specular hematite occurs, shows that the specular hematite veins clearly cut chalcopyrite-sphalerite veinlets. Possibly specular hematite may have formed during several stages of mineralization, but because of the small amount, no positive statement can be made.

BLACK SCHIST ORE

The black schist copper ore is characterized by intersecting and branching veinlets of sulfide minerals, composed largely of chalcopyrite and pyrite (pl. 8*C*). In places, the chlorite has been extensively replaced by veins of almost solid sulfide 1 to 2 inches wide. The higher grade copper ore generally contains relicts of black schist between ramifying chalcopyrite veinlets.

Microscopic studies of the black schist ore confirm the importance of fractures in guiding the copper-bearing solutions in the black schist. Most sulfide veinlets