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Geology and Ore Deposits of the Jerome Area Yavapai County Arizona

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GEOLOGICAL SURVEY PROFESSIONAL PAPER 308

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

FROM C. A. ANDERSON

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1958

UNITED STATES DEPARTMENT OF THE INTERIOR

FRED A. SEATON, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

For sale by the Superintendent of Documents, U. S. Government Printing Office
Washington 25, D. C.

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

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

ABSTRACT

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

INTRODUCTION

LOCATION, CULTURE, AND ACCESSIBILITY

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

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

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

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

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

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

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

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

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

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

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

PHYSICAL FEATURES

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

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

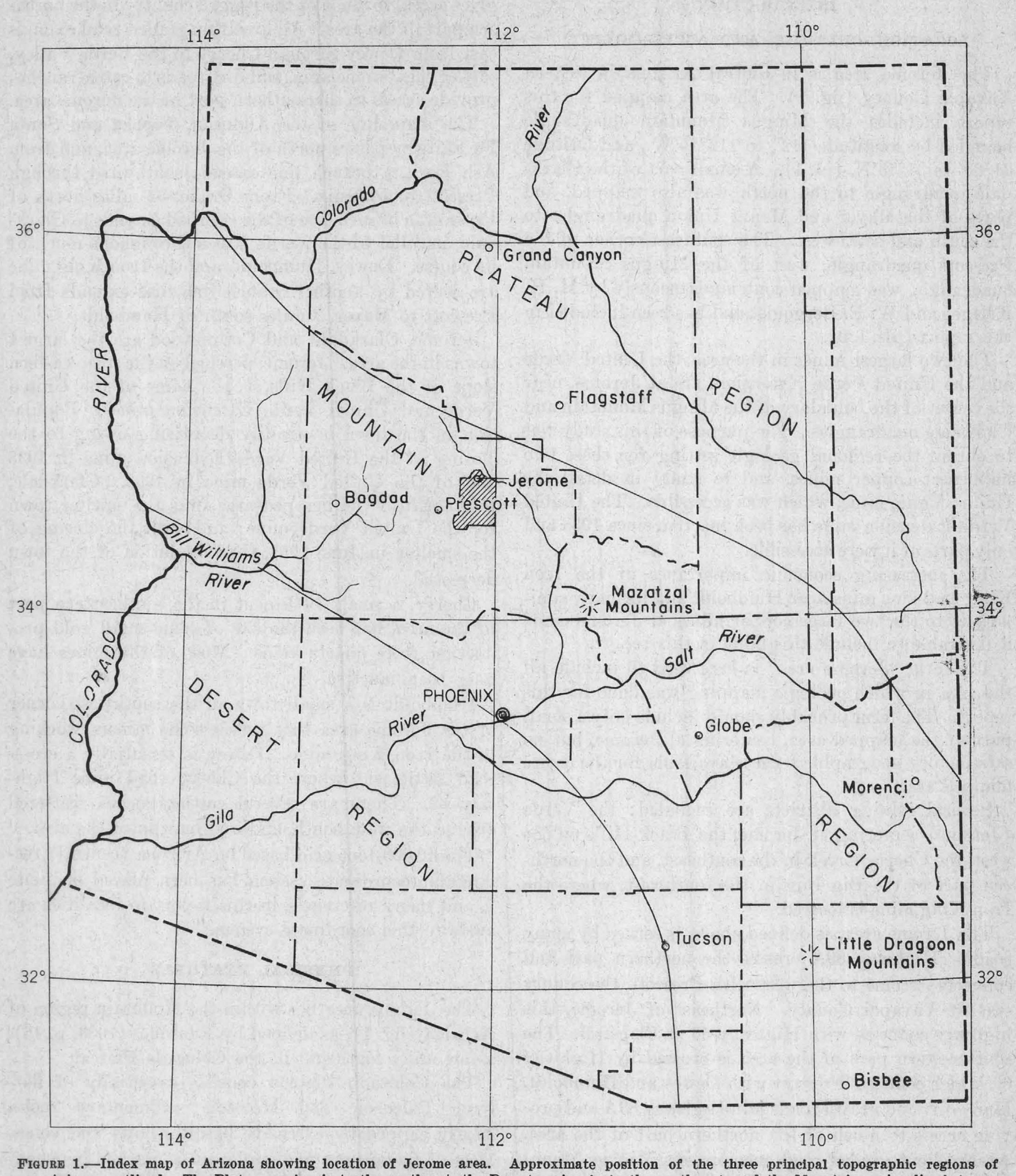

FIGURE 1.—Index map of Arizona showing location of Jerome area. Approximate position of the three principal topographic regions of Arizona outlined. The Plateau region is to the northeast, the Desert region is to the southwest, and the Mountain region lies between them. After Ransome (1903, p. 10).

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

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

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

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

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

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

CLIMATE AND VEGETATION

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

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

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

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

Precipitation averages (inches)

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

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

The vegetation (Nichol, 1937; Benson and Darrow, 1944) in the Jerome area may be grouped into three types: Pine forest, chaparral, and grassland. Pine forest is limited to altitudes above 6,500 feet, in the Black Hills, except for a small area in the Cherry Creek district. The ponderosa pine (*Pinus ponderosa*) dominates this forest, but in some breaks, Gambel's oak (*Quercus gambeli*) and Mexican locust (*Robinia neomexicana*) form small but compact thickets.

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

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

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

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

FIELD WORK AND ACKNOWLEDGMENTS

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

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

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

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

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

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

structural problems and for his critical review of the manuscript.

PREVIOUS WORK

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

1883. Blandy, J. F. The mining region around Prescott, Ariz.: Am. Inst. Min. Metall. Eng. Trans. v. 11, p. 286–291.

Contains a map showing distribution of granite, slate, and limestone. Location of mines of this date are of historical interest.

1905. Jaggar, T. A., and Palache, Charles. Description of Bradshaw Mountains quadrangle, Ariz.: U. S. Geol. Survey Geol. Atlas, folio 126, 11 p.

First adequate description of the Precambrian rocks in the region; only a small part of this quadrangle is included in the study of the Jerome area, but this folio was important pioneer work in the region.

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

Description of the gold veins in Cherry Creek mining district.

1916. Provot, F. A. Jerome mining district geology: Eng. and Min. Jour., v. 102, p. 1028–1031.

General discussion of the geology of the mines at Jerome.

1916. Ransome, F. L. Some Paleozoic sections in Arizona and their correlation: U. S. Geol. Survey Prof. Paper 98–K, p. 133–166.

Brief description of Paleozoic rocks in the Jerome area and faunal lists proving Devonian age of Martin limestone and Mississippian age of Redwall limestone. Correlates basal Paleozoic sandstone with Tapeats sandstone of Cambrian age in Grand Canyon section.

1918. Rickard, T. A. The story of the U. V. X. bonanza: Min. Sci. Press, v. 116, p. 9–17, 47–52.

Fascinating account of the history of mining at Jerome; Rickard states that the United Verde Extension ore body was not originally connected to the United Verde ore body.

1918. Finlay, J. R. The Jerome district of Arizona: Eng. and Min. Jour., v. 106, p. 557–562, 605–610.

First detailed description of the geology of the United Verde mine, Jerome.

1918. Tovote, W. L. Notes on certain ore deposits of the southwest: Am. Inst. Min. Metall. Eng., Bull. 142, p. 1599–1612.

Main point of this report is the objection to the theory that ore deposits at Jerome are of Precambrian age, contrary to the conclusion in our present report.

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

Good petrographic description of rocks and ores.

1922. Reber, L. E., Jr. Geology and ore deposits of the Jerome district: Am. Inst. Min. Metall. Eng. Trans., v. 66, p. 3–26.

Excellent description of local geology and ore deposits at Jerome, based on years of observation in the district. Precambrian rocks subdivided and adequately described. Replacement nature of ore deposits demonstrated and first complete description of the ore bodies and their control given.

1923. Jenkins, O. P. Verde River lake beds near Clarkdale, Ariz.: Am. Jour. Sci., 5th ser., v. 5, no. 25, p. 65–81.

First detailed description of the late Tertiary or early Quaternary lake beds found in the Verde Valley.

1925. Fearing, J. L., Jr., and Benedict, P. C. Geology of the Verde Central mine: Eng. and Min. Jour., v. 119, p. 609–611.

Brief account of the history and geology of a copper mine southwest of Jerome.

1926. Fearing, J. L., Jr. Some notes on the geology of the Jerome district, Ariz: Econ. Geology, v. 21, p. 757–773.

Discusses geology and ore deposits in Jerome. Believes that United Verde Extension ore body was not originally connected to United Verde ore body.

1926. Lindgren, Waldemar. Ore deposits of the Jerome and Bradshaw Mountains quadrangles, Ariz.: U. S. Geol. Survey Bull. 782, 192 p.

Main value of this bulletin is the wealth of information on the mines and prospects in the area. Reconnaissance map of Jerome quadrangle included, but regional geology not treated in a comprehensive fashion.

1927. Anderson, C. A. Voltaite from Jerome, Ariz.: Am. Mineralogist, v. 12, p. 287–290.

Brief description of one of the hydrous sulfates formed by the mine fires.

1928. Lausen, Carl. Hydrous sulphates formed under fumarolic conditions at the United Verde mine: Am. Mineralogist, v. 13, p. 203–225.

Interesting description of 9 hydrous sulfate minerals, 5 of them new, formed at or near surface resulting from mine fires.

1930. Hansen, M. G. Geology and ore deposits of the United Verde mine: Min. Cong. Jour., v. 16, p. 306–312.

First paper to suggest Precambrian movement on Verde fault.

1930. Lausen, Carl. The pre-Cambrian greenstone complex of the Jerome quadrangle: Jour. Geology, v. 38, p. 174–183.

Separated a greenstone complex from Yavapai schist (metasedimentary rocks) and suggested an unconformable relationship between the two; Wilson (1939) later showed that a fault relationship exists between these units.

1930. Newhouse, W. H., and Flaherty, G. E. The texture and origin of some banded or schistose sulphide ores: Econ. Geology, v. 25, p. 600–620.

Brief description of ore from the United Verde mine at Jerome.

1930. Ralston, O. C. Possibilities of zinc production in Arizona: Min. Jour., v. 14, p. 11.

Description of zinc occurrence in the United Verde mine, Jerome.

1931. Ingalls, W. R. World survey of the zinc industry: Min. Met. Soc. America, N. Y.

Emphasized the zinc content of pyritic ore at United Verde mine, Jerome.

1932. Ransome, F. L. General geology and summary of ore deposite, *in* Ore deposits of the southwest: 16th Internat. Geol. Cong. Guidebook 14, Excursion C–1, p. 1–23.

Resumé of Arizona geology, but gives considerable attention to ore deposits of Jerome. Emphasized Precambrian movement on Verde fault.

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

Historical data about mines at Jerome.

1936. Stoyanow, A. A. Correlation of Arizona Paleozoic formations: Geol. Soc. America Bull., v. 47, p. 459–540.

Described the Devonian Jerome formation, which is termed Martin limestone in this report.

1938. Reber, L. E., Jr. Jerome district, *in* Some Arizona ore deposits: Ariz. Bur. Mines Bull. 145, p. 41–65.

Valuable sequel to Reber's 1922 paper; later developments in Jerome district adequately described and illustrated. Much valuable information on the United Verde Extension mine which closed in 1938.

1938. Schwartz, G. M. Oxidized copper ores of the United Verde Extension mine: Econ. Geology, v. 33, p. 21–33.

Petrographic study of chalcocite and oxide ore; paragenesis of minerals.

1939. Wilson, E. D. Pre-Cambrian Mazatzal revolution in central Arizona: Geol. Soc. America Bull., v. 50, p. 1113–1164.

First attempt to divide rocks of the Yavapai schist of Jerome area into separate stratigraphic units and a forward step in the understanding of the Precambrian geology.

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

Brief description of the Iron King mine.

1943. Gutschick, R. C. The Redwall limestone (Mississippian) of Yavapai County, Ariz.: Plateau, v. 16, no. 1, p. 1–11.

Presents measured sections of Redwall limestone as exposed in Jerome area.

1945. Huddle, J. W., and Dobrovolny, E. Late Paleozoic stratigraphy and oil and gas possibilities of central and northeastern Arizona: U. S. Geol. Survey Oil and Gas Inv., Prelim. Chart 10.

Description and measured sections of Paleozoic rocks adjacent to the Jerome area.

1946. Yates, P. F. Bottoming of the United Verde sulphide pipe: Mimeographed paper distributed at meeting of Ariz. Section, Am. Inst. Min. Metall. Eng., Tucson, Ariz., Oct. 1946, 20 p.

Contains level plans and section showing character of United Verde ore body and description of exploration activities on lower levels of the mine.

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

Excellent brief account of the geology of this mine emphasizing character of ore shoots.

1949. Mahard, R. H. Late Cenozoic chronology of the upper Verde Valley, Ariz.: Denison Univ. Bull. Jour. Sci. Lab., v. 41, p. 97–127.

Suggests a complex history for the Verde Valley before deposition of the Verde formation.

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

Preliminary statement on the occurrence of ore in the Iron King mine.

1950. Little, W. W. Radial blast holes for drilling an irregular ore body: Min. Engineering, p. 463–465.

Describes copper-zinc ore to north of main ore body, United Verde mine.

1950. Price, W. E., Jr. Cenozoic gravels on the rim of Sycamore Canyon, Ariz.: Geol. Soc. America Bull., v. 61, no. 5, p. 501–508.

Presents evidence to show that before basaltic eruptions on plateau, drainage from Precambrian rocks in Jerome area was to the north.

1951. Anderson, C. A. Older Precambrian structure in Arizona: Geol. Soc. America Bull., v. 62, p. 1331–1346.

Preliminary statement on the Precambrian stratigraphy and structure in the Jerome area.

1951. McNair, A. H. Paleozoic stratigraphy of part of northwestern Arizona: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 503–541.

Presents evidence for believing Cambrian sedimentation did not extend to Jerome area.

1952. Creasey, S. C. Geology of the Iron King mine, Yavapai County, Ariz: Econ. Geology, v. 47, p. 24–56.

Describes the local geology of this mine, the alteration, metallization, and structural control of the ore.

1952. Huddle, J. W., and Dobrovolny, E. Devonian and Mississippian rocks of central Arizona: U. S. Geol. Survey Prof. Paper 233–D, p. 67–112.

Description of these rocks southeast of the Jerome area, with discussion of the regional distribution and history.

1955. Huff, L. C. A Paleozoic geochemical anomaly near Jerome, Arizona: U. S. Geol. Survey Bull. 1000–C, p. 105–118.

Sampling of the Tapeats(?) sandstone revealed anomalous copper concentrations of more than 100 ppm (parts per million) near Jerome and in an area extending about 2 miles southeast from Jerome. Elsewhere this same basal sandstone contains about 20 ppm copper. Anomalous zinc concentrations of more than 100 ppm are distributed in roughly the same area as the copper. Huff concluded that the abnormal copper and zinc content of the Tapeats(?) represents products of erosion from the ore deposits and that geochemical prospecting techniques can be used to identify and trace heavy-metal anomalies in basal sandstones such as the Tapeats(?).

GENERAL GEOLOGY

OLDER PRECAMBRIAN ROCKS

YAVAPAI SERIES

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

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

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

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

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

ASH CREEK GROUP

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

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

The section of the Ash Creek group is perhaps 20,000 feet thick, not including the base or top. The sum of the maximum thickness of each formation is 23,500 feet, but because some of the volcanic formations are lenticular, a lower figure is a more accurate estimate for the group as a whole.

GADDES BASALT

Distribution

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

Thickness and stratigraphic relationship

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

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

Lithology and internal structure

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

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

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

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

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

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

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

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

	1	2	3
SiO_2	54. 8	59. 59	49. 06
Al_2O_3	16. 2	17. 31	15. 70
Fe_2O_3	2. 4	3. 33	5. 38
FeO	9. 8	3. 13	6. 37
MnO	. 26	. 18	. 31
MgO	3. 7	2. 75	6. 17
CaO	4. 4	5. 80	8. 95
K_2O	. 36	2. 04	1. 52
Na_2O	3. 2	3. 58	3. 11
TiO_2	1. 0	. 77	1. 36
P_2O_5	. 28	. 26	. 45
Ignition loss	2. 0		

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

BUZZARD RHYOLITE

Distribution

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

Thickness and stratigraphic relationship

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

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

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

Lithology and internal structure

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

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

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

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

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

TABLE 3.—*Chemical analysis of Buzzard rhyolite*

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

	Percent
SiO_2	75.9
Al_2O_3	11.9
Fe_2O_3	1.3
FeO	2.7
MnO	.10
MgO	.43
CaO	.96
K_2O	1.2
Na_2O	4.4
TiO_2	.32
P_2O_5	.02
Ignition loss	.86

SHEA BASALT

Distribution

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

Thickness and stratigraphic relationship

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

East of Mingus Mountain, the base of the Shea basalt is exposed north of Oak Wash (1,344,000 N.; 447,000 E.). Interbedded tuffaceous sediments near the top of the Buzzard rhyolite dip northwest, conformable with the rhyolite-basalt contact. Graded bedding in the tuffaceous sedimentary rocks and vesicular flow tops indicate that the tops of the tuffaceous and flow rocks face northwest, and the Buzzard rhyolite is older than the Shea basalt. On the south side of Mingus Mountain, the site of probable contact between the Shea basalt and Deception rhyolite is covered by Paleozoic rocks (1,322,000 N.; 448,000 E.), but the eastward extension of the Deception rhyolite from where it overlies the Buzzard rhyolite indicates that the Deception rhyolite is younger than the Shea basalt which lenses out to the west. In the Mescal Gulch area, the structure involving the Shea basalt-Deception rhyolite contact is complex, and poor exposures in part mask the contact. The major structure in the Mescal Gulch area appears to be a northwestward-plunging anticline, so that younger formations appear progressively to the northwest. By this structural interpretation, the Shea basalt is the oldest and is overlain by the Deception rhyolite and the Grapevine Gulch formation in that order. Supporting evidence occurs in Hull Canyon (1,360,000 N.; 436,000 E.) where the Grapevine Gulch formation clearly overlies the Deception rhyolite. Rhyolite crops out in small masses in the Shea basalt, and some basalt or andesite appears in the Deception rhyolite indicating that rhyolitic and basaltic rocks were erupted simultaneously.

Lithology and internal structure

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

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

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

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

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

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

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

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

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

[1] All iron determined as FeO.

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

BRINDLE PUP ANDESITE

Distribution

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

Thickness and stratigraphic relationship

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

Lithology and internal structure

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

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

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

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

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

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

	Percent
SiO_2	62. 6
Al_2O_3	15. 4
Fe_2O_3	1. 7
FeO	6. 6
MnO	. 24
MgO	2. 1
CaO	3. 2
K_2O	2. 7
Na_2O	3. 1
TiO_2	. 64
P_2O_5	. 22
Ignition loss	1. 8

DACITE OF BURNT CANYON

Distribution, thickness, and stratigraphic relationship

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

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

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

Lithology and internal structure

The dacite of Burnt Canyon contains conspicuous feldspar phenocrysts, ranging from 1 to 2 millimeters in length, in a finely crystalline groundmass in which small quartz crystals are visible. Except for the quartz, the rock is generally similar to the porphyritic Brindle Pup andesite. Local vesicular facies are common, particularly in the southeastern exposures.

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

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

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

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

	Percent
SiO_2	69.6
Al_2O_3	14.0
Fe_2O_3	1.8
FeO	2.7
MnO	.15
MgO	.84
CaO	2.1
K_2O	1.5
Na_2O	4.0
TiO_2	.40
P_2O_5	.15
Ignition loss	2.7

DECEPTION RHYOLITE

Distribution

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

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

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

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

Thickness and stratigraphic relationship

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

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

In the Ash Creek drainage basin, the upper contact with the Grapevine Gulch formation presents some problems. Evidence is generally convincing that the Grapevine Gulch formation lies conformably above the Deception rhyolite, particularly along the contact east of the northward-trending granodiorite porphyry dike (441,400 E.). West of this dike, embayment of the Grapevine Gulch formation within the rhyolite may be interpreted as intertonguing, or may be the trough of a small syncline. However, proof of a fold by reversal of dips or by the outcrop pattern of beds, was not obtained, suggesting that the rhyolitic rocks do intertongue with the younger Grapevine Gulch sedimentary rocks. West of the embayment, there is a slight structural discordance between beds in the Grapevine Gulch formation and the contact with the Deception rhyolite. In part this discordance may represent a depositional contact of different facies on the older rocks, but less than a mile east of Kendall Camp, the discordance is greater, and the occurrence of silicified zones may imply a fault relationship. North of Kendall Camp (1,333,000 N.; 429,500 E.), the Grapevine Gulch formation rests directly on the dacite of Burnt Canyon, whereas east of Kendall Camp (pl. 1, 1,327,500 N.; 436,500 E.), the dacite of Burnt Canyon is 2,000–3,000 feet stratigraphically below the contact between the Grapevine Gulch formation and Deception rhyolite. Because bodies of rhyolitic rock are commonly lenticular, the disappearance of the rhyolite northward beneath the Grapevine Gulch formation cannot be used as positive evidence of a fault relationship or unconformity.

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

Lithology and internal structure

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

GRAPEVINE GULCH FORMATION

Distribution

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

Thickness and stratigraphic relationship

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

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

Lithology and internal structure

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

Lithic tuffaceous sedimentary rock.—The coarse-grained lithic tuffaceous sedimentary rocks generally are poorly bedded; the layers range from 4 to 50 feet in thickness. The units shown on the geologic map (pl. 1) contain fine-grained tuffaceous interbeds ranging from 2 to 50 feet in thickness; some of the units contain about 50 percent of the fine-textured interbeds. The lithic tuffs are pale-greenish gray or greenish black where fresh and are buff colored where weathered. The fragments generally range from 0.1 to 1.0 inch in diameter, although locally, rare fragments are as long as 12 inches.

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

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

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

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

Fine-grained tuffaceous sedimentary rock.—The fine-grained tuffaceous and cherty sedimentary rocks are well-bedded, the layers range in thickness from 1/8 to 2 inches. Graded bedding is well defined in the coarser grained facies, and locally, crossbedding and channeling are present. In Hull Canyon and in the pit of the United Verde mine at Jerome, fine-grained tuffaceous beds are foliated (pl. 4*A*), but the lithic tuff beds are not. Minor folds from 2 to 20 feet across occur in the fine-grained tuffaceous rocks in Yaeger and Hull Canyons and in the pit (pl. 4*B*) and underground workings of the United Verde mine.

The crystal tuff beds consist of angular feldspar and subordinate quartz grains, from 1/2 to 1 millimeter in diameter, embedded in a fine-grained chloritic matrix. The feldspar weathers into conspicuous pale-yellow grains; variations in the size of the crystals accentuate the bedding. Thin-section study reveals that the feldspar grains are albitic and commonly sericitized; the matrix consists of microgranular quartz, albite, hornblende, epidote, chlorite, apatite, and ilmenite or leucoxene.

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

	1	2
SiO_2	65.5	65.68
Al_2O_3	15.2	16.25
Fe_2O_3	3.0	2.38
FeO	4.4	1.90
MnO	.13	-------
MgO	1.3	1.41
CaO	2.0	3.46
K_2O	1.3	2.67
Na_2O	5.6	3.97
TiO_2	.57	.57
P_2O_5	.20	.15
Ignition loss	.95	-------

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

ALDER GROUP

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

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

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

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

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

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

INDIAN HILLS VOLCANICS

Distribution

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

Thickness and stratigraphic relationship

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

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

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

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

Lithology and internal structure

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

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

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

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

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

SPUD MOUNTAIN VOLCANICS

Distribution and thickness

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

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

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

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

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

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

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

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

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

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

Stratigraphic relationship

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

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

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

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

Lithology and internal structure

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

IRON KING VOLCANICS

Distribution, thickness, and stratigraphic relationship

Iron King volcanics, named here from the exposures in Iron King Gulch, occurs in a northward-trending belt south of Humboldt in the southwest corner of the Jerome area (pl. 1). The formation crops out from the southern boundary of the map area northward for about 8,000 feet where it is covered by Cenozoic rocks in Lonesome Valley.

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

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

Lithology and internal structure

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

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

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

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

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

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

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

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

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

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

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

TEXAS GULCH FORMATION

Distribution

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

Thickness and stratigraphic relationship

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

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

Lithology and internal structure

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

CHAPARRAL VOLCANICS

Distribution

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

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

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

Thickness and stratigraphic relationship

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

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

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

Lithology and internal structure

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

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

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

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

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

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

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

GREEN GULCH VOLCANICS

Distribution

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

Thickness and stratigraphic relationship

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

The stratigraphic relationship of the formation could not be determined as the formation is bounded on the southeast by the Chaparral fault and on the west by intrusive rocks, chiefly gabbro and alaskite. Locally, breccia, crystal tuff, and rhyolitic tuffaceous beds in the Green Gulch volcanics somewhat resemble the Spud Mountain volcanics. The basaltic flows, however, more closely resemble those in the Iron King volcanics, which overlie the Spud Mountain volcanics. Despite the local similarities, the lithology of the Green Gulch volcanics is not sufficiently like that of any other formation to permit a correlation. Until such time, therefore, that the Green Gulch volcanics is found in depositional contact with other units, its stratigraphic position will remain unknown and the unit should be given formational rank.

Lithology and internal structure

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

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

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

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

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

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

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

ORIGIN OF YAVAPAI SERIES

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

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

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

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

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

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

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

INTRUSIVE ROCKS

INTRUSIVE RHYOLITE

Distribution and relation to other rocks

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

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

General characters

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

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

QUARTZ PORPHYRY

Distribution

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

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

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

Relation to other rocks

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

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

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

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

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

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

The large masses of quartz porphyry are essentially concordant with the structure of the host rock, and where the host rocks are generally homoclinal, elongate sill-like bodies of quartz porphyry occur, such as at Jerome where the quartz porphyry body is 2,000 to 3,000 feet wide and more than 2 miles long, conformable with the adjacent Grapevine Gulch formation and Deception rhyolite. North of Oak Wash, two bodies of quartz porphyry trend northeastward, essentially conformable to the Shea basalt and Buzzard rhyolite. Where the host rocks are folded, the quartz porphyry masses generally conform to the folds, such as south of Mescal Gulch where the west-northwestward-plunging fold in the Shea basalt is reflected in the western contact of the quartz porphyry. In detail, however, the quartz porphyry masses crosscut all the flows and sedimentary beds along their contacts.

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

General characters

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

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

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

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

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

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

[1] Total iron reported as FeO.

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

The only available chemical analysis of quartz porphyry not collected from areas of hydrothermal alteration is of analysis 1, table 11; it compares closely to the average composition of quartz porphyry (analysis 7), except for a higher Na_2O and lower K_2O content. The remaining analyses are of quartz porphyry that has undoubtedly been hydrothermally altered. The most southerly of these, analysis 2, is richer in Na_2O and poorer in K_2O than the quartz porphyry in and near the United Verde mine where sericite is an important constituent. The CaO content is very low in analysis 2, indicating that albitization must have taken place; this is confirmed in part by microscopic studies of quartz porphyry in the region around the Copper Chief mine. The higher MgO content in analyses 5 and 6 is related to the introduction of chlorite in the rock, a common feature near the United Verde mine. The high CO_2 content of analyses 4 and 6 indicates the presence of calcite or dolomite. The classification of the quartz porphyry is difficult because of the alteration of much of the porphyry in and near Jerome, and Reber's statement that the "usual composition is that of a normal rhyolite, often as albite rhyolite and rarely a dacite" (Reber, 1922, p. 13) is an excellent summary.

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

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

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

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

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

GABBRO

Distribution

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

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

Relation to other rocks

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

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

West of Spud Mountain in the Prescott quadrangle, some gabbro is essentially parallel to the northward-trending Green Gulch volcanics but the dikelike masses of gabbro have a northeast trend, parallel to the Chaparral and Spud faults. Renewed movement on these faults produced a northeastward-trending foliation in some of the gabbro.

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

General characters

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

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

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

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

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

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

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

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

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

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

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

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

[1] Fe_2O_3 calculated as FeO.

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

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

QUARTZ DIORITE

Name and distribution

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

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

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

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

Two outlying masses of quartz diorite are exposed in the Mingus Mountain quadrangle, north and northwest of the eastern part of the quartz diorite mass. The larger mass is in Ward Pocket, on the south wall of Black Canyon (1,325,000 N.; 450,000 E.) and the smaller mass, only 600 feet in its longest dimension, is more than 2½ miles from the main body of quartz diorite. This small mass is exposed in one of the tributaries to Ash Creek (1,317,800 N.; 435,500 E.).

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

Relation to other rocks

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

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

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

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

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

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

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

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

General characters

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

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

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

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

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

TABLE 13.—*Modal composition of quartz diorite*

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

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

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

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

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

A facies of quartz diorite is exposed near 1,286,000 N.–418,500 E. (pl. 1), in which conspicuous large euhedral plagioclase crystals indicate a porphyritic texture, but close examination reveals a seriate texture. This facies contains less quartz and biotite and more hornblende than the typical western facies of quartz diorite, which is intermixed with this seriatic facies. A primary flow structure is parallel to the contacts of the included masses of Spud Mountain volcanics.

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

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

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

GRANODIORITE PORPHYRY DIKES

Distribution

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

Relation to other rocks

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

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

General characters

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

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

Many dikes are almost leucocratic, and consist of quartz and dulled plagioclase phenocrysts in a pale-gray aphanitic groundmass. Other dikes contain abundant hornblende phenocrysts in addition to the quartz and feldspar, and mafic minerals are recognizable in the groundmass. Some of the wider dikes, and particularly the bulbous dike at the Brindle Pup mine and the stubby dike in Black Canyon, have a seriate texture. The quartz and plagioclase crystals range from 0.5 to 2 millimeters.

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

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

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

TABLE 14.—*Chemical composition of granodiorite porphyry*

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

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

[1] Total iron reported as FeO.

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

ALASKITE AND APLITE

Distribution

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

Relation to other rocks

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

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

General character

The unweathered alaskite and aplite are pale red or grayish-orange pink; locally they are nearly white owing to alteration. Dynamic metamorphism induced a pronounced foliation in much of the alaskite and aplite in the southwest corner of the Jerome area. Relicts of the original rock particularly in the northern part of the alaskite mass reveal the original texture and mineralogic composition. Shearing along the margins of the eastern mass of alaskite in the Chaparral volcanics was especially intense; so intense that the narrow southern whiplike prong of this mass may be entirely due to deformation. The contacts of this mass are zones of mechanical mixing of alaskite and rhyolitic tuffaceous rock. Where intensely sheared, these rocks are schistose or gneissose. The quartz grains have been drawn out into "ribbons," and the feldspar partly recrystallized into elongated aggregates.

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

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

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

QUARTZ AND JASPER VEINS

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

Distribution

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

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

Occurrence

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

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

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

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

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

REGIONAL CORRELATION OF THE OLDER PRECAMBRIAN ROCKS

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

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

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

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

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

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

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

At Bagdad, about 40 miles west of Prescott, metamorphic rocks similar to the Yavapai series in the Jerome area have been tentatively correlated with the Yavapai (Anderson, 1951, p. 1336). Three lithologic units were recognized: (1) andesitic-basaltic flows and minor sedimentary interbeds; (2) rhyolitic and rhyolitic-andesitic tuffaceous sedimentary rocks and shale interbeds; (3) shale, sandy shale, and sandstone, recrystallized to quartzite and mica schist. These rocks were folded and faulted before the intrusion of a diverse series of igneous rocks culminating in granite.

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

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

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

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

UNCONFORMITY AT THE BASE OF THE PALEOZOIC ROCKS

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

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

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

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

PALEOZOIC SEDIMENTARY ROCKS

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

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

TAPEATS SANDSTONE(?)

Distribution

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

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

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

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

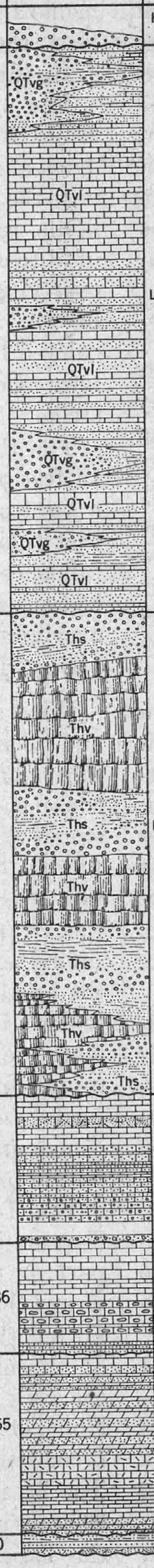

FIGURE 2.—Stratigraphic column of Paleozoic, Tertiary, and Quaternary rocks in the Jerome area.

Thickness and stratigraphic relationship

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

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

Lithology

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

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

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

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

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

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

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

Age and correlation

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

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

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

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

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

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

MARTIN LIMESTONE

Distribution

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

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

Thickness

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

Lithology

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

The Martin limestone comprises four units. These units were distinguished only in measured sections, because detailed mapping of the Martin was not necessary to meet our objectives. For convenience of description, however, each of the units will be described individually. Listed in order of decreasing age, they are: lower unit, lithographic unit, middle unit, and upper unit.

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

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

The lower unit ranges from 34 to 53 feet in thickness and is nearly uniform lithologically. A few thin beds, however, contrast in color, texture, and bedding. The lower unit is dominantly light brownish gray, and subordinately yellowish gray. Color depends partly on grain size and partly on the amount of impurities; the finer grained rocks generally are of lighter hues, and the impurities, presumably clay, locally color irregular zones and beds pinkish and yellowish. Medium- to coarse-grained beds are most abundant, but some are fine, and others, coarse. Beds range in thickness from 1/8 inch to 5 feet. The clastic origin of some beds is shown by etched out crossbedding and channeling. Other beds are massive; whereas a few are distinctly fragmental. The massive beds are always coarse grained, perhaps owing to crystallization.

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

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

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

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

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

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

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

The lower 60 feet of the unit comprises interbedded (1) light-gray, very fine grained, dense dolomitic limestone, (2) light-olive gray, medium- to coarse-grained dolomitic limestone or dolomite, and (3) very light gray, thin-bedded, finely laminated dolomitic limestone. The overlying beds are generally similar to those just described, except for clay and silt impurities and interbeds of claystone or siltstone. Clay mottles the dolomitic limestone purplish to reddish and commonly purple siltstone or claystone interbeds separate these impure beds. The zone from 100 to 140 feet above the base of the upper unit comprises alternating thin-bedded impure slabby dolomitic limestone and poorly exposed purplish limy siltstone that crop out between cliffs of massive dolomitic limestone. These impurities reappear sporadically in lesser amounts in the overlying beds. The uppermost 50 feet of the unit consists of grayish-orange-pink, medium-grained dolomitic limestone beds; it is uniformly and evenly bedded, and crops out in uniform steplike ledges.

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

Age and correlation

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

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

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

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

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

REDWALL LIMESTONE

Distribution, thickness, and stratigraphic relationship

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

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

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

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

Lithology

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

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

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

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

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

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

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

Age and correlation

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

SUPAI FORMATION

Distribution

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

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

Thickness and stratigraphic relationship

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

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

Lithology

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

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

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

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

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

Age and correlation

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

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

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

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

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

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

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

UNCONFORMITY AT THE BASE OF THE TERTIARY ROCKS

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

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

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

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

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

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

CENOZOIC ROCKS

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

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

HICKEY FORMATION

Distribution

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

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

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

Thickness and stratigraphic relationship

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

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

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

Lithology

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

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

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

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

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

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

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

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

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

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

Age and correlation

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

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

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

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

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

In this report the Hickey is tentatively assigned to the Pliocene epoch. The reasons are as follows: (1) much of the satisfactory dating of comparable Cenozoic rocks in Arizona has been to divisions of the Pliocene epoch; (2) the vertebrate fossils found in the sedimentary rock near Prescott do not preclude the possibility of Pliocene age; (3) the tectonic activity subsequent to the deposition of the Hickey formation followed by the deposition and much erosion of the Verde formation do not favor a Pleistocene age for the Hickey formation; (4) bias on our part because of the accurate early Pliocene dating of comparable rocks 20 miles south of Prescott.

VERDE FORMATION

Distribution

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

Thickness and stratigraphic relationship

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

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

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

Lithology

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