Geology and Base-Metal Deposits of West Shasta Copper-Zinc District Shasta County, California

GEOLOGICAL SURVEY PROFESSIONAL PAPER 285

Prepared in cooperation with the State of California, Department of Natural Resources, Division of Mines

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By A. R. KINKEL, Jr., W. E. HALL, *and* J. P. ALBERS

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UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1956

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 BASE-METAL DEPOSITS OF THE WEST SHASTA COPPER-ZINC DISTRICT, SHASTA COUNTY, CALIFORNIA

By A. R. Kinkel, Jr., W. E. Hall, and J. P. Albers

ABSTRACT

The Shasta copper-zinc district of northern California is in the foothills of the Klamath Mountains at the north end of the Sacramento Valley. The district contains two main areas of base-metal ore deposits that are called the West Shasta copper-zinc district and the East Shasta copper-zinc district. The West Shasta district, the subject of this report, is a well-defined, northeastward-trending district about 8 miles long and 2 miles wide, west of the Sacramento River. The south end of the district is 9 miles northwest of Redding, the county seat of Shasta County.

The West Shasta copper-zinc district includes nine base-metal mines that have been productive, and many prospects. The ore consists of large bodies of massive pyrite that contain copper and zinc sulfides and minor amounts of gold and silver. The ore bodies have been mined primarily for copper and zinc, although massive pyrite has been mined at the Iron Mountain mine for sulfur. Only the Iron Mountain mine has been operated continuously in recent years.

The Paleozoic rocks of the West Shasta district range in age from Middle Devonian to Mississippian. Mesozoic rocks, if they were present, have been eroded except for remnants of the Chico formation of Late Cretaceous age. The Chico is overlain by the Red Bluff formation of Pleistocene age and by Recent stream gravels. The Paleozoic rocks are intruded by a pluton of albite granite, the Mule Mountain stock, and by a pluton of biotite-quartz diorite, the Shasta Bally batholith. These intrusives are probably of Late Jurassic or Early Cretaceous age.

The oldest formation that is exposed in the West Shasta district is the Copley greenstone of probable Middle Devonian age. It is composed of volcanic flows, volcanic breccia, and tuffs of intermediate and basic composition, and of a few beds of shale and rhyolitic tuff. The lower part of the formation contains massive flows; the upper part contains much pillow lava and pyroclastic material. The formation is at least 3,700 feet thick, but the base is not exposed in the mapped area.

The Balaklala rhyolite of Middle Devonian age conformably overlies the Copley greenstone, although in the eastern part of the mapped area the Balaklala rhyolite wedges out and locally interfingers with the Copley greenstone. In this area the upper part of the Copley greenstone appears to be equivalent in age to the Balaklala rhyolite. A transition zone, which contains material that is characteristic of both units, locally lies between the two formations. The Balaklala rhyolite is composed of soda-rich rhyolitic flows and pyroclastic material. The stratigraphy in the Balaklala rhyolite has been mapped on the basis of bedded pyroclastic rocks, and on the size of quartz phenocrysts and other lithologic differences in the rhyolite flows. Nonporphyritic rhyolite and porphyritic rhyolite containing phenocrysts of quartz and plagioclase are lithologic subdivisions; the porphyritic rhyolite is further subdivided into medium-phenocryst rhyolite containing quartz phenocrysts 1 to 4 millimeters in diameter and coarse-phenocryst rhyolite with quartz phenocrysts more than 4 millimeters in diameter. The Balaklala is here subdivided into lower, middle, and upper stratigraphic units. Nonporphyritic rhyolite predominates in the lower unit of the Balaklala; medium-phenocryst rhyolite predominates in the middle unit; and coarse-phenocryst rhyolite is characteristic of the upper unit. The Balaklala rhyolite is probably 3,500 feet thick in the central part of the district, but forms a volcanic pile that thins on the edges.

The Kennett formation of Middle Devonian age overlies the Balaklala rhyolite. The Kennett is composed of black siliceous shale, gray shale, rhyolitic tuff, and limestone. The rhyolitic tuff of the Kennett formation is interbedded with shale and grades downward through a transition zone to rhyolitic tuff beds that are part of the upper unit of the Balaklala rhyolite. The maximum thickness of the Kennett formation is probably not more than 400 feet, but folding and repetition by faulting make it impossible to obtain an exact thickness. The Kennett formation is missing in the westernmost part of the area, where the Bragdon formation lies conformably on the Balaklala rhyolite.

The Bragdon formation of Mississippian age conformably overlies the Kennett formation in much of the mapped area, although outside this area part of the Kennett was apparently uplifted by warping and an erosional unconformity separates the two formations. The Bragdon formation is composed largely of gray- and tan-weathering shale, but it also contains beds of conglomerate, grit, and sandstone. The Bragdon formation is 3,500 feet thick in the mapped area, but only the lower part of the formation is present. Diller estimated that the Bragdon is 6,000 feet thick.

The Mule Mountain stock of albite granite intrudes the Copley greenstone and the Balaklala rhyolite. It is a soda-rich siliceous intrusive rock that is composed principally of albite and quartz with minor amounts of epidote. Locally, it contains hornblende. It is probably syntectonic (Nevadan) in age.

The Shasta Bally batholith is composed predominantly of biotite-quartz diorite. It intrudes the Copley greenstone, the Bragdon formation, and the albite granite in the mapped area; west of the mapped area it is nonconformably overlain by rocks of Early Cretaceous age. It is a post-tectonic intrusive of Late Jurassic or Early Cretaceous age.

The Chico formation of Late Cretaceous age is present as erosion remnants in a few places in the mapped area. It is composed of shale, sandstone, and conglomerate. A major unconformity separates the Chico from the pre-Cretaceous rocks. The Chico formation is gently tilted, but is not folded or metamorphosed as are the underlying rocks.

The Red Bluff formation of Pleistocene age overlies the Chico unconformably. It is composed of poorly cemented sand, gravel, and conglomerate that form a veneer on cut surfaces and form part of the fill of the old Sacramento Valley. The Red Bluff formation is deeply entrenched by the present streams.

The Paleozoic rocks in the West Shasta district are folded into a broad anticlinorium that contains many small folds: the axis of the anticlinorium trends N. 15° E. in the central part of the mineral belt through Iron Mountain and Behemotosh Mountain. It has a culmination over the central part of the mineral belt and plunges north at a low angle at the north edge of the mapped area and south at a low angle at the south end of the mineral district. It is interrupted by the Mule Mountain stock south of Iron Mountain. The anticlinorium is flanked by broad synclines on the east and west. Although folding is on such a broad scale that the average dip of the flanks of the anticlinorium is not more than 20°, dips of individual beds range from horizontal to 90°. The rocks are strongly folded in some areas, but adjacent areas are only moderately folded. Folding is particularly erratic in distribution where there is a great difference in the competence of adjacent rocks, as between beds of tuff and thick massive flows, and between shale and conglomerate.

Foliation is common in many parts of the district, although the rocks in many places contain no planar structures. The rocks range from those that are strongly deformed with schistose or gneissic structure through moderately folded rocks having fracture cleavage to rocks that are almost undeformed. In parts of the district foliation is parallel to bedding; in other parts it cuts across the bedding. Where foliation is most intense it is rarely possible to determine the relationship between foliation and bedding. The Mule Mountain stock was intruded into rocks that were already foliated, as it cut across and crumpled the foliation around its borders. The Shasta Bally batholith formed a zone of amphibolite, gneiss, and migmatite from the Copley greenstone along its border.

Faults are abundant in the district. The main faults have two dominant directions, N. 20°–45° W. and N. 60°–80° E.; in both groups the north side is generally downthrown relative to the south.

The Shasta copper-zinc district has produced 54 percent of the copper in California through 1946; the major part of the production has come from the West Shasta copper-zinc district. The zinc production has been small because the ore was direct-smelted and zinc was not recovered; bodies of high-grade zinc ore were mined and treated separately. Gold and silver have been recovered from gossan that overlies massive sulfide ore at Iron Mountain and from sulfide ore that was smelted for copper. Through 1951, 3,600,000 tons of pyrite from the Iron Mountain mine had been treated for its sulfur content, but sulfur was not recovered at any of the other mines. Several thousand pounds of cadmium were recovered from zinc-rich ore at the Mammoth mine.

The copper-zinc ore deposits of the West Shasta district are bodies of massive pyrite that contain chalcopyrite and sphalerite and minor quantities of gold and silver. The most striking features of the ore are its uniformity, its lack of megascopic gangue minerals, and its sharp boundary with barren or weakly pyritized wall rocks. The ore has a brassy, metallic appearance, and some large ore bodies contain as little as 3.5 percent acid insoluble material. In most ore bodies the massive sulfide is separated from barren wall rock by a thin selvage of gouge; gradational contacts between ore and wall rock are very rare. Most of the ore bodies of the district have a lenticular form and are flat lying; their greatest dimension is in a horizontal plane. Several ore bodies are saucer shaped, one is domal, and one is synclinal. Steeply dipping ore bodies are not characteristic of the district, but they do occur in the Hornet mine at Iron Mountain, in the Golinsky mine, and in the Sutro mine.

The bodies of massive sulfide ore, before postmineral faulting, ranged in size from that of the largest body, at the Iron Mountain mine, which has a length of 4,500 feet, a width of several hundred feet, and thickness of slightly more than 100 feet, to small bodies only a few tens of feet in maximum dimensions. Some individual ore bodies are discrete lenses in a broad mineralized zone, but at a few mines an originally continuous ore body has been offset by faults into separate blocks of ore.

All the massive sulfide bodies contain some copper and zinc minerals, but some massive pyrite bodies, as the Hornet ore body at Iron Mountain, contain too little copper or zinc to repay mining as a copper-zinc ore. The only value of the low-grade pyritic bodies at present is in their sulfur content, as the iron-rich residue from roasting has not been utilized. Massive pyrite that is low in copper and zinc occurs both as separate bodies and as low-grade parts of ore bodies that contain minable amounts of copper and zinc. In some mines the upper or lower part of a flat-lying massive sulfide ore body contains so little copper and zinc that it has been left in place. Ore bodies that contain minable amounts of copper and zinc are identical with low-grade pyritic bodies in other respects. Massive sulfide ore that contains a high percentage of zinc occurs mainly at the Mammoth mine, although zinc was recovered from parts of the Richmond ore body at Iron Mountain.

Some disseminated copper ore occurs in the No. 8 mine ore body of the Iron Mountain mine and in the Balaklala mine beneath the massive sulfide ore bodies. This copper ore consists of chalcopyrite and pyrite in about equal amounts as veinlets and disseminations in siliceous schistose rock; there is no gradation between the massive sulfide ore and the siliceous disseminated copper ore.

Three main base-metal ore controls can be recognized in the copper-zinc district. These are: stratigraphic control within the Balaklala rhyolite; structural control by folds and foliation; and feeder fissures along which the solutions ascended.

A conjunction of the three types of ore controls was probably a prerequisite for the formation of a major ore body.

The minable massive sulfide ore bodies that have thus far been found in the West Shasta district are at the same stratigraphic horizon throughout the district, in the upper part of the middle unit of the Balaklala rhyolite. Ore occurs through a stratigraphic thickness of 600 feet in the Balaklala rhyolite at Iron Mountain, and it is possible that locally the favorable zone may have a greater thickness. The top of this zone is the base of the upper unit of the Balaklala rhyolite, which is composed of coarse-phenocryst rhyolite or tuff, but the lower limit is not marked by distinctive flows. The upper part of the middle unit is a group of discontinuous flows and lenticular beds of coarse and fine pyroclastic rocks. The heterogeneous nature of this material is such that the detailed stratigraphy at each mine is unique, yet the fact that this heterogeneous group is capped by a recognizable unit over most of the district makes it possible to locate the ore zone at one general horizon with a fair degree of certainty. Pyritization in this zone is much more widespread and continuous than the scattered distribution of known ore bodies would indicate. Many exploratory drill holes have disclosed heavily pyritized rock in the favorable zone at consider-

able distances from known ore, and hidden ore bodies have been located by systematically drilling areas in the favorable zone.

At some places the character of the rock that was replaced to form ore bodies can be determined; the favored host rock for ore bodies appears to be porphyritic rhyolite that has 2- to 3-millimeter quartz phenocrysts, particularly where this rock is overlain by thinly bedded tuff or fine pyroclastic rocks.

Individual ore bodies or groups of ore bodies tend to be concentrated on or near the axes of broad folds. The ore bodies were formed both in anticlines and in synclines, although those in synclines predominate if basin-shaped structures are included with the synclines. Examples of ore bodies formed in synclines or basin-shaped warps are: the Richmond ore body of the Iron Mountain mine, ore bodies of the Balaklala and Shasta King mines, and probably the ore body of the Early Bird mine. The broad ore zone at the Mammoth mine is on an arch or dome-shaped structure, and at many places mineralization favored the crests of small folds. The ore bodies at the Sutro and Keystone mines and the Old Mine ore body at Iron Mountain are probably on the flanks of folds.

Bedding-plane foliation and fracture cleavage had a considerable effect in localizing ore bodies. The intersection of steep fracture cleavage with gently dipping bedding-plane foliation has provided a shattered area with an impervious cap that has localized some ore bodies.

Some faults formed before mineralization and acted as channelways for ore-bearing solutions. They generally cut the folds and the foliation at a considerable angle and influenced localization of ore bodies in certain parts of the folds.

Many favorable areas where the ore zone has not been eroded remain to be explored, but other large areas can be eliminated for geologic reasons. Areas that are worthy of exploration are those within the main northeastward-trending mineral belt that contain the middle unit of the Balaklala rhyolite; those that can be eliminated are areas in which the middle unit is missing because of erosion or because of original lenticularity and nondeposition, and areas covered by a considerable thickness of younger sediments.

The stratigraphic sequence is the principal feature used in delimiting areas in which new ore bodies may be found. In areas where the rocks of the middle unit have been eroded or were not deposited it appears most unlikely that prospecting would lead to the discovery of bodies of massive sulfide ore. The lateral controls that can be used for prospecting are much less definite, and although crests and troughs of folds are probably more favorable than the flanks of folds, the empirical method of testing the ore-bearing zone between known ore bodies, and testing for extensions of ore in known mines appears to be the most promising. Detailed mapping of folds, foliation, zones of hydrothermal alteration, and possible feeder fissures would probably lead to more detailed controls for guiding exploration.

INTRODUCTION

The Shasta copper-zinc district of northern California has yielded 54 percent of the copper produced in the State through 1946 (Eric, 1948, p. 202). The western part of the district, which has produced most of the ore, is called the West Shasta copper-zinc district and is the subject of this report. This base-metal mining district lies west of the Sacramento River, in the foothills of the Klamath Mountains that border the northwest end of the Sacramento Valley (fig. 1).

The district is the western part of the ore-bearing area formerly known as the Shasta copper belt, or in the older writings as a "copper arc." It was thought to be crescent-shaped and to extend entirely around the head of the Sacramento Valley with the convex side of the arc to the north. Recent studies have shown that the so-called copper arc consists of two districts at either end of the arc that have distinctly different geologic structures and ore occurrences, and that there is only sporadic and unconnected copper mineralization between the two districts. In recent publications (Kinkel and Albers, 1951; Kinkel and Hall, 1951) the two principal base-metal districts have been named the West Shasta copper-zinc district and the East Shasta copper-zinc district. The West Shasta base-metal district extends from the Iron Mountain mine at the south to the Sutro mine at the north. The East Shasta base-metal district includes the area from the Afterthought and Donkey mines near the settlement of Ingot on U. S. Highway 299 east to the Bully Hill, Rising Star, and Copper City mines on the north side of the Pit River, 9 miles to the northwest.

The location of the West Shasta copper-zinc district and the principal base-metal mines are shown in figure 2. The area mapped includes the Igo, Whiskytown, and Shasta Dam 7½-minute quadrangles, and approximately the southern two-thirds of the Behemotosh Mountain quadrangle. The northern part of the Behemotosh Mountain quadrangle was not mapped because it is remote from the mineral district and is underlain entirely by unmineralized rocks of the Bragdon formation. The mines are in a northeastward-trending area about 8 miles long and 2 miles wide in the Whiskytown, Shasta Dam, and Behemotosh Mountain quadrangles.

The West Shasta copper-zinc district overlaps and includes several gold-producing districts, but this report does not include a separate description of them. These districts are described in reports by Ferguson (1914, p. 22–79) and Averill (1933, p. 5–73), and in many reports of the California Division of Mines.

Parts of the mapped area are crossed by State and Federal highways, and the main line of the Southern Pacific Railroad is only a few miles east of the mining district (fig. 1). Redding, Calif., 2½ miles south of the Shasta Dam quadrangle (fig. 2), is the county seat of Shasta County, and is the principal city at the head of the Sacramento Valley. In 1950 the population was 10,734. Redding is at the junction of the main north-south U. S. Highway 99, and the east-west U. S. Highway 299. Other small towns and settlements in the mapped area are Igo, Whiskytown (Schilling), Summit City, Buckeye, and the U. S. Bureau of Reclamation headquarters at Toyon. The old smel-

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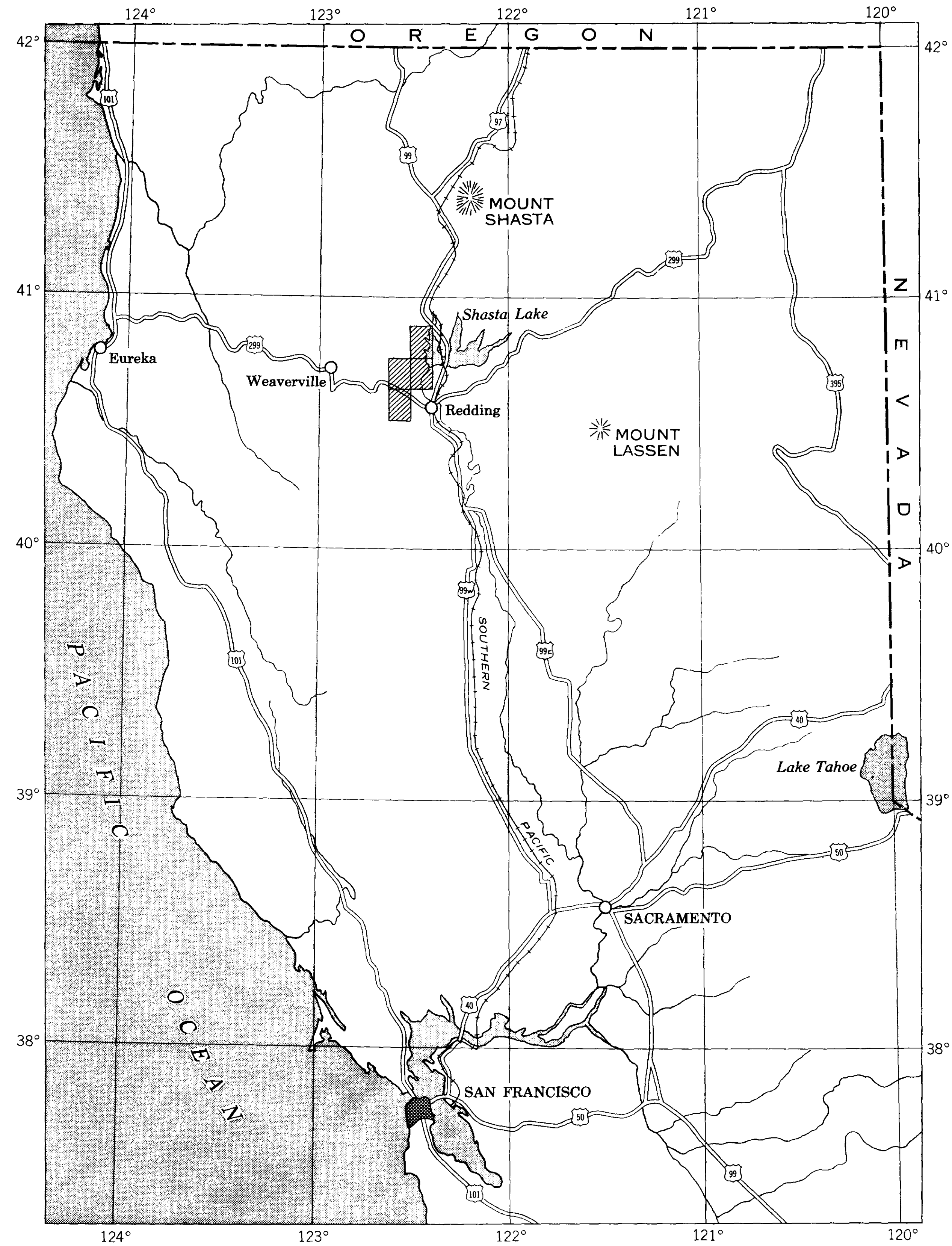

FIGURE 1.—Index map showing the location of the West Shasta copper-zinc district.

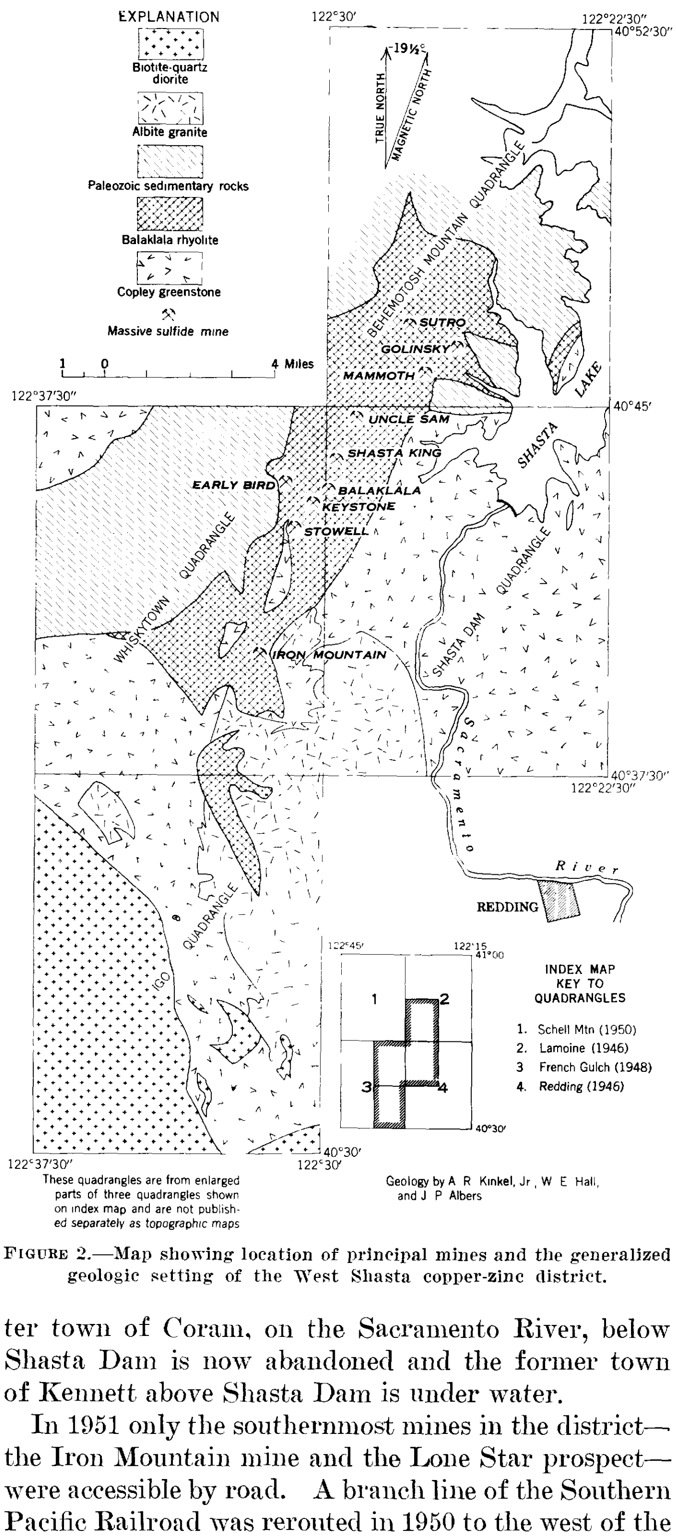

FIGURE 2.—Map showing location of principal mines and the generalized geologic setting of the West Shasta copper-zinc district.

ter town of Coram, on the Sacramento River, below Shasta Dam is now abandoned and the former town of Kennett above Shasta Dam is under water.

In 1951 only the southernmost mines in the district—the Iron Mountain mine and the Lone Star prospect—were accessible by road. A branch line of the Southern Pacific Railroad was rerouted in 1950 to the west of the Sacramento River as far as Shasta Dam to provide rail service to the dam and loading facilities for the ore from the Iron Mountain mine.

The Iron Mountain mine, which was the only mine operating in the West Shasta district in 1952, can be reached by a surfaced road that branches north from U. S. Highway 299W about 3 miles west of Redding. The Lone Star prospect is accessible along a State Forest-Service road either by way of the South Fork Mountain fire-lookout station, or along a dirt road that extends past the Iron Mountain mine. The old road from the former smelter town of Coram, just below Shasta Dam, to the Balaklala mine was repaired in 1951 so that it was possible to drive from Coram to the Balaklala mine; this is a steep one-way dirt road that washes out during the rainy season. The road from the Squaw Creek arm of Shasta Lake to the Uncle Sam and Clipper gold mines was passable in 1950 but is narrow and tortuous and parts of it are covered by wash and debris each spring. Other mines in the district are accessible only on foot or with horses after ferrying across Shasta Lake.

Private boat service is available on Shasta Lake, and the U. S. Bureau of Reclamation has at times supplied barge service to some of the mine owners. However, a road along the west side of Shasta Lake to make accessible the central and northern part of the mineral belt for renewed operations in the district was under construction in 1952.

CLIMATE AND VEGETATION

Climate and vegetation vary greatly within short distances in the West Shasta district because of the difference in topography. The southeast edge of the mapped area is at the end of the Sacramento Valley and has an altitude of 700–800 feet, whereas the western and northwestern parts of the area are in the rugged Klamath Mountains that reach an altitude of 5,189 feet at South Fork Mountain. Temperatures at lower altitudes are high in the summer but moderate in the winter. Daytime temperatures of 110°F are not unusual from June through September; higher temperatures have been recorded. A period of 10 days to 2 weeks of high temperature in the summer is generally followed by a comparable period in which the daytime temperature does not exceed 95°.

Table 1 gives data on the temperature at Redding and at two localities in the mountainous area nearby; Mount Shasta City is 50 miles airline north of Redding at an altitude of 3,555 feet and Weaverville is 29 airline miles west of Redding at an altitude of 2,047 feet. The humidity at the head of the Sacramento Valley is very low in the summer.

Winter temperatures in the Sacramento Valley drop several degrees below freezing a few nights of the year but hard freezes that cause trouble with exposed waterlines occur only at the higher altitudes. Winter temperatures well above freezing are the rule.

TABLE 1.—*Temperature and precipitation, Redding, Calif., and nearby localities*

[Data furnished through courtesy of E. L. Felton, U. S. Weather Bureau, 1952]

Locality	January	February	March	April	May	June	July	August	September	October	November	December	Annual
Average temperature (°F)													
Redding	45.3	49.3	53.9	59.8	66.8	75.0	82.0	80.6	74.0	64.6	54.3	46.7	62.7
Mount Shasta City	33.4	36.8	40.8	46.4	53.4	60.6	67.9	66.4	58.8	51.2	41.4	34.8	49.3
Weaverville	37.4	42.0	46.4	51.3	57.8	64.7	71.4	70.0	63.1	54.5	44.4	38.3	53.4
Average maximum temperature (°F)													
Redding	53.8	58.3	63.9	71.2	79.1	88.0	96.6	95.4	87.9	76.8	64.2	55.2	74.2
Mount Shasta City	41.6	46.6	52.9	59.2	68.2	75.2	84.8	84.5	75.9	66.1	52.7	43.9	62.6
Weaverville	47.2	54.5	61.3	68.0	76.4	85.1	94.6	94.0	85.8	73.9	57.7	47.2	70.5
Average minimum temperature (°F)													
Redding	37.0	40.3	43.7	48.3	54.2	61.3	67.3	65.5	60.1	52.5	44.2	38.4	51.1
Mount Shasta City	24.5	26.7	29.4	33.6	38.7	44.3	49.0	46.9	41.9	36.4	30.0	25.7	35.6
Weaverville	27.6	29.6	31.4	34.9	39.2	44.2	48.1	45.7	40.3	34.8	31.1	29.3	36.4
Average precipitation (inches)													
Redding	7.19	5.99	4.96	2.93	1.83	0.86	0.11	0.06	0.67	2.26	4.06	6.48	37.40
Mount Shasta City	6.07	5.29	4.65	2.65	1.81	.94	.17	.22	.85	2.46	3.95	5.52	34.58
Weaverville	6.59	5.61	3.99	2.79	1.50	.82	.13	.12	.59	2.26	4.89	6.49	35.78
Volmers	10.74	11.85	9.21	4.97	3.01	1.88	.23	.14	.65	6.32	6.35	11.63	66.98

The amount and type of precipitation also varies with the altitude and the season. Snow is common above an altitude of 1,500 feet, and the higher mountains in the district are snow covered most of the winter. During some winters snow falls at the valley level but seldom lasts more than a few days. Precipitation is rare even at higher altitudes from the middle of May until the middle of October. The average precipitation at Redding and at nearby points is given in table 1. The weather station at Volmers (Delta) 24 airline miles north of Redding is an area of unusually high precipitation; several of these areas of high precipitation occur north and west of Redding.

The lower parts of the area are covered by a growth of chaparral, including much manzanita (*Arctostaphylos* sp.) that locally is so dense as to be almost impenetrable. Digger pine (*Pinus sabiniana*), ponderosa pine (*Pinus ponderosa*), sugar pine (*Pinus lambertiana*) and Douglas-fir (*Pseudotsuga taxifolia*), California black oak (*Quercus kelloggii*), canyon live oak (*Quercus chrysolepis*) and interior live oak (*Quercus wislizenii*) grow in the district, but smelter smoke, disastrous fires, and logging near the head of the valley have destroyed the natural balance of vegetation. Most of the lower slopes are subject to rapid erosion because of the loss of ground cover.

The land at higher altitudes was originally heavily timbered with coniferous trees, but much of the best timber was cut for mining operations. Patches of good timber still remain and second-growth timber is coming back in some areas that were denuded by fire or by logging, but in other areas a brush cover has inhibited second-growth timber.

PHYSICAL FEATURES

Most of the West Shasta district is in the Klamath Mountains that border the Sacramento Valley. Part of the district is underlain by a cut surface and a constructional surface consisting of gravel of the Red Bluff formation and is generally considered a part of the valley. The two principal drainage systems are those of the Sacramento River and of Clear Creek. Throughout most of the district these streams have cut deep canyons, although traces of older broad valleys remain at some places. Tributaries to the main streams make a pattern of rugged youthful canyons separated by long spur ridges. The hillsides are steep; many slopes are near the angle of repose whereas others have not reached this stage and have little or no soil or rock debris.

The mountainous areas fall into three groups. A low foothill range east of Clear Creek, dominated by Mule Mountain at an altitude of 2,330 feet, lies on the west side of the Sacramento Valley in the Igo quadrangle. A second group, the high mountainous area west of Clear Creek, part of which is included in the Igo quadrangle, reaches an altitude of 6,962 feet at Bully Choop Mountain west of the mapped area. The mountains in the northern and western part of the district comprise the third group. They are separated from the Mule Mountain range by the low divide between Whiskytown and Old Shasta and by several higher divides west of Whiskytown. Rounded, subdued landforms predominate in the Mule Mountain range although the canyons are steep and V-shaped. The Shasta Bally range also contains many rounded landforms; rugged topography is rare except in the V-shaped canyons. The topography in the mountainous area in the north and northwestern part of the district, on the other hand, is generally rugged except along the crest of ridges. The rounded crest of these ridges, as seen from the valley, is deceptive and conceals the fact that many of the slopes below the ridge are steeper than the angle of repose.

PREVIOUS WORK AND ACKNOWLEDGMENTS

Many geologists have worked on the geology of Shasta County and the surrounding area, and without this background the work on this report would have been much more difficult. Some features of the geology of the Shasta County base-metal district were described in connection with the gold-mining activity, which started a few months after the discovery of gold at Sutters Mill in El Dorado County, but no major study of the geology of the district, except for stratigraphic studies by Smith (1894, p. 588–612) and Hershey (1901, p. 225–245), was made until Diller in 1901–4 mapped the Redding quadrangle for the U. S. Geological Survey. The results of his work were published in 1906 (Diller, 1906), and have been invaluable for all later work.

The writers were continually impressed during the course of their field studies by Diller's accurate observations and his sound conclusions, which were based on a broad knowledge of the geology of northern California. Diller's emphasis was on the structure and stratigraphic sequence in the district; he gave only a cursory study to the base-metal mines that were at that time just beginning their main period of operation. The rapidly increasing importance of copper mining in the district in the early part of the century made it advisable for the U. S. Geological Survey to continue its work here and to put more emphasis on mineral possibilities. In 1906–7 Graton (1909, p. 71–111) mapped areas around the principal mines in the West Shasta district in more detail and on larger scale topographic maps than were available to Diller.

Ferguson (1914, p. 22–79) and Averill (1933, p. 3–73) described the gold deposits of the district, and added information on the general geology. Hinds (1933, p. 77–122) reported on the geology of the district and extended his geologic study to the west and northwest of the area covered by Diller. During 1932–34 G. F. Seager collected much information on the base-metal mines of the district and mapped areas in the central part of the West Shasta District. This information is given in an unpublished report of the California Division of Mines, 1934.

This report is the result of a cooperative project with the State of California, Department of Natural Resources, Division of Mines. Field work by the senior writer was begun in November 1945. J. P. Albers was with the project from 1946 to 1949 when he left to begin geologic work in the East Shasta copper-zinc district. W. E. Hall was assigned to the project in 1949 and remained until the work was completed. The writers were assisted in the work by R. F. Johnson during 1948–49 and by Juan Rossi and Victor Hollister in the summer of 1948. Field work was completed in 1951. A. R. Kinkel, Jr. and W. E. Hall prepared the report in 1951–52.

The writers acknowledge their debt to the staffs of all the mining companies in the district. Access to all the information compiled by the mining companies as well as access to the mines was given to the writers, in addition to many suggestions and discussions on the geology of the mines. Without this cooperation, the present study would have been impossible. R. T. Walker and W. J. Walker provided data on the Shasta King mine as well as production data on some of the other mines in the district and they engaged in many stimulating discussions of the geology in the field with the writers. Data were obtained from geologists of the Coronado Copper and Zinc Co., particularly on sampling and mine mapping that was done in 1948 by the West Shasta Exploration Co., a subsidiary of the Coronado Copper and Zinc Co. Maps of all the mines owned by the United States Smelting Refining and Mining Co., as well as production and assay data, and much geologic information, were furnished by that company. R. N. Hunt of the United States Smelting Refining and Mining Co. contributed information on the geology of the Mammoth mine. W. A. Kerr, owner of the mines that were formerly owned by the Balaklala Consolidated Copper Co., made data available on the Balaklala and Early Bird mines, and the Balaklala Angle Station gossan. Much use has been made of unpublished material collected by G. F. Seager in 1934 on the mines of the district. Some data available to him at that time were not available at a later date, and his compilation of information on the Balaklala mine was particularly useful.

The writers wish to thank the staff of the Iron Mountain mine of the Mountain Copper Co., Ltd. for many courtesies and for aid in compiling the records of the operation of that mine for the past 50 years. Much detailed information on the mine was furnished by C. W. McClung, T. P. Bagley, and R. K. McCallum.

Particular credit is due O. H. Hershey [1] for information on the geology of the underground workings at the Iron Mountain mine. Hershey collected information as a consulting geologist for many years at Iron Mountain during the earlier years of the mine's operation, and his unpublished reports and drill logs were used in the writers' study of the Iron Mountain mine.

Most of the ore bodies of the district were inaccessible at the time of the study, and had been so for many years. The cooperation and interest of the mining companies in compiling all records that were still available so that a complete picture of the metallization of the district could be had has aided the writers greatly.

GEOLOGIC FORMATIONS

The formations in the West Shasta copper-zinc district range in age from Devonian(?) to Recent (pl. 1). The oldest formation exposed is the Copley greenstone of probable Middle Devonian age. It is composed of intercalated volcanic flows, pillow lavas, volcanic breccias and tuffs, most of which are intermediate or basic in composition, and minor, thin, lenticular beds of tuff and shale. The formation is at least 3,700 feet thick, but the base is not exposed in the mapped area. Lack of a distinctive horizon marker in the greenstone and the destruction of primary features by metamorphism makes it impossible to estimate thickness in most parts of the district.

The Balaklala rhyolite is a group of light-colored soda-rich rhyolitic flows and pyroclastic rocks that overlie the Copley greenstone conformably. Pyroclastic rocks make up about one-fourth of the formation. The Balaklala rhyolite formed as a broad elongate volcanic dome, extruded from many vents; the formation is much less extensive laterally than the Copley greenstone. Nonporphyritic rhyolite and porphyritic rhyolite containing phenocrysts of quartz and plagioclase are lithologic subdivisions; the porphyritic rhyolite is further subdivided into flows that contain small phenocrysts and those that contain large phenocrysts. It is possible to map a stratigraphic sequence in the Balaklala rhyolite by using these lithologic subdivisions along with bedded pyroclastic rocks. All the known base-metal ore bodies of the West Shasta district are in the Balaklala rhyolite.

Two features on the geologic maps require explanation. One is the inclusion of all rhyolitic rocks with the Balaklala rhyolite, whether they are intrusive or extrusive masses; the other is the subdivision of the Balaklala rhyolite on a lithologic rather than on a stratigraphic basis.

[1] Hershey, O. H., Private reports prepared for the Mountain Copper Co., Ltd.

On the areal geologic map and on most of the geologic maps of mine areas, intrusive rhyolite is not distinguished from extrusive rhyolite. Subdivisions of the Balaklala rhyolite that can be mapped are based on lithologic differences rather than on a stratigraphic succession. Nevertheless in spite of local recurrence of lithologic types, these subdivisions fall into a sequence sufficiently common and distinctive to be referred to as the lower, middle, and upper units of the Balaklala rhyolite. All rhyolitic rocks, whether they are the rhyolitic flows of the Balaklala or rhyolitic rocks of the Balaklala type in the underlying Copley greenstone, are shown as Balaklala rhyolite. Although the rhyolitic rocks are easily distinguished at most places from the mafic rocks of the Copley greenstone, the mode of emplacement of the rhyolitic rocks is not everywhere evident. Unless pyroclastic material is present, or unless the rocks are exceptionally well exposed, it is commonly impossible in the field to determine whether a particular sheet of rhyolite was intruded into the Copley greenstone as a dike or a sill, whether the sheet of rhyolite represents an early flow of rhyolitic material in the Copley, or whether the rhyolite is a flow that overlies the Copley and is part of the Balaklala rhyolite. For this reason, all the rhyolitic material is given the symbol for Balaklala rhyolite.

The Balaklala rhyolite is subdivided into lithologic types even though three stratigraphic units are recognized in the Balaklala, because the determination of stratigraphic units is impossible outside the central part of the district where all the units are present. The Balaklala rhyolite is a volcanic pile that commonly contains many lithologic types at one general horizon; the repetition of lithologic types from the same vent and repetition caused by overlapping flows and pyroclastic material from different vents, and the lenticular nature of the volcanic units make it impossible to determine the stratigraphic position of an individual flow or pyroclastic bed without taking the sequence into consideration.

The Kennett formation, which is composed of black cherty shale, tuff, and limestone, conformably overlies the Balaklala rhyolite where the latter rock was deposited. The Kennett overlies Copley greenstone locally where no Balaklala is present. It was dated by Schuchert as Middle Devonian in age (Diller, 1906, p. 2) on the basis of abundant fossils collected from the limestone. The greatest thickness of Kennett formation is in Backbone Creek in the northern part of the West Shasta district where Diller measured a section that was 865 feet thick. However, the actual thickness is less, as the beds are repeated by faulting in the area where the section was measured.

The Bragdon formation, which is dated as Mississippian in age by Diller (1906, p. 3), rests upon the Kennett formation and shows no evidence of a major unconformity in the West Shasta district. However, outside this district warping uplifted the Kennett, and in these areas of uplift an erosional unconformity is present between the Kennett and the Bragdon formations. The Bragdon is predominantly shale but contains conglomerate and sandstone. North of Backbone Creek shale and conglomerate of the Bragdon formation is 3,500 feet thick where mapped by the writers, but this represents only the lower part of the formation, which continues north of the mapped area for about 11 miles. The Bragdon formation may be 6,000 feet thick in the Redding 30-minute quadrangle, according to Diller (1906, p. 3).

Two large plutons intrude the layered rocks of the district. The older pluton is the Mule Mountain stock consisting largely of albite granite, which intrudes the Copley greenstone and the Balaklala rhyolite. This stock was intruded as an elongate body the shape of which was determined by structures formed in the enclosing rocks by the Nevadan orogeny. It is syntectonic but later movements have locally sheared the albite granite near its margins. The younger pluton is the Shasta Bally batholith composed of a biotite-quartz diorite. This younger pluton intrudes the Copley greenstone, the Bragdon formation, and the albite granite in the West Shasta district. The biotite-quartz diorite is dated by Hinds (1934, p. 182–192) as Late Jurassic in age. It is overlain outside the mapped area nonconformably by strata that belong to the Paskenta and Horsetown formations of Early Cretaceous age. The albite-granite pluton was named the Mule Mountain stock by Hinds (1933, p. 105) and the biotite-quartz diorite pluton was named the Shasta Bally batholith by the same author.

Hypabyssal intrusive rocks include dikes and sills of diabase, andesite porphyry, lamprophyre, hornblendite, diorite porphyry, dacite porphyry, quartz latite porphyry, and porphyritic rhyolite. The porphyritic rhyolite dikes are pre-Mississippian in age as they do not intrude the Bragdon formation; some were probably feeders for the Balaklala rhyolite flows. Dikes other than the rhyolite cut the Bragdon formation and are post-Mississippian in age.

Cemented gravels of the Chico formation of Late Cretaceous age and of the Red Bluff formation of Pleistocene age unconformably overlie the Copley greenstone in the southeast corner of the Shasta Dam and Igo quadrangles. The geologic column in the West Shasta copper-zinc district is shown in figure 3.

Age	Formation	Lithologic description	Thickness (in feet)
Recent		Alluvium, surface mantle, landslide deposits, gravel	0–150
Pleistocene	Red Bluff formation	Partly to well-cemented gravel	0–100
Cretaceous	Chico formation	Sandstone, shale, conglomerate	0–200
Late Jurassic or Cretaceous	Hypabyssal intrusive	Hornblendite, lamprophyre diorite, dacite, porphyry, quartz latite porphyry, andesite porphyry; diabase	
	Shasta Bally batholith	Biotite-quartz diorite	
Jurassic(?)	Mule Mountain stock	Albite granite	
Mississippian	Bragdon formation	Shale, sandstone, conglomerate	3500+
Middle Devonian	Kennett formation	Limestone, black cherty shale, tuff	0–400
	Balaklala rhyolite	Porphyritic and non-porphyritic rhyolite, rhyolitic pyroclastic rocks	0–3500(?)
Middle Devonian(?)	Copley greenstone	Greenstone, keratophyre, meta-andesite, pyroclastic rocks, metagabbro	3700+

FIGURE 3.—Geologic column in the West Shasta copper-zinc district, Shasta County, Calif.

The Paleozoic rocks in the West Shasta copper-zinc district are folded into a broad arch that contains many small folds and forms a broad, low anticlinorium. The axis of the anticlinorium trends N. 15° E. in the central part of the mineral belt and passes through Iron Mountain and Behemotosh Mountain (pl. 1). Beds on the flanks of this fold generally dip at angles of 20°–30°, but locally they may be vertical; beds in the central part of the district generally dip at low angles. The broad arch has a gentle culmination in the central part of the mining district at the Uncle Sam mine; in the northern part of the district, north of the Mammoth mine, this structure plunges gently to the north. The arch is broken by many faults and at the south end by intrusive masses. Two sets of faults are prominent; one set strikes N. 20°–45° W., and the other strikes N. 60°–80° E. Both vertical and horizontal movements are recognized; in nearly all the faults the north block is downthrown.

COPLEY GREENSTONE

DISTRIBUTION

The Copley greenstone is the basement rock in the West Shasta copper-zinc district and is the most extensively exposed formation. It was originally called Copley meta-andesite by Diller but the name was changed to Copley greenstone by Kinkel and Albers (1951) because the latter name is more expressive of the lithology. The principal areas underlain by this formation are in the Shasta Dam quadrangle, of which 70 percent is underlain by Copley, and in the Igo and Whiskytown quadrangles, where the Copley greenstone is exposed in a belt 9 miles long striking N. 10° W. between the albite granite and the biotite-quartz diorite plutons. Windows of Copley greenstone are also exposed through the overlying Balaklala rhyolite in deep canyons that cut through the mining district. It immediately underlies the Bragdon formation in part of the Whiskytown quadrangle, and is brought to the surface by a large fault in the northwest quarter of that quadrangle.

Although the structure of the copper-zinc district is that of a broad low arch with its axis through the heart of the mining district, the younger Balaklala rhyolite crops out mainly along the crest of the arch and the older Copley greenstone crops out as parallel bands on the flanks of the arch. This apparent anomaly is caused by the rugged topography, the pinching out of the rhyolite toward the flanks of the arch, and by faulting. The Balaklala rhyolite, which is thick at the crest of the arch and is thin on the flanks, forms the crests of the rugged hills in the mining district whereas the greenstone forms the lowlands of broad rolling hills.

THICKNESS AND RELATIONSHIP TO OTHER ROCKS

Partial sections of the Copley greenstone have been measured by Louderback (1928, p. 56–59) and by Hinds (1933, p. 87). Louderback measured a partial section, now flooded, near Kennett that showed a thickness of 2,000 feet; Hinds measured two sections: one on Stacy Creek measured 1,200 feet, and the other on Shirttail Peak measured 1,500 feet. The writers measured a partial section of Copley, 3,700 feet thick, in Modesty Gulch in the Whiskytown quadrangle.

The Copley greenstone may be 6,000 feet thick in the belt that is exposed between the albite granite and the biotite-quartz diorite plutons in the Igo quadrangle. The biotite-quartz diorite intrudes the Copley in the southwestern part of the Igo quadrangle, and here it has metamorphosed part of the Copley to hornblende schist, amphibolite, and migmatite. The foliation has a uniform N. 30° W. strike and dips 60° NE. in this belt and except for local transgressions is parallel to the biotite-quartz diorite contact. This foliation may also be parallel to bedding, although in most places in this belt bedding in the Copley is obscure. The only definite bedding is near Brandy and Boulder Creeks where a few shaly tuff beds were seen to be conformable to foliation. If the foliation is parallel to bedding in most of the belt, a section of Copley 6,000 feet thick is exposed. However, much of the banding is caused by metamorphic differentiation by solutions traveling along planes of foliation and it may or may not be parallel to bedding.

The Copley greenstone is overlain by Balaklala rhyolite and has a gradational contact with the Balaklala. A pyroclastic layer ranging from a thin bed to as much as 150 feet in thickness is present at many places at or near the top of the Copley. The pyroclastic layer contains rounded fragments of Balaklala-type nonporphyritic and porphyritic rhyolite in a tuffaceous, andesitic matrix. It is conformable with the overlying Balaklala rhyolite.

A few thin bands of Copley-type greenstone flows and pyroclastic rocks are interbedded in the lower 200 to 300 feet of the Balaklala rhyolite. They indicate that eruption of rocks of intermediate composition continued sporadically after eruption of Balaklala rhyolite started and they are regarded as part of the Balaklala rhyolite.

The Copley greenstone is cut by many intrusive bodies. The Mule Mountain stock intrudes the Copley in the eastern part of the Igo quadrangle and in the southern parts of the Whiskytown and Shasta Dam quadrangles, and the Shasta Bally batholith intrudes the Copley in the western third of the Igo quadrangle. The Copley is also cut by many dikes and sills of nonporphyritic and porphyritic rhyolite. Some of these dikes and sills were feeders for the Balaklala rhyolite flows and pyroclastic material. Others, consisting of sugary, aplitic rhyolite may be related to the stock of albite granite. The best exposures of these dikes are in the western part of the Shasta Dam quadrangle, on Copley Mountain, and in Spring Creek. Dikes of hornblendite, diabase, diorite porphyry, dacite porphyry, quartz latite porphyry, and lamprophyre also cut the Copley greenstone.

GENERAL DESCRIPTION

The Copley greenstone consists of interlayered volcanic flows, tuffs, agglomerate, and a few thin layers of tuffaceous shale and black shale of small areal extent. No distinctive horizon marker was recognized in the Copley, and the stratigraphic units are extremely lenticular. The lower and middle parts of the exposed section of Copley consist predominantly of fine-grained chloritic lava flows and tuff beds of keratophyric composition, and some shale. The upper part is composed largely of amygdaloidal pillow lava, fine to coarse pyroclastic material, and diabase. The top of the Copley consists of extensive but not continuous volcanic breccia as much as 150 feet thick, which generally is too thin to be shown on the quadrangle maps.

The Copley is strongly metamorphosed by the Shasta Bally batholith in the western part of the Igo quadrangle and is progressively less metamorphosed eastward away from the batholith. In the western part of the quadrangle, near the contact with the Shasta Bally batholith, the Copley is altered by contact metamorphism to amphibolite, epidote amphibolite, and hornblende gneiss and migmatite for as much as 4,000 feet from the contact. In some areas the Copley is strongly foliated, but in others it is massive; it is altered to chlorite-quartz-albite-epidote rock (green schist facies of Eskola) and few original minerals are preserved in the recrystallized rocks. However, megascopic textures are preserved except in the metamorphic zone along the Shasta Bally batholith, or where the Copley is foliated. East of the Sacramento River in the eastern part of the mapped area the rocks are much less foliated and altered.

The Copley is composed of keratophyre, spilite, and albite diabase, and some meta-andesite and meta-gabbro. In the field it was not possible to differentiate between these petrographically different types of rocks, and they were mapped on megascopically recognizable types such as greenstone flows, pillow lava, amygdaloidal greenstone, greenstone tuff, or coarser pyroclastic material. Nearly all these rocks are of keratophyric composition, however, except for a small amount of spilite and albite diabase. Meta-andesite, which cannot be differentiated in the field from keratophyre or spilite, crops out only in the eastern part of the area in the Shasta Dam quadrangle. The lithologic types are described, but the petrographic description is given separately because of the difficulty of correlating the two in the field.

LITHOLOGIC DESCRIPTION

Fine-grained chloritic lava and tuff.—The lower and middle parts of the Copley are mainly light to dark-green, fine-grained lava and tuff of keratophyric composition. Some flows are finely porphyritic and include phenocrysts of plagioclase, or chlorite pseudomorphic after hornblende, in an aphanitic, greenish groundmass. In the western part of the district much of the fine-grained Copley is massive and hard due to metamorphism by the Shasta Bally batholith and to a lesser degree by the Mule Mountain stock. The Copley is softer and more chloritic and is foliated in the central part of the district and in the western part of the Shasta Dam quadrangle. Locally the fine-grained Copley, especially fine tuffaceous material, is schistose; foliation is emphasized by weathering, and road cuts commonly expose hard, massive lava under a few feet of soft chloritic, schistose greenstone.

Amygdaloidal greenstones are abundant locally. The amygdules have smooth, rounded outlines, and rarely show evidence of stretching. Most of the amygdules are 4–5 millimeters in diameter, although some are as much as 2½ centimeters in diameter. Quartz, calcite, chlorite, epidote, clinozoisite, albite, and zeolites are common vesicle fillings. Calcite is commonly removed from calcite-filled amygdules by weathering leaving a vesicular rock.

Pillow lavas.—Pillow lavas are abundant in the upper 1,000 feet of the Copley, and in most places they are associated with volcanic breccia. Three varieties of pillow lavas occur in the district. One variety in Modesty Gulch in the Whiskytown quadrangle has pillows that average 2 feet in diameter. A cross section of an individual pillow shows a dense light-green core about 20 inches in diameter surrounded by a rim of darker green amygdaloidal lava a few inches thick. The amygdules are about a quarter inch in diameter and are filled with quartz or chlorite. Pillows of this variety are in a layer about 20 feet thick, which is underlain by about 200 feet of dark-green lava that weathers into rounded forms 1–2 feet in diameter that have no apparent internal structure.

In the second variety of pillow lava, also in Modesty Gulch, the cores of the pillows are light colored and siliceous and almost without exception contain much white vein quartz. The average major axis of the elliptical pillow cores is about 18 inches; the minor axis is about 8 inches. The cores are surrounded by a rim of variolite 2–3 inches thick. The variolite has rounded white spots, 2–3 millimeters in diameter, which make up 75 percent of the material of the rim, in a dark-green chloritic matrix. This pillow layer is about 100 feet thick in Modesty Gulch, where it was traced for more than a mile.

A third variety of pillow lava is exposed in Boulder Creek north of the Hornet mine at Iron Mountain. The pillows average 3 feet in diameter and have sharp-edged rims about 1 foot thick. The rims are a lighter shade of green than the centers and are slightly amygdaloidal.

Coarse volcanic breccia.—The areas of pyroclastic material shown on the geologic map do not have sharp boundaries at many places and contain smaller irregular areas of nonpyroclastic material. The map shows the principal areas that contain much pyroclastic material, but many small areas throughout the Copley in addition to those shown contain pyroclastic rocks. However, large bodies of greenstone do not contain pyroclastic material and for this reason an attempt has been made to differentiate these bodies from areas that are composed predominantly of pyroclastic material. Most of the coarse pyroclastic material lies in the upper part of the Copley greenstone. The volcanic breccia consists largely of a mixture of bombs, lapilli, irregular-shaped fragments, and ropy material in a greenish, aphanitic matrix having the same general appearance as the rock that forms the fragments. The bombs and fragments rarely are more than 6 inches in diameter and average about 4 inches. All gradations are present between ropy lava, lava with a few bombs and fragments, and pyroclastic layers composed almost entirely of bombs and fragments.

The bombs in the pyroclastic layers are commonly well rounded and are of slightly different color than the matrix (fig. 4). Some angular fragments as well as

FIGURE 4.—Volcanic breccia near the top of Copley greenstone in Grizzly Gulch in the Whiskytown quadrangle.

bombs in volcanic breccia have light-colored rims suggesting that the rims may have resulted from reactions between the fragment and the matrix or from fumarolic action.

A distinctive volcanic breccia is exposed at many places at the top of the Copley, and forms a transition between Copley greenstone and Balaklala rhyolite. This breccia contains rounded, light-colored fragments, generally 3–4 inches in diameter but locally as much as 1 foot in diameter, in a green chloritic matrix. Some fragments are porphyritic and nonporphyritic rhyolite; others which may be silicified greenstone, are light colored and have abundant quartz amygdules and chlorite flecks. This breccia is as much as 150 feet thick, and is well exposed in the head of Modesty Gulch south of Mad Mule Mountain but it is present at many other localities in the transition zone along the Copley-Balaklala contact.

Shale and shaly tuff.—Beds of shale are interbedded in the Copley greenstone east of the Old Diggings district in the Shasta Dam quadrangle; they are exposed in road cuts along the road to the Walker gold mine and north and south of the road. Some are black or gray siliceous shale that closely resembles the siliceous shale of the Kennett formation. These shale beds are interlayered with thin beds of tan sandy shale, shaly

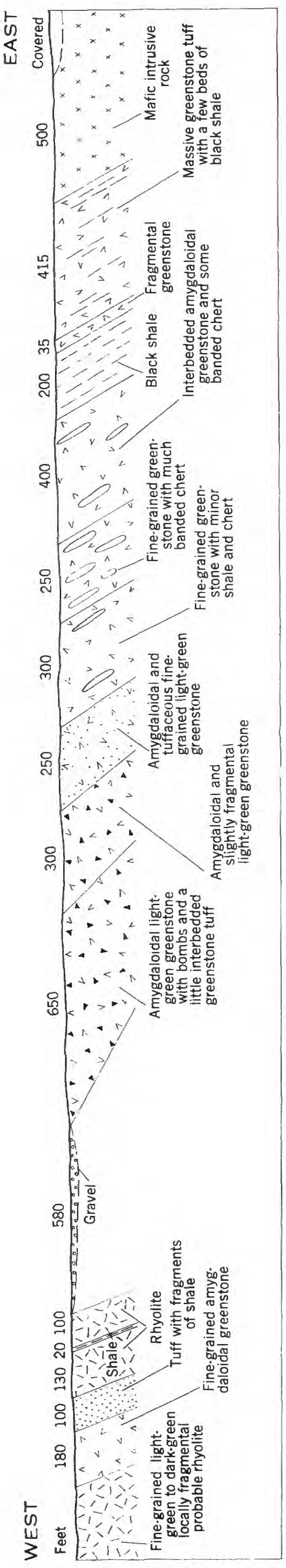

FIGURE 5.—Partial section of the Copley greenstone along the Sacramento River about 1½ miles N. 70° W. of Redding, Calif.

tuff, and Copley-type flows. A similar series of interbedded siliceous, cherty shale and greenstone crops out in the canyon of the Sacramento River outside the map area about 1½ miles west of Redding (fig. 5). These shale beds may be a continuation of the shale near the Walker mine. In some areas rhyolitic tuff and shale occur together in the Copley as in Brandy Creek and east of Coram.

A few thin beds of shaly tuff and black shale are exposed in road cuts on the logging roads near the head of Brandy Creek and in the canyon of Boulder Creek in the Igo quadrangle. Also, some of the strongly sheared rocks in the Copley probably are shaly tuff as they are associated with volcanic breccia. However, they are so altered that it is difficult to be certain of their original character.

PETROGRAPHIC DESCRIPTION

Keratophyres.—Most of the Copley greenstone is keratophyre. The keratophyres are greenish, aphanitic rocks that commonly are foliated and chloritized and locally are schistose. Some are silicified and are difficult to distinguish megascopically from Balaklala rhyolite. Although many of the primary structures and textures are destroyed, pyroclastic, amygdaloidal, and pillow structures, and pilotaxitic, porphyritic, and variolitic textures have been recognized where the rocks are not foliated.

Most of the keratophyres have a pilotaxitic texture and are composed of albite or sodic oligoclase, chlorite, secondary quartz, zoisite, clinozoisite, epidote, montmorillonite, green biotite, leucoxene, and small amounts of calcite, relict hornblende, apatite, and opaque minerals. Plagioclase, chlorite, and secondary silica commonly constitute more than 90 percent of the rock. The plagioclase is mostly cloudy owing to inclusions of chlorite and in places epidote or zoisite, and it may be either massive or lath shaped. It ranges in composition from $Ab_{97}An_3$ to $Ab_{78}An_{22}$.

Mafic minerals include chlorite, epidote, clinozoisite, zoisite, and relict hornblende. Hornblende is largely altered to chlorite and epidote, and only rarely is relict primary hornblende present. Chlorite is the predominant mafic mineral; it occurs as pseudomorphs after hornblende, as tiny flakes disseminated through plagioclase, as irregular patches and veinlets through the matrix, and as filling of amygdules. The chlorite has negative elongation and the indices of refraction are: $n_\alpha=1.617$ and $n_\gamma=1.622$; it probably is prochlorite. Epidote, clinozoisite, and zoisite are present as tiny inclusions in plagioclase and as fillings in some amygdules. They occur in quantities sufficient to account for an original plagioclase only slightly more calcic than the albite now present. Veinlets of quartz are common.

Amygdaloidal structures are common in the keratophyres (fig. 6). The amygdules are quartz, chlorite, epidote, calcite, zoisite, albite, and zeolites listed in approximately decreasing order of abundance. Most of the amygdules are 2 to 3 millimeters in diameter and are spherical or ellipsoidal. Some quartz amygdules are as much as 2.5 centimeters in diameter. Although distorted amygdules have been recognized in the western part of the district, most amygdules show no evidence of distortion.

FIGURE 6.—Photomicrograph of amygdaloidal keratophyre. Groundmass composed of laths of albite ($Ab_{90}An_{10}$) with interstitial chlorite and minor eipdote. Amygdule filling (*a*) is calcite and minor epidote and quartz. Crossed nicols, × 20.

Some of the keratophyres have a variolitic texture (fig. 7) instead of the more common pilotaxitic texture. The variolites are conspicuous megascopically on a weathered surface and stand out as round white spots 1 to 2 millimeters in diameter in a dark-green chloritic matrix. Some pillow structures have rims of variolite. The variolites are made up of radial growth of untwinned plagioclase that ranges in composition from $Ab_{92}An_8$ to $Ab_{85}An_{15}$ and contain 5–10 percent tiny disseminated grains of quartz and a little chlorite and zoisite. Variolitic growths constitute as much as 75 percent of the rock, and the rest is mainly interstitial chlorite.

FIGURE 7.—Photomicrograph of Copley greenstone showing variolitic texture. Variolites are sodic oligoclase surrounded by chlorite. Crossed nicols, × 14.

Spilite.—Spilite comprises only a small percentage of the Copley. Spilite that crops out on Vista Point in sec. 22, T. 33 N., R. 5 N., 1 mile south of Shasta Dam is a hard light-green aphanitic rock containing quartz-filled amygdules. The rock is finely porphyritic and has a few unaltered phenocrysts of augite and plagioclase in an aphanitic groundmass. The phenocrysts range from 1–2 millimeters in diameter.

Under the microscope the spilite is seen to be composed of phenocrysts of albite, augite, olivine, and epidote in a fine-grained holocrystalline groundmass made up of albite, augite, epidote, hornblende, and chlorite. Albite phenocrysts are clear unzoned crystals as much as 0.75 millimeter long and are twinned according to Carlsbad and albite twin laws. They have a composition of $Ab_{93}An_7$. Euhedral augite phenocrysts as much as 2 millimeters in diameter comprise about 5 percent of the rock. Some augite crystals are unaltered, whereas others are partly altered to epidote and chlorite. A few unaltered olivine phenocrysts are present, but most of the olivine is partly altered to chlorite.

The groundmass has a pilotaxitic texture, and is composed predominantly of albite, augite, and epidote. The albite laths in the groundmass are clear and unaltered and have sharp albite twinning. Some unaltered augite occurs in the groundmass, but most of it is altered to epidote and chlorite.

Composition of the spilite is given in table 2. Analyses of spilite from Oregon and California and a computed analysis of average spilite as given by N. Sundius are listed for comparison.

The spilite from Shasta County contains more alumina and magnesia and less ferrous iron than the average spilite as listed by Sundius.

TABLE 2.—*Chemical analyses of spilite*

Samples: 1. From sec. 22, T. 33 N., R. 5 W., 1 mile south of Shasta Dam, Calif. Analyst, R. N Eccher. 2. From NE¼ sec. 35, T. 7 S., R. 42 E., Baker quadrangle, Oregon.[1] Analyst, J. F. Fairchild. 3. Average spilite according to N. Sundius.[2]

Constituent	Spilite		Average spilite
	1	2	3
SiO_2	52.56	53.15	51.22
Al_2O_3	17.00	14.39	13.66
Fe_2O_3	.91	1.28	2.84
FeO	6.12	9.33	9.20
MgO	7.11	4.74	4.55
CaO	5.55	7.04	6.89
Na_2O	5.28	4.58	4.93
K_2O	.84	1.01	.75
H_2O^-	.10	.19	1.88
H_2O^+	3.15	2.02	----
TiO_2	.40	1.50	3.32
CO_2	.86	.10	.94
P_2O_5	.05	.19	.29
MnO	.16	.14	.25
BaO	----	0	----
FeS_2	----	Tr.	----
	100.09	99.66	100.72

[1] Analysis 8 (Gilluly, 1935, p. 234).
[2] Sundius, N. (1930).

The spilite differs from the keratophyres in showing more of a basaltic character. The spilite has relict olivine and augite and contains abundant epidote, whereas hornblende is the only relict mineral in the keratophyres, which are composed predominantly of albite and chlorite and contain very little epidote.

Meta-andesite.—The meta-andesites are hard, dark green to greenish-gray aphanitic rocks that commonly are finely porphyritic. The phenocrysts are mostly plagioclase but some are relict hornblende and augite. Amygdaloidal structures are common. Locally shearing is intense and the rock is altered to chlorite schist.

The meta-andesite is composed of plagioclase, chlorite, epidote, clinozoisite, augite, hornblende, montmorillonite and some quartz and accessory opaque minerals. It has a pilotaxitic texture (fig. 8). Plagioclase, epidote, clinozoisite, and chlorite commonly make up 90 percent or more of the rock. However, as much as 30 percent augite and hornblende are present in the least metamorphosed andesite. Augite commonly occurs as corroded relicts in epidote and clinozoisite, but may also be found as tiny euhedral phenocrysts. Green hornblende laths are also present in the least altered andesites, but they are replaced by chlorite and epidote in the meta-andesites.

FIGURE 8.—Photomicrograph of meta-andesite of the Copley greenstone with pilotaxitic texture. Rock is composed of plagioclase laths ($Ab_{71}An_{29}$) with interstitial chlorite, epidote, and magnetite. Crossed nicols, × 14.

Plagioclase typically makes a felted texture of lath-shaped crystals. In a single thin section plagioclase ranges in composition from albite to andesine. Clear plagioclase laths with sharp albite twin lines are andesine; in laths of oligoclase composition slight saussuritic alteration has occurred, and tiny epidote and zoisite inclusions are present. Albite that locally replaces the original andesine is very cloudy and is massive and untwinned and has negative relief.

In the field it is impossible to separate the keratophyres from meta-andesite. Texturally the two rocks are similar, but differences can be recognized under the microscope. Whereas chlorite is the predominant mafic mineral in the keratophyres, the predominant mafic minerals in the meta-andesite are epidote and clinozoisite, which replace augite and hornblende and are present as inclusions in plagioclase that has saussuritic alteration. The spilite is distinguished from the meta-andesite by its many unaltered phenocrysts of augite, albite, and olivine in a groundmass of albite, augite, and epidote. Much of the albite is in clear euhedral laths. The composition of the plagioclase in the meta-andesite ranges from albite to andesine; the albite was formed by saussuritic alteration. The meta-andesites have about the same mineralogic character as the spilite, although spilite contains relict olivine phenocrysts, and meta-andesite in contrast, has relict hornblende phenocrysts.

Diabase and albite diabase.—Locally diabase and albite diabase are present in the Copley greenstone, probably as sills and dikes. They are hard dark-green, fine-grained rocks composed of plagioclase, hornblende, augite, chlorite, epidote, calcite, and some quartz, apatite, and magnetite. Some of the diabase is porphyritic (fig. 9), and the texture is ophitic where the rock is unaltered.

All the diabase is altered to some extent, and part of it is completely altered to secondary minerals. Augite is mostly replaced pseudomorphously by green hornblende, but a few relict cores of augite are present. Epidote and chlorite are the predominant mafic minerals, and in some specimens they have completely replaced hornblende. Where the alteration is most intense, the original ophitic texture is nearly destroyed.

Plagioclase ranges in composition from albite to andesine and everywhere contains inclusions of epidote and zoisite. The plagioclase in the diabase in the eastern part of the Shasta Dam quadrangle is saussuritically altered and has many relict cores of andesine. There is no doubt in the eastern part of the Shasta Dam quadrangle that the origin of the albite is by saussuritic alteration of an originally more calcic plagioclase.

Some of the diabase in the Whiskytown quadrangle that is no more altered than that in the Shasta Dam quadrangle has albite as the only plagioclase, and is an albite diabase. The plagioclase contains some inclusions of epidote and zoisite but not enough to account for an originally andesine or labradorite composition by saussuritic alteration. The albite diabase has an ophitic texture similar to the diabase.

Metagabbro.—One specimen from north of the Gannim mine in the Whiskytown quadrangle is a metagabbro. It is probably part of a small body intrusive into the Copley but the outcrop is too poor to be certain of the shape. The rock has a granitoid texture and has an average grain size of about half of a millimeter. It is very altered and is composed of plagioclase, hornblende, epidote, chlorite, a little relict olivine, calcite, and magnetite (fig. 10). The plagioclase is albite or sodic oligoclase formed by saussuritic alteration. Hornblende is pseudomorphous after augite. Similar small bodies of metagabbro intrude the Balaklala rhyolite southwest of the Stowell mine, west of the Sutro mine, and in the Shasta Dam quadrangle.

Keratophyre tuff.—Pyroclastic rocks make up 25 percent or more of the Copley greenstone. They range in texture from tuff that is too fine grained to recognize the individual grains to coarse breccia containing fragments more than a foot in diameter. Most of the tuff

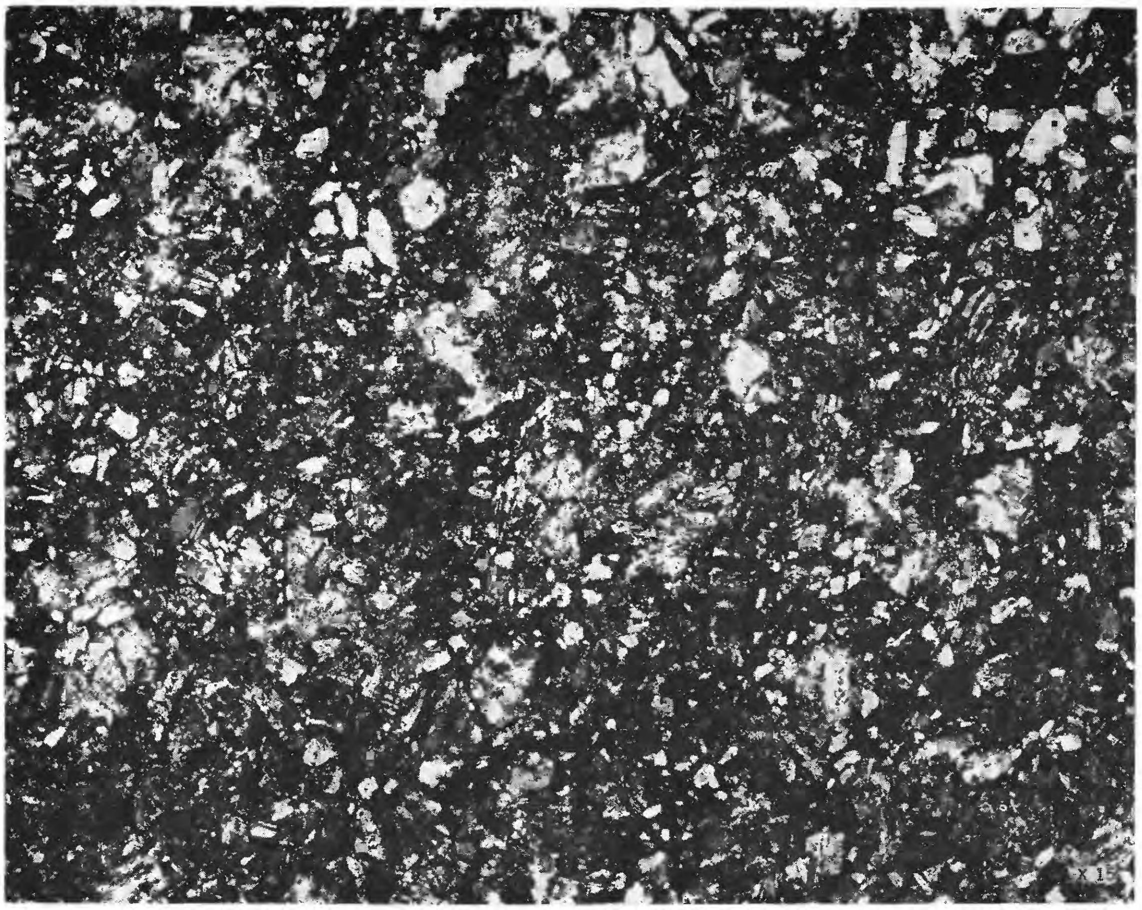

FIGURE 9.—Photomicrograph of altered porphyritic albite diabase. Crossed nicols, × 15.

beds are extremely sheared and are altered to fine-grained chlorite, epidote, quartz, albite, montmorillonite, and possibly other minerals too fine to recognize. Many beds that are extremely sheared and altered are probably tuff because they grade laterally or upward into coarse pyroclastic material.

AGE

The Copley greenstone is believed to be Middle Devonian in age, but no fossils have been found to accurately date it. It conformably underlies the Balaklala rhyolite, and the upper part of the Balaklala has been dated as of Middle Devonian age. Therefore the main part of the Copley is probably Middle Devonian or possibly slightly older.

Some of the rocks mapped as Copley along the east edge of the Shasta Dam quadrangle are younger than some of the Balaklala rhyolite, although all are greenstone with no distinct lithologic break between them and the main part of the Copley. Diller's Bass Mountain diabase (Diller, 1906, p. 7), which he considered to be Mississippian in age, lies mostly east of the mapped area, but extends into the southwestern part of the Shasta Dam quadrangle. Diller's description of the Bass Mountain diabase is as follows:

> The Bass Mountain diabase of the southern slope of Bass Mountain is generally a dark, somewhat greenish, compact lava which is not porphyritic, but is occasionally vesicular and more frequently fragmental. Where fresh this lava has darker spots of augite embedded in the lighter colored groundmass.

Diller correlated this band of greenish lava and pyroclastic material, which crops out south and east of Bass Mountain and extends to the southeastern part of the Shasta Dam quadrangle, with what he considered to be flows interbedded with the Bragdon formation in the vicinity of Middle Salt Creek in the Lamoine 15-minute quadrangle 8 miles north of Backbone Creek. The band he mapped as Bass Mountain diabase southeast of Bass Mountain underlies the Bragdon formation (Diller, 1906, p. 7). Therefore the correlation of this band with beds interbedded in the upper part of the Bragdon seems doubtful.

Hinds (1933, p. 91) called the Bass Mountain diabase of Diller's Bass Mountain basalt because of the surficial origin of the rocks. Hinds states that the Bass Mountain basalt rests unconformably on the Copley meta-

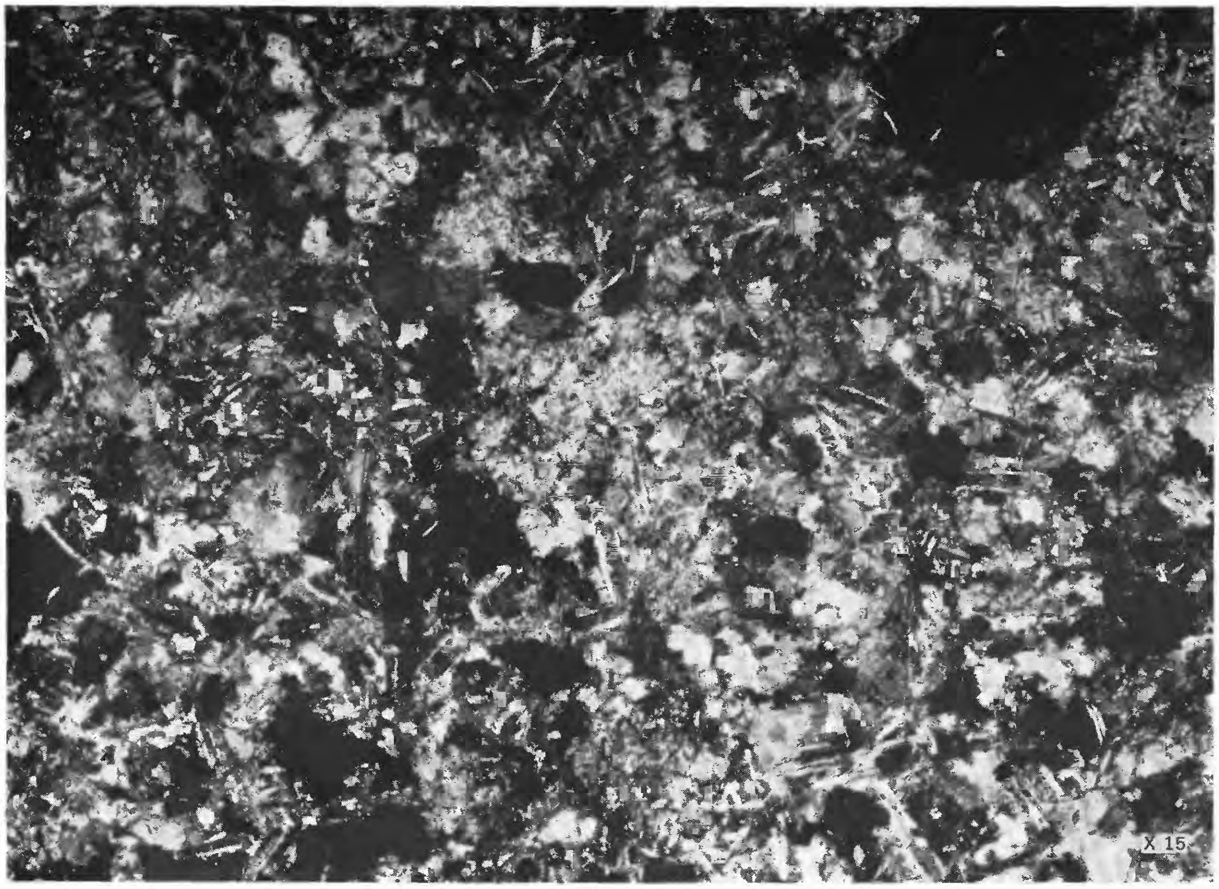

FIGURE 10.—Photomicrograph of metagabbro showing plagioclase laths showing saussuritic alteration. Crossed nicols, × 15.

andesite and the Kennett formation and is interbedded with and overlain by Bragdon strata in the area from Bass Mountain southward to Newton. The basis for this statement is not known, as he does not show any Bragdon in this area on the map, and it is not further explained in the text.

The writers were not able to map a lithologic or a time break in the greenstone in the mapped area, and they consider all the greenstone to be Copley, with the Copley becoming less altered and less metamorphosed to the east, although recognizing that it became younger also. J. P. Albers and J. F. Robertson, who in 1951 mapped the southeastern part of the Lamoine 15-minute quadrangle adjoining the West Shasta district, mapped the continuation of the same rock units (the Balaklala rhyolite and slightly metamorphosed greenstone pyroclastic rocks) that crop out along the northeastern border of the Shasta Dam quadrangle in the West Shasta district. They found tuff beds in the Balaklala that strike northeast and dip 40°–60° SE. under slightly metamorphosed greenstone that correlates with the pyroclastic greenstone along the eastern border of the Shasta Dam quadrangle shown on plate 1. These greenstone pyroclastic rocks, in turn, dip under shale of the Kennett and Bragdon formations. Thus it seems probable that flows and pyroclastic rocks of intermediate to basic composition were being erupted in the Shasta Dam quadrangle while Balaklala rhyolite type flows and pyroclastic rocks as well as flows and pyroclastic materials of intermediate composition were erupted to the northeast in the Lamoine quadrangle. The pyroclastic greenstone in the eastern part of the Shasta Dam quadrangle (Diller's Bass Mountain diabase) apparently is younger than part of the Balaklala rhyolite and older than the Kennett formation.

BALAKLALA RHYOLITE

GENERAL DESCRIPTION

The Balaklala rhyolite is made up of many rhyolitic flows and beds of coarse and fine pyroclastic material that together form a broad elongate volcanic pile. Although the rocks of the Balaklala include many lithologic types that can be mapped separately, they are almost identical chemically and mineralogically. The mode of formation of the rocks and their reaction to the secondary processes to which they have been subjected account for many of the differences in appearance. Where the rhyolitic rocks are not weathered and are not deformed, they typically are hard light-gray or light-green felsitic rocks, many of which contain megascopically visible quartz and feldspar phenocrysts. The number and size of the phenocrysts vary in different flows, and this feature has been used to distinguish between flows that are otherwise identical. Amygdaloidal, flow-banded, flow-brecciated, and platy rhyolitic rocks are varietal types of flows. Pyroclastic rocks range in texture from coarse volcanic breccia or conglomerates through lapilli tuff to shaly tuff.

Foliation has been formed in the rhyolite by dynamic metamorphism. Sericite schist has been formed from rhyolite in some areas; sheeted rhyolite has formed in other areas. The rhyolite is altered in some places to dark-green or dark-gray rocks and in other places to shades of pink, lavender, green, or white by regional metamorphism or by hydrothermal alteration. The common hydrothermal alterations are pyritization, silicification, and sericitization. Much of the rhyolite is iron stained.

DISTRIBUTION AND RELATIONSHIP TO OTHER ROCKS

The Balaklala rhyolite crops out in a belt 16 miles long and 5 miles wide that strikes N. 15° E. from the south half of the Whiskytown quadrangle to the southwest quarter of the Behemotosh quadrangle. It forms the rugged hills west of the Sacramento River from South Fork Mountain and Iron Mountain in the Whiskytown quadrangle northeast to Mammoth Butte and Behemotosh Mountain in the Behemotosh Mountain quadrangle. The total thickness of the siliceous flows and pyroclastic rocks may be as much as 3,500 feet; it is thickest in the central part of the district and thins toward the periphery but varies greatly in thickness from place to place because of the difference in the amount of material that was extruded from many separate vents.

Small dikes, sills, and plugs of Balaklala rhyolite cut the underlying Copley greenstone throughout the mapped area. Also some thin, Balaklala rhyolite-type pyroclastic and flow rocks are in areas that are predominantly Copley greenstone (fig. 11). In the eastern part of the area, the eruption of lava of Copley greenstone continued while lava of Balaklala rhyolite was erupting farther west, and at some places the two lava types interfinger and are contemporaneous. Some of the bands of rhyolite in the Copley greenstone are erosion remnants in the troughs of synclines along the thin outermost edges of the volcanic pile, but others are rhyolitic material that was erupted locally before the

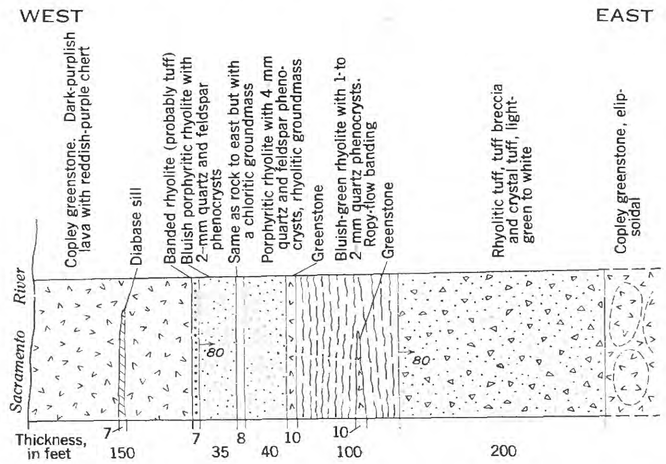

FIGURE 11.—Partial section in the Copley greenstone that contains interbedded rhyolitic flows and pyroclastic rocks, showing distribution of map units in the Balaklala rhyolite on the east flank of Behemotosh Mountain. Locality is 4,000 feet N. 15° W., from the Walker mine.

eruption of the flows of Copley greenstone had ceased and are interlayered with the Copley. Examples of rhyolitic flows that are infolded in the Copley are some of the rhyolite in the southwest quarter of the Whiskytown quadrangle, and some rhyolite along the Sacramento River in the southern part of the Shasta Dam quadrangle. Rhyolitic flows and pyroclastic rocks that are interlayered with the flows of Copley greenstone are generally in the upper part of the Copley and near the Copley-Balaklala contact; they can rarely be distinguished from rhyolitic sills in the Copley unless they are accompanied by rhyolitic pyroclastic material.

At many places along the Copley-Balaklala contact there is a transition zone. At these places, the upper part of the Copley contains bombs and fragments of rhyolite similar to Balaklala rhyolite, mixed with Copley greenstone pyroclastic material. At some localities the transition zone grades downward into mafic pillow lava and upward into rhyolitic pyroclastic rocks.

Minor flows of Copley greenstone-type lava and pyroclastic rocks are interlayered with the Balaklala rhyolite at some localities. These have been distinguished on the geologic map from the Copley greenstone because they are flows younger than part of the Balaklala; the same differentiation could not be made for early flows of rhyolite in the Copley because sills of Balaklala rhyolite in the Copley are common. It was difficult to determine whether some areas of greenstone in the Balaklala near the Stowell mine are windows of the underlying Copley, or whether they are younger greenstone flows in the Balaklala, but they are shown on the geologic map as greenstone flows in the Copley.

The Balaklala is conformably overlain by the Kennett formation. At many places—as along the east side of Backbone Creek between Upper Limestone Valley and Lower Limestone Valley; along the road to the Mammoth mine; east of the Golinsky mine; and in the vicinity of Butcher Creek—rhyolitic tuff and pyroclastic material grades upward into shale and sandstone of the Kennett without a stratigraphic break. The lower limit of the Kennett is placed where shale predominates over rhyolitic tuff.

STRATIGRAPHIC RELATIONSHIP

The Balaklala rhyolite is a complexly interlayered broad volcanic pile. Although the volcanic rocks were extruded from many vents, and sills and dikes of the same age and lithology intrude the flows, a recognizable stratigraphy is present in the pile, so that the group of volcanic rocks that were extruded early in the sequence can be mapped separately from those extruded later. Geologic mapping has shown that there was a progressively coarser crystallization of phenocrysts of quartz and feldspar in the magma chamber from which the flows were derived, and although there are many exceptions and reversals, the earliest and most widespread flows were nonporphyritic, later flows contained phenocrysts 1–4 millimeters in diameter, and the latest flow contained coarse phenocrysts more than 4 millimeters across. These types have been used as lithologic units on the maps, but it was found that the stratigraphic significance of the lithologic types must be interpreted with caution; later rhyolitic rocks intrude earlier flows as dikes and sills that are not always distinguishable; in areas where vents were quiescent for long periods some types of flows are absent; and lava from several vents interfinger.

About one-fourth of the Balaklala rhyolite is made up of fine- to coarse-rhyolitic pyroclastic material. Although pyroclastic beds are widely distributed throughout the Balaklala, they are more common in some parts of the stratigraphic section than in others. Concentrations of pyroclastic material are most common in the upper part of the lower unit of the Balaklala in the central part of the district, and in the upper part of the middle unit. Pyroclastic beds are also present locally at the base and at the top of the upper unit of coarse-phenocryst rhyolite and form transition zones to the rocks above and below. The pyroclastic rocks range in size from small bodies a few tens of feet in length to continuous layers that have been traced for several thousand feet. They are composed of shaly tuff, crystal tuff, lapilli tuff, fine and coarse volcanic breccia, flow breccia, and volcanic conglomerate. It is not always possible to determine whether a volcanic breccia was of pyroclastic origin or whether it is a flow breccia which was formed by the breakage and inclusion of fragments of crust into the liquid interior of a flow. Generally, if the fragments and matrix were of the same material, the breccia was regarded as a flow breccia.

Much of the pyroclastic material is water deposited, but some areas of pyroclastic material are roughly equidimensional in plan and may have been formed as explosion pipes of volcanic breccia. Few widespread tuff layers that would serve as horizon markers have been found in the Balaklala. Individual flows and pyroclastic beds are lenticular, and only a few can be traced for as much as 3,000 feet. The most notable exception is the cumulo dome of coarse-phenocryst rhyolite that forms the upper unit of the Balaklala rhyolite from the Stowell mine to the Golinsky mine—a distance of about 6 miles.

All the flows in the Balaklala rhyolite except the interlayered mafic flows are of uniform chemical and mineralogic composition, and where pyroclastic material was absent it was impossible to map individual flows except by texture. The phenocryst size is the main criterion used in mapping individual flows. In flows of porphyritic rhyolite in the West Shasta district, the quartz and feldspar phenocrysts maintain a nearly uniform maximum size, and a seriate texture is seldom present. Megascopically, the quartz phenocrysts are conspicuous, whereas the feldspar phenocrysts blend with the groundmass and are inconspicuous. The same criterion of phenocryst size was used by Gavelin (1939, p. 146) in mapping similar soda-rich rocks in the Malanäs district, Sweden. Flow-banding is common in some of the nonporphyritic rhyolite at the base of the Balaklala, and was an aid locally in mapping structure. It is uncommon in the porphyritic rhyolite.

Three texturally distinctive varieties of rhyolite make up the flow and pyroclastic rocks of the Balaklala: (1) nonporphyritic rhyolite, in which quartz grains are microscopic, except a few as much as 1 millimeter in diameter; (2) medium-phenocryst rhyolite, in which quartz phenocrysts range from 1 to 4 millimeters; and (3) coarse-phenocryst rhyolite, in which quartz phenocrysts are more than 4 millimeters. The nonporphyritic rhyolite is characteristic of the lower part of the volcanic sequence, the medium-phenocryst rhyolite of the middle part, and the coarse-phenocryst rhyolite is characteristic of the upper part. Each of the three distinctive varieties of rhyolite is somewhat irregularly distributed laterally, so that at many places, especially around the periphery of the volcanic pile, only one or two varieties are present. The variety most characteristic of a particular part of the sequence at places contains minor flows, pyroclastic layers, and dikes of the other kinds. Provided these deviations are kept in mind, it is convenient and useful to refer to lower, middle, and upper units of the Balaklala rhyolite, to emphasize the stratigraphic significance of the three distinctive varieties of rhyolite. Recognition of this general sequence has led to a practical interpretation of structure and favorable zones for ore in many of the mine areas; more widely it clarifies the regional structure and provides a basis for mineral exploration.

The lithologic description of the lower, middle, and upper units of the Balaklala rhyolite is given below.

Generalized stratigraphy of the Balaklala rhyolite

Unit	*Description*	*Feet*
Transition zone to Kennett formation.	Tuff containing some quartz phenocrysts more than 4 millimeters in diameter and rhyolitic tuff interbedded with thin shale beds.	0–300
Upper	Coarse-phenocryst rhyolite containing quartz phenocrysts more than 4 millimeters in diameter. Transition zone at the base locally is composed of coarse-phenocryst rhyolite tuff and breccia.	0–1, 400
Middle	Predominantly flows of medium-phenocryst rhyolite and pyroclastic material with quartz phenocrysts 1 to 4 millimeters in diameter but contains many small flows of nonporphyritic rhyolite. Widespread pyroclastic beds in the upper part of the middle unit. Contains a few small mafic flows of Copley greenstone-type lava.	0–1, 500
Lower	Predominantly nonporphyritic rhyolite and rhyolitic pyroclastic material but includes a few flows of medium-phenocryst rhyolite and a few mafic flows.	0–2, 000
Transition zone to Copley greenstone.	Mixed mafic and silicic pyroclastic rocks.	0–150

The early rhyolitic flows that form the lower unit of the Balaklala are more widespread and continuous than later flows. The lower unit consists mostly of nonporphyritic rhyolite, some of which is flow banded, but it also contains rhyolitic tuff and volcanic breccia. Flow-banded rhyolite (part of which may be tuff) and flow breccia are more common in the lower unit than in the middle, and shaly tuff is less common. Several varieties of pyroclastic breccias are in the lower unit; they are predominantly flow breccias, but some contain principally fragments of flow-banded rhyolite in an unbanded, tuffaceous matrix; others contain fragments of nonporphyritic rhyolite in a matrix of crystal tuff with quartz and albite crystals, and still others have fragments of porphyritic and nonporphyritic rhyolite together with fragments of greenstone in a matrix that may be either felsic or mafic. A few flows of medium-phenocryst rhyolite and pyroclastic material, and a few Copley greenstone-type mafic flows, are interlayered with the predominantly nonporphyritic rhyolite of the lower unit of the Balaklala. This unit is as much as 2,000 feet thick at some places, although at other places rocks of the middle unit rest on Copley greenstone where the flows of the lower unit were not deposited.

The middle unit of the Balaklala, the medium-phenocryst rhyolite, consists characteristically of porphyritic rhyolitic flows and pyroclastic material containing quartz and feldspar phenocrysts 1–4 millimeters in diameter. It contains many flows of nonporphyritic rhyolite in some localities, and at such places it is difficult to distinguish the lower unit of the Balaklala from the middle unit. The upper part of the middle unit contains abundant pyroclastic layers in the medium-phenocryst rhyolite, but flow breccias are rare. The pyroclastic layers in the lower and middle units are well exposed north of the Mammoth mine in the canyon of Little Backbone Creek, and in the North Fork of Squaw Creek arm of Shasta Lake. North of the Mammoth mine single beds of coarse pyroclastic rocks have been traced continuously for more than 3,000 feet, and in this locality the pyroclastic beds attain a thickness of 300 feet.

The upper part of the middle unit of the Balaklala is the ore zone, and the deposits occur as replacement bodies in medium-phenocryst rhyolitic flows that lie beneath pyroclastic beds.

The flows of the middle unit of the Balaklala did not extend as far laterally in some areas as those of the lower unit, but they are more extensive than those of the upper unit. The middle unit is generally 500 to 1,500 feet thick, but at a few places the upper unit rests directly on the lower unit.

The upper unit of the Balaklala is characteristically a single, continuous body of massive, uniform, coarse-phenocryst rhyolite, which contains many quartz and feldspar phenocrysts that are more than 4 millimeters in diameter. Most of the upper unit is structureless, without bedding or layering, and locally the rocks are poorly foliated. This unit is about 1,400 feet thick at Mammoth Butte, but thins abruptly toward the periphery. Although rare within the main body of the upper unit, a few beds of pyroclastic rocks do occur. These rocks, composed in part of coarse-phenocryst rhyolite, are common at the top and bottom of the upper unit and as extensions, in the same stratigraphic horizon, beyond the periphery of the massive, nonpyroclastic part of the upper unit. The coarse-phenocryst rhyolite forms a cumulo dome, and the massive structureless character of a large part of the rock tends to obscure the less prominent areas of associated coarse-phenocryst pyroclastic rocks that give evidence of its surface origin. The pyroclastic rocks are described in more detail than their relative abundance would justify because they are horizon markers and because they are critical features used in determining the origin of the main body of coarse-phenocryst rhyolite.

Only one area of pyroclastic rocks is recognized within the main body of the upper unit of the Balaklala rhyolite; this area is exposed on the southeast side of Mammoth Butte above Shoemaker Springs and above the Mammoth mine, where a layer of coarse pyroclastic material 70 feet thick contains fragments of coarse-phenocryst rhyolite mixed with fragments of nonporphyritic rhyolite in a tuffaceous matrix.

A thin bed of tuff or fine volcanic breccia that contains quartz phenocrysts more than 4 millimeters in diameter occurs at many places at the base of the coarse-phenocryst rhyolite of the upper unit. Tuff beds are also present above this rhyolite and are interbedded with thin beds of shale of the overlying Kennett formation. The pyroclastic zones in the uppermost and lowermost parts of the upper unit are widely distributed but are characteristically lenticular and discontinuous, although locally they may be traced, except for erosion gaps in the present topography, over an area of several square miles.

The coarse-phenocryst rhyolite is overlain by a fine volcanic breccia and sandy tuff that contains fragments of this rhyolite. The volcanic breccia is overlain in turn by interbedded crystal tuff, fine-grained tuff beds, tuffaceous shale, and thin beds of gray shale. The beds of tuff and fine pyroclastic material at the top of the coarse-phenocryst rhyolite are best exposed on the ridge east of the Mammoth mine, about 700 feet west of the prominent outcrop of limestone of the Kennett formation. This tuffaceous zone is also well exposed east of the Golinsky mine (fig. 12). The tuff beds grade upward into shale of the Kennett, and the contact between Balaklala and Kennett is placed where shale predominates over tuff.

One of the most continuous zones of pyroclastic material is the transition zone at the base of the upper unit of the Balaklala rhyolite, directly under the coarse-phenocryst rhyolite. This zone is commonly made up of one or more beds of tuff and is 10 to 50 feet thick,

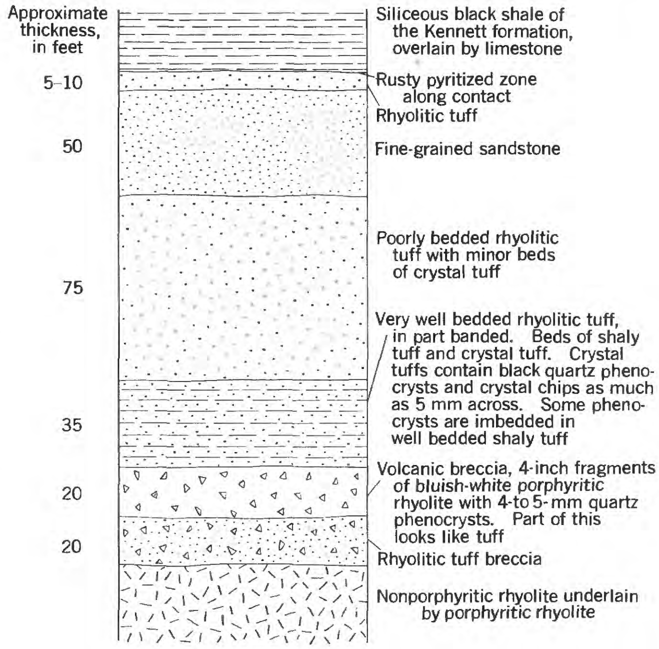

FIGURE 12.—Vertical section across the contact between the Balaklala rhyolite and shale of the Kennett formation east of the Golinsky mine.

but, where it is composed mostly of coarse pyroclastic material, it is as much as 150 feet thick. Even this pyroclastic zone, it must be emphasized, is made up of several discontinuous pyroclastic beds, and in many places the transition zone is absent. Beds of tuff also overlie the dome of coarse-phenocryst rhyolite and one coarse pyroclastic bed is interbedded with the coarse-phenocryst rhyolite west of the Mammoth mine. The pyroclastic zone that underlies the coarse-phenocryst rhyolite consists mostly of light-colored, rhyolitic tuff or fine pyroclastic material, although a coarse pyroclastic layer occurs 500 feet south of the Copper Crest ore body of the Mammoth mine. The pyroclastic zone under the coarse-phenocryst rhyolite is exposed at many places in the Mammoth mine area; it is also exposed under the coarse-phenocryst rhyolite north and northeast of the Uncle Sam mine, at an altitude of about 2,750 feet, and at the surface over much of the topographic flat between the Keystone and Stowell mines.

The coarse-phenocryst rhyolite dome is not as extensive laterally as the tuff beds above and below it, and where the coarse-phenocryst rhyolite wedges out, the overlying and underlying tuff beds merge and continue beyond the limits of the dome.

The composition and texture of the pyroclastic beds in the transition zone at the base of the upper unit of the Balaklala rhyolite are not uniform. The transition zone consists predominantly of thin-bedded rhyolitic

FIGURE 13.—Coarse-phenocryst Balaklala rhyolite containing dark quartz phenocrysts, many of the dark phenocrysts are angular clasts of quartz, and the rock may be a crystal tuff.

tuff interbedded with tuff containing quartz crystals and chips that are more than 4 millimeters in diameter. A coarse-phenocryst rhyolite containing dark-gray to smoky quartz phenocrysts in a greenish-gray matrix is associated at many places with the tuff layer at the base of the coarse-phenocryst rhyolitic dome (fig. 13). In most places the rhyolite containing smoky quartz phenocrysts is a tuff, but in other places the matrix is too altered to be certain. Lithic tuff containing fragments of coarse-phenocryst rhyolite is fairly common, and at some places the tuff grades into volcanic breccia and fragments of nonporphyritic and porphyritic rhyolite one-half to 2 inches in diameter in a matrix of fine tuff or crystal tuff. Some of the fragments of rhyolite in this volcanic breccia are well rounded and appear to be waterworn, and part of the bed is a volcanic conglomerate. The transition zone at many places grades downward into the tuff and flows of the middle unit of the Balaklala.

The center of the cumulo dome of the coarse-phenocryst rhyolite is probably about 1 mile southwest of the Mammoth mine. What appear to have been the main feeders for this rhyolite are exposed near the Uncle Sam gold mine. There the tuff of the upper unit of the Balaklala at the base of the dome of coarse-phenocryst rhyolite, and the underlying flows and pyroclastic rocks of the middle unit have gentle dips. They are cut by plugs of coarse-phenocryst rhyolite that have steep contacts. These plugs, which were the feeders for the dome, cut the coarse-phenocryst rhyolitic tuff at the base of the dome although at this locality they cannot be traced into the main dome of coarse-phenocryst rhyolite because of an erosion gap in the present topography.

PETROGRAPHIC DESCRIPTION

The silicic flows and pyroclastic beds of the Balaklala rhyolite are leucocratic rocks that have a very high soda content. The composition of the Balaklala is shown in table 3. The rhyolite contains mainly albite and quartz and some epidote, orthoclase, chlorite, sericite, clay minerals, and magnetite. Some glass was present, but it is now devitrified.

TABLE 3.—*Chemical analyses of extrusive and intrusive phases of Balaklala rhyolite*

1. Flow of medium-phenocryst rhyolite from Igo quadrangle. Analyst, M. K. Carron.
2. Flow of coarse-phenocryst rhyolite near Balaklala mine. Analyst, W. J. Blake, Jr.
3. Weathered coarse-phenocryst rhyolite from Balaklala mine. Analyst, W. J. Blake, Jr.
4. Dike of coarse-phenocryst rhyolite near north portal of Spread Eagle mine.[1]
5. Plug of coarse-phenocryst rhyolite near Clipper mine.[2]
6. Rhyolite from Iron Mountain mine.[3]

Constituent	Chemical analyses					
	1	2	3	4	5	6
SiO_2	77.01	73.60	68.71	68.75	76.47	74.52
Al_2O_3	12.04	13.08	16.61	16.75	13.90	--------
Fe_2O_3	.67	1.16	2.79	.48	.18	--------
FeO	1.51	2.69	1.43	1.72	.40	--------
CaO	.86	.22	.20	.89	.19	--------
MgO	1.29	1.32	1.65	.83	1.74	--------
K_2O	.59	1.65	1.17	.80	.65	.40
Na_2O	4.58	3.93	.94	6.95	4.76	5.16
H_2O^-	.13	.13	.69	.84	.29	--------
H_2O^+	1.19	1.73	5.30	1.52	1.34	--------
TiO_2	.18	.27	.36	.27	--------	--------
MnO	.03	.05	.04	.04	--------	--------
CO_2	--------	.12	.01	None	None	--------
P_2O_5	--------	.07	.03	.16	--------	--------
ZrO_2	--------	--------	--------	None	--------	--------
SO_2	--------	--------	--------	None	--------	--------
S	--------	--------	--------	None	--------	--------
BaO	--------	--------	--------	.03	--------	--------
SrO	--------	--------	--------	.03	--------	--------
	100.08	100.02	99.93	100.06	99.92	--------

[1] Analysis from Butler, B. S., 1909, Pyrogenetic epidote: Am. Jour. Sci., 4th ser., v. 28.
[2] Analysis from notes of Graton, L. C., and Butler, G. S., 1906, U. S. Geol. Survey.
[3] Analysis from notes of Diller, J. S., 1901–04, U. S. Geol. Survey.

NONPORPHYRITIC RHYOLITE

The nonporphyritic rhyolite is a dense felsitic rock without megascopic phenocrysts or with only sparsely disseminated quartz and feldspar phenocrysts as much as 1 millimeter in diameter in an aphanitic groundmass. The color ranges from white to light bluish green and buff, and more rarely to red, greenish black, and black. At the surface the nonporphyritic rhyolite is generally hard and siliceous looking, but locally is soft and altered. Some of the nonporphyritic rhyolite is banded, as shown by the delicate flow lines that are brought out by weathering, by color differences, or by stretched vesicles that are filled with quartz and epidote. Contorted flow banding is common, but in some localities the flow structures are regular, and in the field it is difficult to distinguish whether the banding was due to flow, to schistosity, or to bedding in a fine-grained tuff. Much of the banded rhyolite proved to be tuff when examined under the microscope. A dark-greenish-black aphanitic rock that was mapped as black felsite is exposed in the lower part of the middle unit of the Balaklala in the northern part of the district. The dark felsite is in part dacite and in part dacitic tuff. It has sharp contacts with the lighter colored soda rhyolitic flows and pyroclastic rocks.

The rhyolite is called nonporphyritic if it contains no phenocrysts or only sparsely disseminated quartz and albite phenocrysts less than 1 millimeter in diameter. The quartz phenocrysts are subhedral and are shattered by closely spaced fractures. The albite phenocrysts in the nonporphyritic rhyolite are lath shaped, while those in the medium- and coarse-phenocryst rhyolite are more nearly equant in outline. Carlsbad twinning is predominant; albite twinning is much rarer than in the albite phenocrysts in the porphyritic rhyolite.

The groundmass is a fine-grained aggregate of albite and quartz that contains some chlorite, epidote, clay minerals, sericite, and hydromica, and small quantities of apatite, magnetite, biotite, and sphene. The groundmass texture may be pilotaxitic, trachytic, or microgranitoid, and fluidal, microspherulitic, and amygdaloidal structures are common.

Where unaltered, the groundmass consists mainly of a felted mass of albite laths averaging about 0.1 millimeter in length, and interstitial quartz. Albite ranges in composition from $Ab_{95}An_5$ to $Ab_{92}An_8$. In some specimens the albite laths are alined and flow around phenocrysts, forming a trachytic texture. A few thin veinlets containing albite and quartz cut the rhyolite. Mafic minerals are secondary green biotite, chlorite, and epidote, which occur in fractures in the rock and as vesicle fillings. All gradations exist from unaltered nonporphyritic rhyolite consisting mainly of quartz and albite to sheared and altered rhyolite composed of quartz, chlorite, sericite, hydromica, clay minerals, epidote, and green biotite.

Much of the nonporphyritic rhyolite consists of a granular aggregate of quartz, sericite, chlorite, epidote, clay minerals, and magnetite, and has a uniform grain size of 0.01 to 0.02 millimeter. Some of the nonporphyritic rhyolite shows excellent fluidal structures in ordinary light. Globulites of magnetite are drawn out to form sinuous bands. These structures are not evident under crossed nicols, which brings out a microgranular texture. This texture probably indicates a devitrified glass, as the microgranular texture does not reflect the fluidal structures.

Microspherulitic structures are common in the nonporphyritic rhyolite but are rare in the porphyritic rhyolite, which has a slightly coarser grained groundmass. These microspherulites are most common in the rhyolite that has a pilotaxitic or trachytic texture. The spherulites are 0.1 to 0.3 millimeter in diameter and have a fibrous, radial growth. In some, the fibers have a positive relief and consist entirely of quartz; others have a strong negative relief and consist of albite and possibly cristobalite. Most of the spherulites have a random pattern, although some are arranged in chains. Some of the nonporphyritic rhyolite has many rounded forms of quartz about 0.1 to 0.3

millimeter in diameter. They have the same shape as the microspherulites, but most of them have no internal structure. They probably are spherulites that have been replaced by quartz.

Amygdaloidal structures are present in the nonporphyritic rhyolite, but are absent in the porphyritic rhyolite. The vesicles are lenticular and are as much as 1 centimeter long and 3 millimeters wide. They are filled with quartz, epidote, and chlorite.

A dark-greenish-black felsitic rock that occurs locally in the middle and lower units of the Balaklala rhyolite is in part dacite and in part dacitic tuff. Some layers of this rock are uniformly dark throughout, as observed in the beds near the Shasta King mine, whereas others are mottled or have irregular dark seams through a lighter green matrix. The dacite forms separate flows or pyroclastic layers interbedded with the lighter colored rhyolite and has sharp contacts with the rhyolite. It forms a very small part of the lower unit of the Balaklala near the Shasta King mine. The dacitic flows have a trachytic texture and contain plagioclase, epidote, quartz, and chlorite. The dacite contains at least 20 percent epidote, which gives the rock its dark color. Epidote is in part disseminated through the rock, and in part fills vesicles and fractures. The plagioclase is unzoned oligoclase ($Ab_{80}An_{20}$) and is much more calcic than in other flows in the Balaklala. Although no primary mafic minerals remain, the large percentage of epidote suggests that mafic minerals were present and that they were altered to epidote during regional metamorphism. The epidote is later than the quartz veinlets that cut the rock, as shown by replacement of originally continuous quartz veinlets by epidote. Much of the dark felsite has a clastic texture and is a fine-grained dacitic tuff. No gradation between dacite and rhyolite is evident—either in the field or from the petrographic study.

Diamond drilling at Iron Mountain revealed a mottled siliceous, chloritized nonporphyritic rhyolite. The rock is composed mainly of quartz and chlorite and contains some hydromica, sericite, calcite, epidote, pyrite, and chalcopyrite. The chlorite is prochlorite and has the following optical properties: optic sign positive, $n_\alpha = 1.608$, and $n_\gamma = 1.615$. The rhyolite has been sheared and brecciated, and the original texture has been destroyed except for some round, radial quartz growths that are probably silicified spherulites. Locally the mottled, dark-green, chloritic rock looks like Copley greenstone. Some of the chloritic rock has a few quartz phenocrysts and is a chloritized rhyolite, but one 10-foot layer, which contains finely disseminated leucoxene and no quartz may be a thin flow of andesite interlayered with the Balaklala rhyolite.

MEDIUM-PHENOCRYST RHYOLITE

Medium-phenocryst rhyolite is a porphyritic rock that has quartz and feldspar phenocrysts 1 to 4 millimeters in diameter in an aphanitic groundmass. The color may be white, buff, light cream, gray, light greenish gray, or pink. This rhyolite ranges in structure from hard unsheared rhyolite to a strongly foliated rock that contains quartz phenocrysts in a groundmass of secondary minerals. The porphyritic rhyolite generally consists of about 10–20 percent quartz and feldspar phenocrysts in a felsitic groundmass. Quartz phenocrysts are conspicuous and constitute from 5 to 20 percent of the rock. The quartz commonly occurs in the form of glassy, euhedral stubby dipyramids. Usually one rhombohedron predominates and the prism faces are absent; this combination gives a pseudocubical form to many of the phenocrysts. Some of the quartz phenocrysts are rounded or anhedral. Glomerocrysts, that is, clusters of 2-millimeter quartz phenocrysts aggregating 6 to 12 millimeters in diameter, are common, and they caused some trouble in mapping on the basis of phenocryst size. However, the size of individual phenocrysts in the glomerocryst can usually be recognized.

Feldspar phenocrysts are as abundant as quartz phenocrysts in the unaltered rock, but in much of the porphyritic rhyolite the feldspar crystals are completely altered to sericite and clay minerals. They are euhedral and average about the same size as the quartz phenocrysts.

No mafic minerals are recognizable megascopically in most of the rhyolite, but locally small amounts of epidote were observed.

Platy, medium-phenocryst rhyolite makes up much of the Balaklala rhyolite roof pendant that projects into the Mule Mountain stock along Clear Creek, east and south of Whiskytown, and occurs also on Iron Mountain peak. The rock is a light-greenish-gray medium-phenocryst rhyolite that has well-formed parting planes one-fourth to 1 inch apart. Although the parting planes appear to be parallel when viewed from a distance, in detail the plates are lenticular and few of them can be traced as much as 10 feet (fig. 14). The porphyritic rhyolite between plates is massive and unsheared. The platy structure probably is a primary flow structure that differs in origin from the sheeted rhyolite and from rhyolite having closely spaced fracture cleavages.

Under the microscope the medium-phenocryst rhyolite is seen to be a porphyritic rock consisting of 1- to 4-millimeter quartz and feldspar phenocrysts in an extremely fine grained holocrystalline groundmass of pilotaxitic or microgranitoid texture. Spherulitic

FIGURE 14.—Platy medium-phenocryst rhyolite near Mount Shasta mine east of Clear Creek. Note the lenticular character of the plates.

structures and granophyric textures are rarely found. No seriate texture was observed; the change from phenocrysts to fine-grained groundmass is extremely abrupt. The rock consists mainly of albite and quartz, and small amounts of sericite, hydromica, epidote, chlorite, clay minerals, probable orthoclase, zoisite, biotite, carbonate, magnetite and sphene listed in decreasing order of abundance. The unaltered rock contains more than 90 percent albite and quartz. Generally the albite in the groundmass is partly altered to sericite and hydromica, and to smaller amounts of kaolinite, chlorite, epidote, and zoisite. Where shearing is pronounced, only quartz phenocrysts remain in a completely secondary groundmass.

PHENOCRYSTS

Quartz.—Quartz phenocrysts constitute from 5 to 20 percent of the rock; quartz also constitutes about one-third of the groundmass, and is present in veinlets with or without albite. The quartz phenocrysts range in outline from rounded to stubby dipyramids. They are colorless or are red due to thin coatings of hematite around and in cracks through them. Liquid inclusions are abundant, but are lacking in the quartz of the groundmass and veinlets. Most of the quartz phenocrysts have straight, sharp faces against the groundmass, but a few are extremely corroded.

The quartz phenocrysts in both sheared and unsheared porphyritic rhyolite are typically shattered by closely spaced, uniformly distributed fractures. The shattering probably was caused by shrinking in conversion from beta to alpha quartz during cooling (Wright and Larsen, 1909, p. 438).

Albite.—Albite occurs as euhedral phenocrysts in about the same number as quartz phenocrysts, but is also found as a felted mass of tiny laths in the groundmass, as intergrowths with quartz, and in quartz-albite veinlets cutting the rock. Many of the albite phenocrysts have indices of refraction less than 1.54 and have the maximum extinction angle perpendicular to (010), X′ to (010) of 15° to 16°, showing that the albite is more sodic than $Ab_{95}An_5$; the phenocrysts range in composition from $Ab_{96}An_4$ to $Ab_{92}An_8$. They are not zoned, and have both albite and Carlsbad twinning. Zoisite and epidote are found in a few phenocrysts, but many are unaltered and contain no inclusions.

Epidote.—Epidote, in amounts usually less than 5 percent, is disseminated as tiny grains throughout the groundmass, and occurs locally in small amounts in altered albite, and as phenocrysts that look pseudomorphic after pyroxene.

Orthoclase.—A feldspar, which is probably orthoclase, is irregularly distributed in the groundmass in a few thin sections; it has a negative optic sign, negative relief, and an optic angle of about 80°.

GROUNDMASS

The holocrystalline groundmass of the unaltered porphyritic rhyolite consists predominantly of albite and quartz, but contains small quantities of orthoclase(?), epidote, and magnetite. The albite in the groundmass has the same composition as that in the phenocrysts. Groundmass textures are pilotaxitic, microgranitoid, myrmekitic, and granophyric. The most characteristic groundmass has a pilotaxitic texture and consists of a felted mass of unoriented plagioclase laths and interstitial quartz, small amounts of orthoclase (?), epidote, and magnetite. Pilotaxitic texture is dominant in the nonporphyritic rhyolite, which has the finest grained groundmass; microgranitoid texture is more common in the medium-phenocryst rhyolite, which has a more coarsely grained groundmass. Myrmekitic and granophyric textures are less common and probably indicate reworking of the original rock by deuteric solutions.

Some of the porphyritic rhyolite with the finest grained groundmass contains spherulitic structures. These are colorless microscopic spheres that show a dark cross under crossed nicols. The spherulites consist of a radial fibrous growth that has positive relief and is probably quartz.

Veinlets of quartz and albite cut the rock; the composition of the albite in these veinlets is $Ab_{96}An_4$. The veinlets have sharp walls and usually cause no change in the texture or composition of the adjacent groundmass, but in a few sections a recrystallization of the groundmass to a micrographic intergrowth was observed.

Micrographic and myrmekitic textures are common. Abundant myrmekitic intergrowths of albite and quartz surround quartz phenocrysts or replace the rhyolite groundmass. Most of the quartz phenocrysts have megascopically straight crystal faces, but thin sections show that they are extremely irregular. The quartz in the myrmekitic intergrowth surrounding the quartz phenocrysts is in optical continuity with the phenocryst, and grows out from it. Some quartz phenocrysts have cores of myrmekitic intergrowths of quartz and albite (fig. 15).

Micrographic intergrowths are present mainly in albite phenocrysts and less commonly surround them. These intergrowths are much coarser than the myrmekite and seem to be a replacement by quartz along cleavage planes in albite.

COARSE-PHENOCRYST RHYOLITE

The coarse-phenocryst rhyolite is similar in appearance to the medium-phenocryst rhyolite except that the quartz and albite phenocrysts are more than 4 millimeters in diameter. Where unweathered most of the rhyolite is a hard massive leucocratic rock that contains 20 to 30 percent quartz and albite phenocrysts in an aphanitic matrix; mafic minerals constitute less than 5 percent. This leucocratic rock is white, cream, light greenish gray, or buff. Over much of the district it has a soft "punky" appearance near the surface due to alteration of the feldspar to clay minerals.

Both extrusive and intrusive coarse-phenocryst rhyolite is present. The main extrusive mass is at Mammoth Butte, where it forms a cumulo dome that has an exposed thickness of about 1,400 feet. At least 90 percent of this dome consists of homogeneous, leucocratic, porphyritic rhyolite containing subhedral to euhedral quartz phenocrysts as much as 1 centimeter in diameter, and euhedral albite phenocrysts as much as 6 millimeters, in a white, cream, or light greenish-gray aphanitic groundmass.

Intrusive coarse-phenocryst rhyolite forms plugs, sills, and dikes. At the Uncle Sam mine in the central part of the mining district, plugs consisting of this rhyolite constitute the main conduit for the extrusive coarse-phenocryst rhyolite dome. Other plugs, sills,

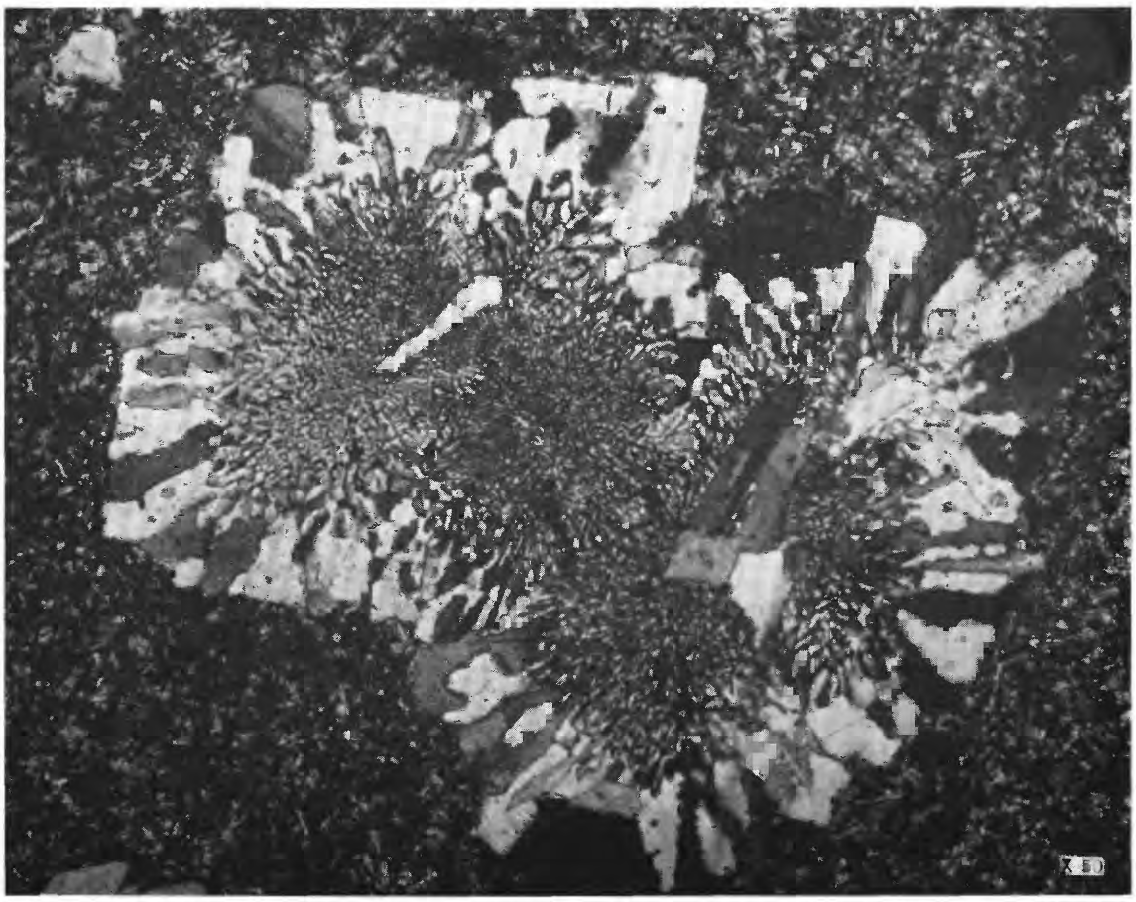

FIGURE 15.—Myrmekitic intergrowth of quartz and albite in medium-phenocryst Balaklala rhyolite. Crossed nicols, × 50.

and dikes of this rhyolite occur in the lower and middle units of the Balaklala rhyolite and in the Copley greenstone.

EXTRUSIVE COARSE-PHENOCRYST RHYOLITE

The extrusive coarse-phenocryst rhyolite is a porphyritic rock that contains mainly quartz and albite, but also contains biotite, epidote, zoisite, chlorite, sericite, hydromica, clay minerals, and magnetite. Quartz and albite phenocrysts constitute 20 to 30 percent of the rock in about equal amounts. The groundmass has a micrograniitoid texture. No flow-banding or spherulitic structures are present such as are common in the nonporphyritic rhyolite.

Quartz phenocrysts.—Quartz phenocrysts generally are as stubby dipyramids that range from 2 to 10 millimeters in diameter, but some are irregular or rounded. Most of them have a milky color; however, some are dark gray or smoky (fig. 13). The dark color is due to dark, minute dustlike inclusions near or along arcuate fractures in the quartz (fig. 16).

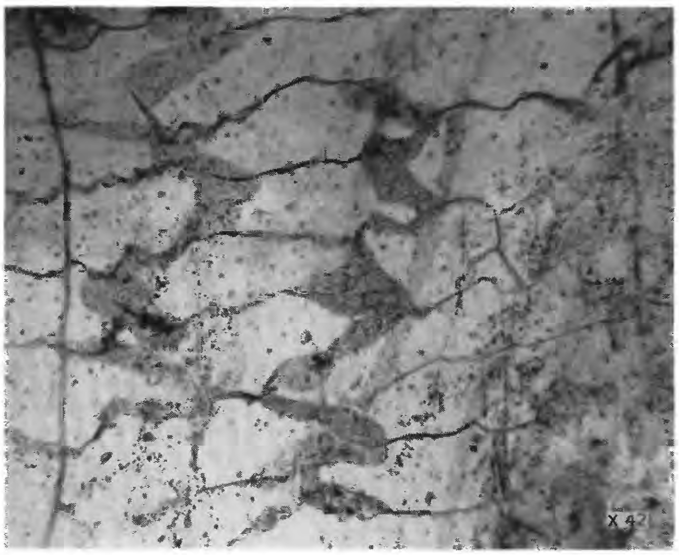

FIGURE 16.—Photomicrograph of dark quartz phenocryst in coarse-phenocryst rhyolite. Dark color is due to minute dark inclusions along arcuate fractures. Crossed nicols, × 42.

Most of the quartz phenocrysts in the coarse-phenocryst rhyolite are shattered by minute, closely spaced fractures similar to those in the medium-phenocryst rhyolite. This shattering is probably due to inversion from beta to alpha quartz. Megascopically the quartz phenocrysts seem to have sharp, straight faces. Under the microscope some crystals are seen to be very irregular in outline and have thin growth rims. The growth rims and the original quartz phenocryst both contain liquid inclusions, but the growth rims are not crackled. Quartz grains in the groundmass next to a quartz phenocryst are commonly in optical continuity with the phenocryst. Many quartz phenocrysts show strain shadows. The shattering that caused the strain shadows formed more widely spaced and irregularly distributed fractures than the crackling believed to have formed during inversion of the quartz. The irregular fractures in the quartz causing the strain shadows were probably formed during orogeny.

Albite phenocrysts.—Albite phenocrysts, ranging from 3 to 6 millimeters in diameter, constitute 15 to 25 percent of the rock. They are most commonly euhedral but some are irregular in outline. The composition of the albite ranges from $Ab_{97}An_3$ to $Ab_{92}An_8$. The phenocrysts are unzoned and have albite and Carlsbad twinning.

All the albite phenocrysts are altered to some extent and are replaced by sericite, hydromica, chlorite, clay minerals, and epidote. Chlorite and hydromica are the most abundant alteration products; epidote is rarely present in the albite.

Epidote.—Epidote constitutes as much as 5 percent of the coarse-phenocryst rhyolite. It is in grains less than 0.2 millimeter across that are disseminated throughout the groundmass, or is in clusters of tiny granules that may aggregate 4 millimeters across. Some of the epidote is pseudomorphic after hornblende.

Groundmass.—The groundmass has an equigranular texture and is composed mainly of quartz and albite, but also contains chlorite, biotite, epidote, sericite, hydromica, clay minerals, and magnetite. The grain size averages about 0.2 millimeter. No flow or spherulitic structures are present. The groundmass of the extrusive coarse-phenocryst rhyolite is coarser grained than the groundmass of the smaller phenocryst rhyolite and is equigranular instead of pilotaxitic. Quartz and albite are the principal constituents, but chlorite, in irregular patches as much as 1½ millimeters in diameter, is sparsely disseminated through the matrix and along fractures. Some of the chlorite has an anomalous blue interference color. Green biotite is present but includes chlorite as a secondary mineral along fractures and in irregular clusters. Both minerals generally constitute less than 5 percent of the rock.

Myrmekitic and micrographic intergrowths of quartz and albite are present similar to those in the medium-phenocryst rhyolite. Micrographic intergrowths are more abundant and are formed in isolated albite phenocrysts as centripetal replacement by quartz along cleavage planes in albite. In some instances the whole phenocryst may be a micrographic intergrowth, but generally an unreplaced core of albite remains. Myrmekitic intergrowths are in the center of a few quartz phenocrysts and in the groundmass surrounding quartz phenocrysts, and have a finer texture than the micrographic intergrowths.

There is no pronounced foliation in the coarse-phenocryst rhyolite. The secondary minerals are unoriented in contrast to the medium-phenocryst rhyolite, which may have oriented sericite and hydromica. However, near the Mammoth mine the rocks are poorly foliated near the thin edge of the coarse-phenocryst rhyolite dome.

INTRUSIVE COARSE-PHENOCRYST RHYOLITE

The intrusive coarse-phenocryst rhyolite is commonly a white to greenish-white porphyritic rock containing quartz and albite phenocrysts as much as 7 millimeters in diameter in a siliceous aphanitic groundmass. Plugs of coarse-phenocryst rhyolite cut the lower unit of the Balaklala rhyolite at the Uncle Sam gold mine in the central part of the mining district, and these plugs constitute the main conduit for the extrusive coarse-phenocryst rhyolite dome. Most of the plugs consist of white hard porphyritic rhyolite that is extremely silicified; parts of the rock contain more than 50 percent quartz. No mafic minerals are visible megascopically.

The intrusion of the plugs must have been accompanied by explosive activity, as the plugs and the surrounding nonporphyritic rhyolite, for several hundred feet, are shattered. This shattering permitted the rock to be more thoroughly altered by hydrothermal solutions than the unshattered rhyolite. Little spots and irregular masses of talc as much as 4 inches long are disseminated through much of the altered rock; in some places the rhyolite was altered to a quartz-talc rock. Silicification of the coarse-phenocryst rhyolite plugs and the nonporphyritic rhyolite near the plugs is much more widespread than the talc alteration. The quartz causes a wholesale replacement of the originally porphyritic rhyolite; this replacement decreases the size of the individual phenocrysts and increases the grain size of the groundmass, giving the rhyolite a granitoid appearance. Albite phenocrysts are extremely corroded, and albite occurs as irregular relict grains in secondary quartz.

The quartz and quartz-talc alteration is not limited to the plugs, but extends out into the shattered wall rock of nonporphyritic rhyolite. Where the alteration is intense, it is impossible to distinguish whether the rock was part of the coarse-phenocryst rhyolite plug or part of the surrounding nonporphyritic rhyolite.

A plug of coarse-phenocryst chloritic rhyolite exposed in the adit of the Uncle Sam gold mine, 1,800 feet from the portal, has a different appearance and composition from the other plugs in the Uncle Sam mine area. It is composed of a hard greenish-gray rhyolite that has no distinct porphyritic texture, but by close examination some glassy, anhedral quartz phenocrysts can be outlined in a green secondary groundmass. The nonporphyritic rhyolite wall rock of this plug is altered to a predominantly quartz-chlorite rock. Under the microscope the greenish-gray porphyritic rhyolite is seen to be composed of quartz and albite phenocrysts in a groundmass of quartz, albite, epidote, hornblende, chlorite, sericite, talc, clay minerals, sphene, and magnetite. Much of the rock is replaced by quartz and a few sulfide minerals. Quartz and albite occur as anhedral to subhedral phenocrysts as much as 5 millimeters in diameter in a groundmass averaging 0.5 millimeter in grain size. Albite is clouded by inclusions of epidote and by veinlets of chlorite and sericite that have cut through and replaced it. Euhedral green hornblende grains 0.2-millimeter across, which appear to be primary, make up as much as 2 percent of the rock; the hornblende is in part replaced pseudomorphously by epidote. The darker color of the rock is due to the presence of 15 to 20 percent chlorite that has replaced albite. The plagioclase phenocrysts are unzoned crystals that have albite twinning. They have an extinction angle of 15° in sections perpendicular to (010), X' to (010), which corresponds to albite of composition $Ab_{95}An_5$. The difference in mineralogy of the darker colored plug in the adit of the Uncle Sam gold mine is not understood. The talc alteration of nearby plugs in this area indicates the addition of magnesia-rich solutions. Possibly primary hornblende in the darker colored plug favored the formation of chlorite and epidote by the magnesia-rich hydrothermal solutions rather than talc, as in the other plugs in the vicinity of the Uncle Sam gold mine.

PYROCLASTIC BEDS

The rhyolitic pyroclastic beds consist of shaly tuff, crystal tuff, lapilli tuff, fine and coarse volcanic breccia, flow breccia, and volcanic conglomerate. In most of the pyroclastic beds the fragments are angular to subrounded and consist of coarse material, but in a few beds the fragments are well rounded and resemble volcanic conglomerate. The character of the material in a single pyroclastic bed is rarely uniform; coarse unstratified rock is commonly interlayered with finer pyroclastic rocks and with well-bedded tuff. Figure 17 shows the nature of a pyroclastic bed located 2,800 feet northwest of the Mammoth mine in the middle unit of the Balaklala rhyolite. The base is a jumble of coarse unstratified material that has fragments of porphyritic and nonporphyritic rhyolite. Some of these fragments are flow banded. This unstratified material is overlain by a poorly bedded fine pyroclastic

FIGURE 17.—Pyroclastic layer in middle unit of the Balaklala rhyolite northwest of the Mammoth mine in the canyon of Little Backbone Creek. Coarse-porphyritic and nonporphyritic rhyolite fragments as much as a foot in diameter below, grading to rhyolitic tuff and shaly tuff above.

breccia, which is in turn overlain by a well-bedded rhyolitic tuff and shaly tuff. Unsorted, angular fragments are characteristic of many of the volcanic breccias.

The beds range from fine, well-bedded, apparently water-deposited, shaly layers of tuff, through tuff containing quartz and albite crystals or crystal fragments, to lithic tuff containing fragments of porphyritic and nonporphyritic rhyolite as much as a quarter of an inch in diameter. Most of the fine tuff layers are well bedded; some are apparently reworked and contain waterworn fragments. In most of the crystal and lithic tuff layers the bedding is not apparent or is poorly developed. Most of the beds are only a few inches to a few feet thick and have a lateral extent of a few hundreds of feet. They may either lens out or grade into volcanic breccia. A group of tuff beds, such as a deposit in a basin, is much more extensive and in some places can be traced for several thousand feet. The larger units of tuff are composed of several individual beds; they commonly grade downward into coarser pyroclastic layers, although there may be no uniform sequence of rhyolite flows, tuff, and breccia.

Minor amounts of fine-grained dacitic tuff beds are present in both the lower and middle units of the Balaklala. They are exposed along the trail to the Shasta King mine on the South Fork of Squaw Creek. The dacitic tuff beds are dark-greenish-black, aphanitic rocks that have a cherty appearance. In the field they were mapped as black felsite, but under the microscope a clastic texture was recognized in some of the dacitic rock.

Tuff beds containing quartz and feldspar crystals and chips more than 4 millimeters in diameter are characteristic of the pyroclastic zone at the base of the coarse-phenocryst rhyolite. The tuffaceous origin of some of the crystal tuffs is difficult to recognize. Quartz crystals, more than 4 millimeters across are identical to the quartz phenocrysts in the coarse-phenocryst rhyolite in an apparently aphanitic matrix. These crystal tuff beds, though, have a heterogeneous character in contrast to the uniform character of the coarse-phenocryst rhyolite. Quartz crystals in the tuff have a very uneven distribution, and the crystal tuff beds grade laterally and vertically into rhyolitic tuff beds that have no quartz crystals. Also, the quartz crystals in the tuff are more rounded than those in the coarse-phenocryst rhyolite, and chips of phenocrysts are common. Another characteristic of the crystal tuff is a much higher ratio of quartz to feldspar phenocrysts than is present in the porphyritic rhyolite; the ratio of quartz to feldspar phenocrysts in the porphyritic rhyolite is about 1:1.

TUFF BEDS

Under the microscope the rhyolite tuff is seen to have the same mineralogy as the flow rocks of the Balaklala rhyolite, although the proportions of minerals are usually different. They contain grains of quartz and albite and lithic fragments of porphyritic and nonporphyritic rhyolite in a fine-grained clastic matrix of quartz, sericite, hydromica, chlorite, albite, kaolinite, montmorillonite and less commonly rutile, magnetite, biotite, and zeolites. Albite is usually much more altered than in the rhyolite flows, and is often completely altered to sericite, hydromica, chlorite, and clay minerals. Albite when unaltered has the composition $Ab_{96}An_4$ to $Ab_{92}An_8$, similar to that in flows in the Balaklala rhyolite.

Many of the tuff layers in the Balaklala are easily recognizable as they are well bedded or contain lithic fragments. However many of the fine-grained tuff and crystal-tuff layers have no apparent bedding, and megascopically they have an aphanitic matrix; these are difficult to recognize in the field as tuff because their clastic nature is not apparent. The rhyolite tuff beds have the following characteristics by which they may be recognized where bedding and clastic textures are not obvious in an outcrop:

1. In thin sections, minerals in the matrix of the tuff beds have a rounded, clastic texture. This is much more conspicuous under the microscope in plane light than under crossed nicols.
2. Angular chips of phenocrysts are common.
3. Quartz and feldspar fragments and crystals have a heterogeneous distribution, whereas the distribution of quartz and feldspar phenocrysts in the porphyritic rhyolites is much more uniform.

4. The matrix of the tuff is heterogeneous. Parts may be very fine grained and others rather sandy.
5. Quartz fragments and crystals greatly predominate over feldspar fragments and crystals.
6. Quartz and feldspar phenocrysts tend to be more anhedral than those in the rhyolite flows, and they commonly are rounded.
7. Porphyritic or nonporphyritic rhyolite fragments may be present, but they are rarely abundant or conspicuous in the fine-grained tuff or crystal tuff beds.

These criteria were developed by the microscopic study of fine-grained rocks that were obviously bedded and of clastic origin. Points 2, 3, 5, and 6 are criteria in recognizing crystal tuff in the field. Megascopic grains of quartz and feldspar commonly have rounded outlines, although euhedral crystals and crystal chips are present. The distribution of the megascopically visible quartz and feldspar grains is not uniform, and in places they may be completely lacking. Quartz is often the only mineral that can be recognized in crystal tuff; where feldspar is discernible, a ratio of quartz to feldspar phenocrysts of 3:1 or 4:1 is common, whereas the ratio of quartz to feldspar phenocrysts in porphyritic rhyolite flows is about 1:1.

Points 1 and 4 are the most reliable criteria for recognizing fine-grained tuff or crystal tuff microscopically. Particularly the round clastic appearance of the matrix is characteristic of most of the fine-grained rhyolitic tuff and crystal tuff in this district.

A photomicrograph of rhyolitic tuff is shown in figure 18.

An unusual rhyolitic tuff containing pellets is found in the Iron Mountain area. It is a fine-grained, buff-colored, partly iron-stained tuff that contains spherical and ellipsoidal pellets of the same composition as the matrix (fig. 19). Most of the pellets range in size from spheres 2 millimeters in diameter to prolate ellipsoids 18 millimeters long with a minor axis of 10 millimeters. Small pellets are sparsely disseminated, whereas the larger ones may constitute as much as 30 percent of the rock.

The pellets are harder and more resistant to erosion than the matrix and stand out on weathered surfaces (fig. 19). These rocks are very fine grained clastic rhyolitic tuff beds composed of quartz, sericite, and hydromica. The pellets have a megascopic concentric structure, but commonly contain fragments that have microspherulitic structure.

Following the classification of Wentworth and Williams (1932, p. 37), this rock is an accretionary lapilli tuff. Similar tuff pellets have been called volcanic pisolites, mud balls, and mud pellets (Williams, 1926, p. 232). They have been described by Richards and Bryan (1927, p. 54–60) from the Brisbane tuff of Australia and by Pratt (1916, p. 450–455) from the 1911 eruption of Taal volcano in southwestern Luzon, Philippine Islands. The pellets were found on the slopes and the margin of Taal volcano 6–8 kilometers from the crater (Pratt, 1916, p. 451).

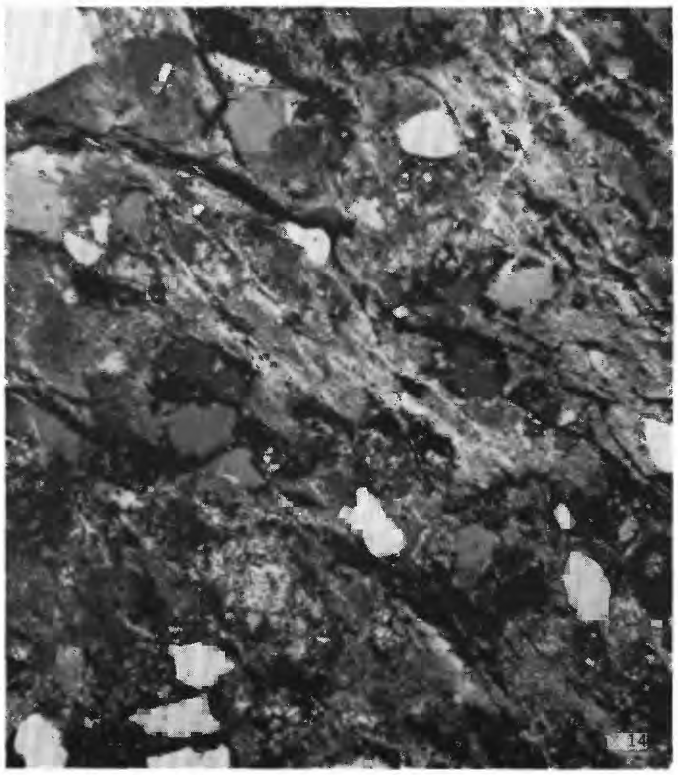

FIGURE 18.—Photomicrograph of bedded rhyolitic tuff. Fragments are all angular quartz clasts in a fine clastic matrix of quartz, sericite, and minor albite. Crossed nicols, × 14.

FIGURE 19.—Accretionary lapilli rhyolitic tuff. Most of the pellets are prolate ellipsoids having an average major axis of 6 millimeters. The composition of the pellets and the matrix is the same.

The tuff pellets are formed in the air by the aggregation of very fine tuff particles by condensing steam. Wentworth and Williams (1932, p. 37) note that similar pellets may be formed less commonly by rolling of lapilli nuclei over fresh ash surfaces. The pellets in the West Shasta district lack the lapilli nuclei and undoubtedly fell to the earth as spherical mud balls that were later distorted during orogeny. They indicate that part of the Balaklala rhyolite was above water when the pellet tuff was formed.

The lower and middle units of the Balaklala rhyolite contain some very fine grained dactic tuff beds. They are dark greenish-black rocks that commonly have a mottled appearance due to light-green spots or irregular light-colored veinlets. Megascopic grains of quartz may be present in an aphanitic matrix. The fine-grained dacitic tuff beds are composed of quartz, epidote, calcite, albite, zeolites, chlorite, and some clay minerals. The feldspars are saussuritically altered to epidote, calcite, and zeolites. The dark color is due to finely disseminated epidote; the mottled appearance of lighter green is caused by silicification.

ORIGIN OF THE BALAKLALA RHYOLITE

The Balaklala rhyolite was named and first described in detail by Diller (1906, p. 6), who described the Balaklala rhyolite as a series of siliceous lava and tuff beds. Graton (1909, p. 81) later studied part of the Shasta copper-zinc district and concluded that the Balaklala rhyolite of Diller was intrusive into the surrounding rocks, and he called it an alaskite porphyry.

An intrusive origin for the rhyolitic rocks was accepted by all the geologists who published information on this area after Graton's work (Averill, 1939, p. 122; Ferguson, 1914, p. 30; Hinds, 1933, p. 107; Seager, 1939, p. 1958–1959), although geologists for some mining companies continued to regard the rocks as flows and pyroclastic rocks.[2] Mapping by the writers has shown that Diller was correct in his belief that the rhyolitic rocks are of volcanic origin, and they have revived Diller's name of Balaklala rhyolite for the formation.

The disagreement among geologists on the mode of origin of the Balaklala rhyolite stems largely from the difficulties inherent in the interpretation of complex volcanic structures. Owing to inadequate exposures and insufficient information on the area as a whole, the origin and relationship of individual rock types and structures is difficult to determine. The essential problem in determining the origin of the Balaklala rhyolite is whether the rhyolite is entirely intrusive, whether it is in part older and in part younger than the Kennett formation, or whether it is a volcanic pile composed mainly of extrusive rock but containing some intrusive rock. An important part of this problem is the origin of the upper unit of the coarse-phenocryst rhyolite. The lower and middle units of the rhyolite are considered by the writers to be unquestionably extrusive because of the large amount of pyroclastic material interlayered in the rhyolite. The upper unit, however, is a broad, flat-lying body of massive rock as much as 1,400 feet thick that contains very little pyroclastic material; its mode of origin is not obvious. The massive character of this rock suggests that it might be either an intrusive sill or a laccolith between the Kennett formation and the middle unit of the Balaklala rhyolite, but the writers believe that the field evidence favors its formation as a volcanic dome extruded on the middle unit of the Balaklala rhyolite before Kennett time. The discussion of the origin of the coarse-phenocryst rhyolite is given in full, both because the upper coarse-phenocryst rhyolite contains the least internal evidence of its surficial origin, and because the ore-bearing zone lies at or a short distance below the base of the upper unit of the Balaklala, and this contact is used as a horizon marker throughout the district in the search for ore bodies.

The evidence that has been adduced by others in postulating the intrusive character of all or part of the Balaklala rhyolite can be summarized as follows:

1. Dikes or sills of rhyolite are reported to cut all the enclosing rocks, including the Kennett formation.
2. The absence of a basal conglomerate and of debris from Balaklala rocks in the overlying Kennett formation shows that the Kennett was not deposited on the surface of the Balaklala.
3. Breccias, which are interpreted as autobreccias in an intrusive, rather than as volcanic breccias.
4. The absence of glass in the rhyolite.

Evidence that is thought by some geologists of the mining companies to demonstrate the intrusive nature of the upper coarse-phenocrysts rhyolite is:

5. "Xenoliths" of shale occur in the coarse-phenocryst rhyolite.
6. The coarse-phenocryst rhyolite appears to have metamorphosed the overlying shale.
7. The intrusion of the rhyolite crumpled the overlying shale along the contract of the intrusive rock.
8. The lower and middle units of the Balaklala rhyolite are foliated but the upper unit is not, indicating intrusion after the foliation was formed.

A discussion of the specific points is given because it is recognized that some of the field evidence is controversial and open to several interpretations.

1. The tuff beds, coarse and fine pyroclastic material, and some shaly material that are interlayered with the rhyolite are adequately described under "Pyroclastic rocks." The large amount of pyroclastic material and the fish plate found in one tuff bed appear to the writers to give conclusive evidence that the rocks of the lower

[2] Based on private reports of the mining companies.

and middle units of the Balaklala rhyolite originated at the surface. However, many bodies of rhyolite have been intruded into their present position as sills, dikes, necks, and probably in a few places as breccia pipes; at many places, unless pyroclastic beds occur there are no petrographic features of the rhyolite bodies to indicate that one was intrusive and another was extrusive. Feeders for a rhyolite flow may cut all the underlying rocks, but the writers have found no locality where rhyolite intrudes the Kennett formation. Rhyolitic tuff beds, particularly crystal tuff, occur in the transition zone between the upper unit of the Balaklala rhyolite and the Kennett formation and in the lower part of the Kennett formation at some places. It is probably these occurrences, interpreted as rhyolitic sills, that led earlier writers to believe that rhyolite intrudes the Kennett formation.

2. Basal conglomerate and waterworn rhyolitic debris would not be expected at the base of the Kennett formation, as the fish plate that was found in tuff in the upper part of the Balaklala rhyolite, and other evidence given under "Geologic history," indicate that the Balaklala was deposited largely at or below sea level. Debris derived from a few volcanic islands that projected above water level would be similar in appearance to the waterworked tuff beds that are found in the transition zone between Balaklala and Kennett.

3. Bodies of breccia where the groundmass and fragments are alike are flow breccias formed by the incorporation of crustal material in a flow, rather than being examples of autobrecciation in an intrusive. They are layered rocks, commonly bounded by other types of pyroclastic layers.

4. Some textures that appear to be caused by devitrification of glass and some fluidal textures have been seen in thin sections, but they are rare. Apparently very little glass was formed in the rhyolites.

5. No xenoliths of shale of the Kennett formation in rhyolite were found by the writers. The xenoliths of shale between "sills" of rhyolite, reported at some of the mines, are interlayered rhyolitic crystal tuff and shale in the transition zone between Balaklala and Kennett formations.

6. The hard, bedded material under the coarse-phenocryst rhyolite that has been called metamorphosed shale or hornfels at the Mammoth mine was found to be a normal soft, shaly tuff where it was traced away from the mine. Thin sections showed that the hard, flinty character of the tuff bed at the mine was due to hydrothermal silicification.

7. The crumpling at the shale-rhyolite contact and the difference in the amount of foliation between the rocks of the middle and upper units of the Balaklala are due to their difference in competence as described under "Relationship to folds and foliation." Some minor crumpling is due to syngenetic sliding.

Evidence for an extrusive origin for the upper coarse-phenocryst rhyolite unit of the Balaklala in particular is:

1. The presence of a tuff bed which contains fragments of coarse-phenocryst rhyolite and coarse quartz crystals along the base of the dome of coarse phenocryst rhyolite. These beds indicate an explosive phase of the rhyolite before the extrusion of the thick dome. The base of the main body of the rhyolite is everywhere conformable to this tuff except at the vent.

2. The gradational contact at the top of the upper unit of the Balaklala rhyolite from coarse-phenocryst rhyolitic tuff to shaly tuff and shale at some localities indicates an explosive phase after the extrusion of the main body of coarse phenocryst rhyolite, and indicates continuous deposition.

3. The tuff and pyroclastic rocks above and below the main body of massive coarse-phenocryst rhyolite come together at the edge of the dome, and continue as one zone of tuff beyond the limits of the dome.

4. The presence of at least one coarse pyroclastic layer in the rhyolite dome, which contains rounded fragments of coarse-phenocryst rhyolite mixed with other varieties of rhyolite, shows that the formation of the dome was interrupted locally by a period of explosive activity during which a coarse pyroclastic layer was deposited.

5. Lack of any apophyses of rhyolite extending up into the shale of the Kennett formation is negative evidence that the coarse-phenocryst rhyolite is not intrusive, although a few dikes accompanying the rhyolitic tuff beds in the lower part of the Kennett might be expected.

KENNETT FORMATION

The Kennett formation is composed almost entirely of shale and limestone, but minor beds of shaly tuff and crystal tuff are interbedded with shale near the base of the formation. The shale forms the lower part of the formation; it is commonly a black siliceous thinly bedded rock which is locally crumpled and cut by a network of tiny quartz veins. The limestone forms the upper part of the formation and is largely a coral reef.

Fairbanks (1903, p. 48) first described the shale and limestone between Squaw Creek and Backbone Creek and north of Backbone Creek, and Smith (1894, p. 591) credits him with naming these strata the Sacramento formation in a manuscript. However, Smith (1894, p. 591) called these rocks the Kennett limestone and shale and implied in a table that the Sacramento formation and Kennett limestone and shale are equivalent.

Subsequently the term Sacramento formation was dropped and the term Kennett formation was used by later writers (Diller, 1906, p. 2; Graton, 1909, p. 79).

DISTRIBUTION

The Kennett formation is exposed throughout the northern part of the West Shasta district as discontinuous erosion remnants on present topography of a former much more continuous formation. Most of the Kennett caps hills or ridges, and is seldom found in the valleys.

The Kennett formation is most extensively exposed in the southern part of the Behemotosh Mountain quadrangle. It is exposed northeast of Backbone Creek on the southwestern slope of Backbone Ridge between the Bragdon formation to the northeast and the Balaklala rhyolite to the southwest, and as erosion remnants on the crests of hills and ridges southwest of Backbone Creek to Mammoth Butte. Good exposures are found above the road leading from Shasta Lake to the Golinsky mine and along the upper part of the road from Shasta Lake to the Mammoth mine. The Kennett is also exposed as a thin, discontinuous band as much as 800 feet wide between the Balaklala and the Bragdon formations in the northeastern part of the Whiskytown quadrangle and the southwestern part of the Behemotosh Mountain quadrangle from Mad Mule Mountain to Behemotosh Mountain, and as isolated remnants resting on Copley greenstone on the east side of Shasta Lake in the northeastern part of the Shasta Dam quadrangle.

THICKNESS AND RELATIONSHIP TO UNDERLYING ROCKS

The maximum thickness of the Kennett formation in the West Shasta district is probably not more than 400 feet, although Diller (1906, p. 2) describes a partial section 865 feet thick, and Stauffer (1930, p. 95–96) describes a section 815 feet thick. This discrepancy is in part due to repetition of beds by faulting and in part to a lesser thickness of material which is here included in the Kennett.

Geologic section northeast of Backbone Creek (Stauffer)

	Feet
Mississippian:	
Bragdon formation:	
9. Conglomerate with quartz pebbles and fossiliferous fragments, these latter often dissolved leaving holes	30
Devonian:	
Kennett formation:	
8. Shale, mostly dark, with thin sandy beds	150
7. Limestone, massive, light-gray containing a small amount of chert. Part is filled with corals	100
6. Sandstone, thin-bedded; and dark-gray shale	140
5. Limestone, dark-gray to bluish	10
Devonian—Continued	
Kennett formation—Continued	
4. Shale, cherty, gray	150
3. Limestone, thin-bedded to massive, often cherty and parts of it full of various types of corals. A black chert band at the base	200
2. Shale, siliceous, often sandy and usually very thin bedded. Shale mostly black to gray and often partly metamorphosed	65
Balaklala rhyolite:	
1. Rhyolite.	

The thickness of Kennett formation is here considered to be less than the 800+ feet shown by Diller and Stauffer because there appears to be repetition by faulting of the limestone and adjacent shale beds. There is no evidence that there is more than one limestone layer; although the outcrop is poor where the upper limestone beds are shown in Diller's and Stauffer's sections, these are probably fault slivers rather than separate beds of limestone. The Backbone Creek fault projects through where the dike of Birdseye porphyry cuts the Kennett at this locality and repeats part of the section.

The following geologic section was measured by the writers on a ridge in sec. 22, T. 34 N., R. 5 W., that projects westerly into the Backbone Creek arm of Shasta Lake immediately south of Lower Limestone Valley.

Geologic section northeast of Backbone Creek

[The base of the section is at lake level, altitude, 1,000 feet]

	Feet
Bragdon formation:	
Shale, light-brown and gray; and sandy shale	500+
No outcrop	250
Rhyolitic tuff, brown, sandy, not well bedded; some well-bedded sandy mudstone. Some of the tuffaceous rock contains 1/8 to 1/4 inch lithic fragments (poor exposure)	50±
Kennett formation:	
Mudstone, dark-gray, siliceous, possibly cherty, similar to the siliceous dark-gray shale that weathers light gray below the limestone (poor exposure)	20±
Mudstone, limy, siliceous, 2- to 6-inch beds. Abundant martinioid brachiopods in lower 6 inches	6
Limestone, light-gray to bluish-gray, bedded. Contains abundant corals. The upper 25 feet contains brachiopods and cup corals (lower contact not exposed)	200±
Shale, black, siliceous, in part cherty, 1- to 6-inch beds. Locally crumpled	75±
Balaklala rhyolite:	
Porphyritic rhyolitic tuff containing 3- to 4-millimeter quartz crystals. The tuff is well bedded but does not part on bedding planes. Bedding is marked by color bands and by compositional and grain-size variations. The top 6 inches to 3 feet is slightly crumpled and iron stained and contains small, local areas of gossan and minor siliceous black shale lenses mixed with the tuff	40±
Porphyritic rhyolite containing 1- to 2-millimeter quartz phenocrysts. Light-gray to light-green felsitic matrix	200+

This is the most complete section now available above the level of Shasta Lake, but it is incompletely exposed. Measurements start from the lakeshore at an altitude of 1,000 feet, and the stratigraphic thicknesses were measured by altimeter readings or were estimated because the beds are locally crumpled and exact contacts are not exposed.

Stauffer's section includes the shale above the limestone as far as the first conglomerate of the Bragdon formation as part of the Kennett formation and gives the thickness of shale above the limestone as 150 feet. Diller does not mention the first overlying conglomerate specifically, but he also includes 140 feet of shale above the limestone as a part of the Kennett formation, and presumably placed the top of the Kennett at the first conglomerate bed above the limestone. The writers see no advantage and much difficulty in including shale above the limestone as part of the Kennett formation. There is no change in the lithology of the shale above and below the first conglomerate bed above the limestone, the contacts are conformable, and deposition apparently has been continuous. Also, the conglomerate beds are not everywhere present, and even where present they are lenticular, and the "first" conglomerate in one area is not at the same stratigraphic horizon as the "first" conglomerate in a nearby area. Therefore the writers have considered the shale below the first conglomerate in this locality to be part of the Bragdon formation, and they have placed the contact of Kennett and Bragdon at a lithologic break at the top of the limestone where it is present or at the top of black, siliceous shale where limestone is absent.

The maximum thickness of the Kennett formation appears to be about 400 feet in the West Shasta district. The greatest thickness of the Kennett is exposed southwest of Backbone Creek on Quarry Ridge east of the Golinsky mine. In this area the uppermost part of the limestone has been eroded in the present erosion cycle and the rocks are locally crumpled, but a thickness of 350 to 400 feet of the Kennett is exposed. On the ridge between Little Backbone Creek and Squaw Creek the Kennett is so crumpled that no accurate measurement of thickness is possible, but an estimate of thickness based on lithology is 300 to 400 feet.

The shale, sandstone, and tuff beds of the Kennett formation grade downward conformably into breccia, tuff, and flow-banded facies of the Balaklala rhyolite. This gradation is clearly shown in many outcrops scattered throughout the area. The best exposures showing this sedimentary relationship are on the northeast side of Backbone Creek between the Upper Limestone and Lower Limestone Valleys and just east of the Mammoth mine on the road leading from Shasta Lake to the mine. The contact of Balaklala and Kennett formations also is exposed in some of the tributaries both north and south of Little Backbone Creek and in Butcher Creek and the small gullies east of Butcher Creek.

The lower contact of the Kennett is placed between the dark-gray to black cherty or sandy shale and the light-tan to light-green rhyolite or rhyolitic tuff. In some places the contact is sharp; in others it is gradational; for example, east of the Mammoth mine, the contact is placed where shale predominates over rhyolitic tuff. On Behemotosh Mountain limestone rests directly on rhyolitic tuff of the middle unit of the Balaklala rhyolite. No shale is present under the limestone, and no bedding can be seen in the tuff; it seems probable that this locality is one in which the tuff was deposited above sea level and remained above it during early Kennett time. The tuff sank below or was eroded to sea level in late Kennett time, and coral limestone was deposited on the tuff.

RELATIONSHIP BETWEEN THE KENNETT AND THE BRAGDON FORMATIONS

Diller (1906, p. 2–3) and Stauffer (1930, p. 95–96) believed that the Kennett formation was overlain unconformably by the Bragdon formation and that much of the Kennett was eroded before the Bragdon was deposited. Their evidence for this is that the Kennett differs in thickness in several parts of the area, that it is missing in some areas, and that fragments of Kennett limestone are present in the Bragdon conglomerate. The writers believe that the Kennett is essentially conformable under the Bragdon in the West Shasta district. Although they recognize that minor erosional unconformities and depositional overlaps occur locally, fairly well exposed sections across the Kennett-Bragdon contact on the northeast side of Backbone Creek and along the continuous band of Kennett that crops out in the northeast quarter of the Whiskytown quadrangle show no evidence of an angular unconformity, channeling, or a major erosion period before the deposition of the Bragdon in these areas. Sedimentation seems to have been continuous in these areas; it is improbable that an interval of erosion sufficient to remove the limestone and some of the Kennett would have left a layer of shale of the Kennett of uniform thickness for an outcrop length of 5 miles, such as occurs in the Whiskytown quadrangle.

The limestone that forms a thick bed in the Behemotosh Mountain quadrangle occurs in the Whiskytown quadrangle as only a few scattered lenses; but such isolated lenses also occur in the Kennett east of the Behemotosh Mountain quadrangle. There is no evidence that the limestone was once a continuous bed over the entire area which was removed by erosion before the deposition of the Bragdon, but there is evidence of patchy deposition of limestone, and there are considerable areas where it was not deposited. It is a coral reef-type limestone; its greatest thickness probably is at Backbone Creek, and it thins to isolated patches away from this center. In places where the limestone is missing, the shales of the Bragdon and Kennett are conformable and the contact is drawn at the upper limit of the dark siliceous shale that characterizes the Kennett. It seems probable that the volcanic pile of Copley and Balaklala rocks was built up to or near sea level and coral reefs formed on the seamount and rimmed projecting volcanic islands; cessation of volcanic activity, erosion of volcanic islands, and continued subsidence lowered the surface below the coral zone and allowed the deposition of Bragdon sediments.

The variation in the thickness of the shale in the Kennett formation, and absence of the shale in some areas, is not regarded as evidence for a major erosional unconformity, although local erosion undoubtedly occurred. The variations in thickness that are found can be explained by unequal deposition on a surface of relief, erosion by currents of high points on the sea floor, differential compaction over topographic highs, and syngenetic sliding. The Kennett was not deposited in areas that were above sea level; erosion of these areas contributed part of the tuff that is in the lower part of the shale of the Kennett in some areas. In the vicinity of the Mammoth mine, evidence for the deposition of the Kennett on a surface of relief is the different thickness and type of material in the Kennett. The thickness of the Kennett rocks below the limestone ranges from a few feet to slightly more than 100 feet, and the composition ranges from siliceous black shale to mixed shale and tuff and in one locality to a crystal tuff. The nature of the limestone, a coral-reef type, is further evidence of a nearshore environment.

As great masses of the volcanic rocks which formed the Balaklala rhyolite were poured out on the sea floor, it is probable that the base of the volcanic pile subsided as the top rose, and that the submarine volcanic pile was a broad low dome. Large basins, such as the one north of the Mammoth mine were present, in which considerable thicknesses of bedded, water-deposited tuff and volcanic conglomerate collected; they indicate a local surface of low relief. Some of the material appears to be waterworn and may have been derived from projecting volcanic islands. On the other hand, it is possible that some of the Kennett was deposited on fairly steep initial slopes; syngenetic sliding at the base of the Kennett indicates that the Kennett was deposited on a sloping surface.

The Kennett formation did not completely cover the volcanic seamount, parts of which were at times above sea level. In some places local topographic highs in the dome of Balaklala rhyolite were present, as north of Spring Gulch in the Whiskytown quadrangle just west of peak 3,893. The Kennett never covered these local peaks so that the Bragdon formation was deposited as an overlap on the Balaklala with no erosional unconformity . Also, the Kennett sedimentary rocks were not deposited on the Balaklala over the center of the volcanic highland, as in the NW¼ sec. 14, T. 33 N., R. 6 W., shales of the Bragdon rest directly on soft porphyritic rhyolitic tuff.

As described under "Geologic history," fragments of fossiliferous limestone that have been identified as limestone of the Kennett in conglomerate beds in the Bragdon prove that some limestone of the Kennett was eroded and incorporated in the younger strata of the Bragdon. The lack of evidence of erosion except locally, and the evidence of continuous deposition in much of this area, however, implies that warping occurred; the West Shasta district remained largely a depressed area of essentially continuous sedimentation while areas outside this district were uplifted and eroded, and contributed debris to the Bragdon sediments.

LITHOLOGIC DESCRIPTION

The Kennett formation is composed predominantly of dark-gray to black siliceous shale, and massive fossiliferous limestone; rhyolitic tuff and tuffaceous shale are prominent in the lower part of the formation. Some conglomerate, limy shale, sandstone, and porphyritic rhyolite occur locally in the black shale. The fossiliferous limestone and black siliceous shale are the distinctive lithologic units by which the Kennett is recognized (fig. 20).

FIGURE 20.—Kennett formation northeast of Backbone Creek arm of Shasta Lake. View looking northeastward. Bold limestone outcrop of the Kennett formation on ridge (*a*) : black siliceous shale of the Kennett formation (*b*) ; tuff in the Balaklala rhyolite (*c*).

Black siliceous shale layers, which locally are cherty, compose the lower part of the Kennett at most localities, although interbedded tuff and shale form the base of the formation at some places. Except for two small localities of limestone, the basal black cherty shale is the only Kennett stratum that is present southwest from Behemotosh Mountain to Mad Mule Mountain in the Behemotosh Mountain and Whiskytown quadrangles. No fossils were found in this black siliceous shale layer, but it was correlated with the Kennett because of its stratigraphic position and because of the similarity in the lithologic character of the black siliceous shale to that present in the Kennett elsewhere. Two small lenses of limestone in the black siliceous shale further suggest that this layer is Kennett. Black siliceous shale is also present under the limestone in most exposures in the Behemotosh Mountain quadrangle; it ranges in thickness from a few inches to 150 feet. The black shale layers are in part thinly bedded; bedding planes marked by color bands range in width from less than a millimeter to several inches; parting planes are 1 to 6 inches apart. The black shale layers are commonly contorted and are cut by many thin quartz veinlets.

All the black shale beds are siliceous and carbonaceous and contain quartz, carbonaceous material, sericite, clay minerals, limonite after pyrite, sphene, and rutile. They contain angular quartz grains averaging about 0.1 millimeter in diameter in a fine-grained matrix of carbonaceous material, sericite, quartz, clay minerals, sphene, and rutile in which the grains average about 0.01 millimeter across. Locally the black shale is cherty and contains poorly preserved structures that are probably radiolaria. The cherty units are brittle and were crackled during orogeny. The fractures are filled with quartz, but there was little or no introduction of quartz through the matrix. Locally two sets of quartz veinlets, which intersect the bedding at 45°, are prominent. The black siliceous, clastic shale beds rarely contain the abundant quartz veinlets that occur in the cherty units.

Other rocks in the Kennett formation below the limestone are coarse- to fine-grained tuff, conglomerate, sandstone, siltstone, and porphyritic rhyolite. All these rock strata are lenticular, and collectively they form units of heterogeneous clastic and pyroclastic material. The largest areas of these lenticular masses of tuffaceous sediments are in the Butcher Creek basin southeast of Mammoth Butte and in the S½ sec. 26 and the NW¼ sec. 35, T. 34 N., R. 5 W., northeast of the Backbone Creek arm of Shasta Lake. Locally a particular bed within this zone is distinctive enough to be traced in mapping, for example, the coarse-grained tuff bed that lies under black cherty shale on the north side of the Squaw Creek arm of Shasta Lake. This coarse-grained tuff layer extends from Butcher Creek eastward to the point where the divide separating Squaw Creek from Little Backbone Creek intersects the lake level—a distance of about 2 miles. This bed averages about 100 feet in thickness, and it is composed of pea-sized fragments of rhyolite and greenstone. It is considered part of the Kennett rather than Balaklala rhyolite because at the eastern end it overlies black cherty shale. Conglomerate occurs in small, irregular pods in this heterogeneous zone, but makes up less than 1 percent of the Kennett at this locality and is not present elsewhere. The conglomerate is composed of angular to subrounded fragments of shale and chert averaging half an inch in diameter embedded in a silty matrix. Individual lenses of conglomerate commonly are 1 to 3 feet thick and 5 to 10 feet long.

A layer of flow-banded porphyritic rhyolite that has small dark-gray and colorless quartz phenocrysts is exposed in the vicinity of Butcher Creek. This porphyritic rhyolite appears to lie conformably between sedimentary layers of the Kennett formation; it has an average thickness of about 30 feet and can be traced with a fair degree of certainty for about 1 mile. A similar body of rhyolite that appears to be interbedded with shale is located about 1,000 feet north of the one at Butcher Creek, but poor exposure and many faults make its inclusion in the Kennett formation questionable; it may be an inlier of Balaklala rather than interbedded in the Kennett rocks.

Rhyolitic tuff beds and possibly thin rhyolitic flows are interbedded with sandy or tuffaceous shale at some places and these tuff beds are gradational to porphyritic rhyolitic tuff at the top of the Balaklala rhyolite. A good exposure of the gradational contact between the Balaklala and Kennett formations is on the ridge east of the Mammoth mine. At this locality there is a gradual change, through a stratigraphic thickness of about 100 feet, from porphyritic and nonporphyritic tuff and tuff breccia containing some sandy tuff and gray shale beds upward to predominantly gray to tan shale and a few rhyolitic tuff beds. Most of the rhyolite in this transition zone is definitely tuffaceous, but some consists of massive light-green felsitic rocks that may be either sills or flows. No crosscutting relationship was observed. The contact between the Balaklala rhyolite and the Kennett formation is also well exposed east of the Golinsky mine (fig. 12).

The considerable thicknesses of pyroclastic rocks of Balaklala rhyolite type that occur in the lower part of the Kennett formation indicate that eruptions of Balaklala rhyolite type rocks continued into Kennett time. Some feeders for the overlying pyroclastic rocks may cut the lower strata of the Kennett, but no such crosscutting relationship has been observed by the writers. However, much of the rhyolitic pyroclastic rocks interbedded with the Kennett are water deposited; they are not necessarily derived from explosive activity during Kennett time, but may in part be due to the reworking of rhyolitic tuff on volcanic islands and the deposition of these tuff beds along shorelines where they were interlayered with shale of the Kennett.

The limestone in the Kennett formation is a thinly to thickly bedded but in part massive rock that is light to bluish gray. The lower part of the limestone is mostly massive or thickly bedded and contains only sparse corals. The upper 50 feet is highly fossiliferous and contains abundant coral debris and some chert nodules and bands. Most of the corals are concentrated in definite bands in the limestone. A few brachiopods occur in the upper 25 feet but were not found in the lower part of the limestone. Much of the upper part of the limestone is of clastic origin.

Limestone talus commonly covers the contact between the limestone and the underlying shale. The writers found this contact exposed at only one point on the east side of Backbone Creek in Dark Canyon at an altitude of 1,550 feet. Figure 21 gives a section across the contact at this point. Balaklala rhyolite-type rhyolitic tuff thus is present at both the upper and lower contact of the Kennett limestone.

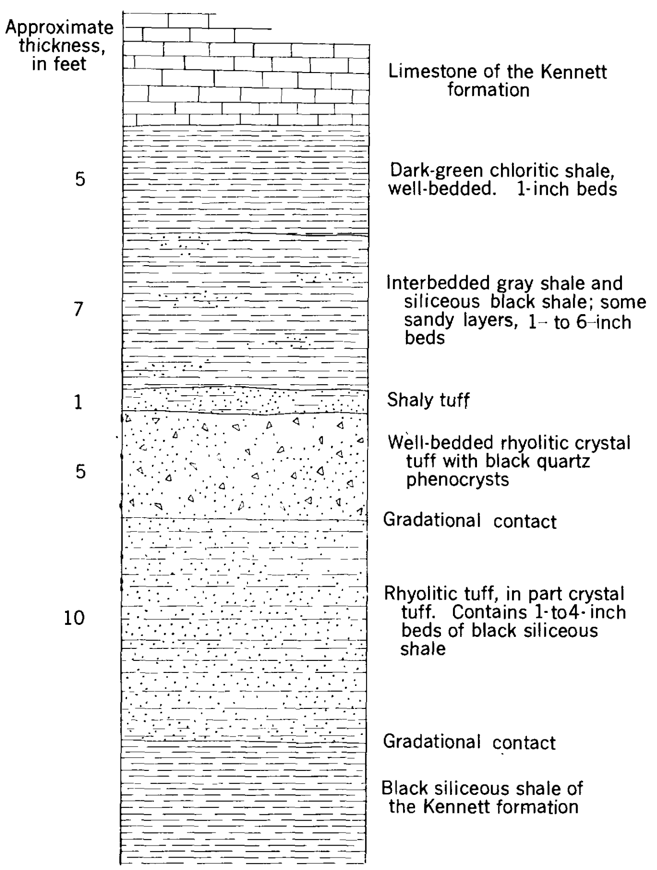

FIGURE 21.—Partial section beneath the limestone of the Kennett formation on the east side of Backbone Ridge, Behemotosh Mountain quadrangle.

The limestone is all very fine grained, but is not recrystallized to marble. Thin veinlets filled with white calcite are common in the limestone; in places these tend to dissolve out at the surface. Void spaces left by solution of these veinlets and vertical fluting (lapiés) formed by rainwater, make the surface of the limestone quite cavernous.

Logan (1947, p. 324) has the following statement on the composition of the limestone of the Kennett:

> The few analyses available show 95 to 97 percent calcium carbonate, 1 to 4.4 percent silica, 0.5 to 2.25 percent magnesium carbonate, and very little iron oxide and alumina. However, it must have been on the average of good quality as lime made from it was used over a large part of northern California, the old kiln at Briggsville having supplied lime to many pioneer towns of early mining days, and the deposits near Kennett having been in operation from at least 1884 until 1925.

This limestone was mined from a quarry on the ridge in the W½ sec. 34, T. 34 N., R. 5 W. An analysis of the limestone of the Kennett from a sample collected by the writers in sec. 4, T. 33 N., R. 5 W., along the road from Shasta Lake to the Mammoth mine was made by the Calaveras Cement Co.

Analysis of limestone of the Kennett formation

	Percent
SiO_2	0.22
Fe_2O_3	.12
Al_2O_3	.26
$CaCO_3$	99.84
MgO	0
	100.44

AGE

The age of the Kennett formation has been established as Middle Devonian from fossil collections made by Fairbanks, and referred to by Smith (1894, p. 591), and from collections by Diller (Diller, 1906, p. 2; Diller and Schuchert, 1894, p. 416–422), Stauffer (1930, p. 95–96), and the writers.

Diller collected fossils from shale and limestone in the Kennett formation. He collected the fossils from the shale near the Sacramento River a short distance north of Morley, which was 2.6 miles N. 25° E. of the former town of Kennett, and from the Kennett formation near Backbone Creek. This fossil locality is now flooded by Shasta Lake. Charles Schuchert studied Diller's collection and concluded that it is of Middle Devonian age (Diller, 1906, p. 2).

A fossil collection was made of the upper 25 feet of the limestone of the Kennett formation and the overlying 6-foot calcareous shale bed from a ridge northeast of the Backbone Creek arm of Shasta Lake in sec. 22, T. 34 N., R. 5 W. and referred to G. Arthur Cooper of the U. S. National Museum, who reports:

> The specimens from the Kennett limestone definitely belongs to the Middle Devonian. Those specimens from the limy shale horizon may be Middle Devonian but might be from the lower part of the Upper Devonian. The martinioid is difficult to identify but suggests Devonian relationships rather than Mississippian.

Dr. Cooper identified the following fauna:

Calcareous shale bed:
- Martinioid brachiopod
- *Leiorhynchus* (?) sp.

Limestone of the Kennett formation:
- *Heliolites* sp.
- *Amphipora* sp.
- *Atrypa* sp.
- *A.* sp. aff. *A. spinosa*
- *Schizophoria* sp. (large)
- Cup corals (poor)

BRAGDON FORMATION

DISTRIBUTION

The Bragdon formation, which was named from the community of Bragdon on the Trinity River 8 miles south of Trinity Center by Hershey (1904, p. 347–360), occurs as an arcuate band 17 miles long and as much as 4 miles wide at the north end of the West Shasta district. It is part of a much larger area continuing northward that covers most of the Schell Mountain and Lamoine quadrangles between the Trinity and the McCloud Rivers. Erosion remnants of the Bragdon also cap the peak 3,500 feet northeast of Behemotosh Mountain in the Behemotosh Mountain quadrangle and in the NW¼ sec. 14, T. 33 N., R. 6 W., in the Whiskytown quadrangle. Shale and grit that are probably part of the Bragdon formation are exposed also on Jerusalem Creek 10 miles west of Igo, on the southwest flank of the Shasta Bally batholith. The Bragdon originally covered the West Shasta district, but was entirely eroded from some areas before deposition of the Cretaceous sediments.

THICKNESS AND STRATIGRAPHIC RELATIONSHIP

The Bragdon formation overlies the Kennett formation, but in some places the Kennett was not deposited or was locally eroded, and in these places the Bragdon rests upon either the Balaklala rhyolite or Copley greenstone. On the west side of the Whiskytown quadrangle the Bragdon rests directly on Copley, and the Kennett probably was not deposited here, as the band of black siliceous shale of the Kennett that extends almost continuously for a distance of about 15 miles to the northeast does not underlie the Bragdon continuously west of Iron Mountain.

Much of the contact between the Bragdon formation and Balaklala rhyolite in the Behemotosh Mountain quadrangle is a fault contact. The fault along Backbone Creek drops the Bragdon into contact with the Balaklala for a distance of 2.75 miles. A fault contact also separates the Balaklala and Bragdon northwest of Behemotosh Mountain in the Behemotosh Mountain quadrangle and in sec. 22 (projected), T. 33 N., R. 6 W., in the Whiskytown quadrangle. The contact between the Copley and Bragdon in the northwestern part of the Whiskytown quadrangle north of Shirttail Peak is a fault contact striking northeast that brings the Copley up on the northwest side against the Bragdon.

At most places in the mapped area the contact between the Kennett and Bragdon formations is sharp and conformable, and was placed at the top of the black siliceous or cherty shale of the Kennett or at the top of the limestone where it overlies the cherty shale beds. On the northeast side of the Backbone Creek arm of Shasta Lake, it was difficult to place the contact. A lenticular mass of interlayered sandy, rhyolitic tuff, calcareous shale, and dark nonsiliceous shale separate limestone of the Kennett and dark siliceous and cherty shale from the typical light-brown and gray shale of the Bragdon. The record of sedimentation here appears to be complete; there are no signs of channeling or erosion of the top of the limestone. The only feature that suggests an erosional unconformity is some red-banded sandstone above the limestone that resembles sandstone oxidized at the surface. The sedimentary rocks in the Backbone Creek locality grade upward from calcareous shale through sandy and tuffaceous beds to tan shale layers that are uniform as far as the first conglomerate of the Bragdon. In the eastern part of the Whiskytown quadrangle a sharp lithologic break between cherty and silicified shale and noncherty shale marks the contact between Kennett and Bragdon rocks, but the two types of shale are conformable.

The top of the Bragdon is not exposed in the West Shasta district. Diller (1906, p. 3) reports, however, that the Bragdon is overlain conformably by the Baird formation northeast of the West Shasta district in the Lamoine quadrangle. He places the top of the Bragdon at the top of the highest conglomerate bed.

In the Behemotosh Mountain quadrangle a partial section of the Bragdon has a thickness of 3,500 feet from Backbone Creek to the north edge of the mapped area on Backbone Ridge, but this is only the lower part of the formation, and the thickness may be considerably greater, as the Bragdon extends northward from the mapped area for a distance of about 11 miles.

The thickness of the Bragdon may be considerably more than 3,500 feet in the Whiskytown quadrangle, but the whole area is so broken up by faults that no section can be measured or even closely estimated. An ill-defined zone of conglomerates occurs in the lower part of the Bragdon, but a few conglomerate beds occur above and below this zone. The base of the main zone of conglomerate in the Bragdon in the Whiskytown quadrangle is about 500 feet above the base of the Bragdon and the top of the zone is about 1,500 feet above the base of the Bragdon.

The least faulted partial section of the Bragdon in the Whiskytown quadrangle extends from the Balaklala-Bragdon contact near the Bright Star mine in the Whiskytown quadrangle N. 26° W. to the center of the syncline 3,000 feet north of hill 3564, which is south of Cline Gulch; the thickness of the Bragdon in this area is 1,450 feet. From the center of this syncline eastward the Bragdon has a fairly uniform strike of N. 10° to 30° W. and dips 15° to 70° W. These dips give an apparent thickness to the Bragdon of 15,000 feet. However, most of this area contains the main conglomerate zone of the Bragdon, and apparently is mostly within the lower 2,000 to 3,000 feet of the Bragdon which is here cut by a series of northwest-striking faults that drop the west side and continually repeat the section. These faults are generally poorly exposed and occupy debris filled canyons, but they are marked by the abrupt termination of a series of conglomerate beds. Some of the faults have prominent topographic expression as seen on aerial photographs.

Diller (1906, p. 3) stated that the maximum thickness of the Bragdon in the Redding (30-minute) quadrangle may be as much as 6,000 feet, but that the broad area of the Bragdon is so affected by small folds that its thickness is difficult to determine. He states that south of Castella on Hazel Creek, where the Bragdon forms a band between the Kennett and Baird formations, the thickness is estimated at 2,900 feet.

LITHOLOGIC DESCRIPTION

The Bragdon formation in the West Shasta district is composed of interstratified shale, siltstone, sandstone, grit, conglomerate, and a small amount of rhyolitic tuff, and at one locality contains a small flow of mafic lava. Beds of shale make up more than 75 percent of the formation. They are dark greenish gray to black on unweathered surfaces, but are buff to brown on weathered surfaces. The shale is thinly bedded, and the stratification is easily recognized by parting and by color differences, or by variations in grain size in adjacent laminae. Thin beds of sandy shale and grit interbedded with the shale also show the structure in the shale. Bedding in the shale is very regular. In places graded bedding is common, but little crossbedding or channeling was observed.

Thin siltstone and fine-grained sandstone layers are interbedded with shale throughout the Bragdon. The sandstone beds are generally a few inches to a few feet thick, but beds of gritty sandstone as much as 50 feet thick occur north of Backbone Creek and near the head of Spring Creek. These beds are commonly light gray or buff colored, although some are dark gray. The dark sandstone beds contain abundant rock fragments of shale and chert and are graywacke.

The characteristic beds of conglomerate in the Bragdon are the most distinctive feature of this formation; they do not occur in older or younger formations. The only known occurrence of coarse detrital material at the base of the Bragdon in this district is just above the junction of Jackass and Backbone Creeks along the north edge of the mapped area. At this locality a bed of grit 10 to 20 feet thick lies along the contact between shale of the Bragdon and the underlying Balaklala rhyolite. None of the Kennett formation is exposed at this locality, although Kennett strata crop out a short distance to the southeast. The grit at the base of the Bragdon is composed of angular chips of chert and quartz, one-eighth to one-quarter inch in diameter, in a dark-gray sandy matrix. Other than this, basal conglomerate and grit are lacking along the base of the Bragdon.

Conglomerate beds are concentrated between 500 and 1,500 feet above the base of the Bragdon in one general zone. However, some thin beds occur higher in the section north of Backbone Creek, and lower in the section northwest of Iron Mountain, but they are more lenticular and less abundant in these areas. Within the general zone there are usually 4 to 6 persistent conglomerate beds that can be traced from a few thousands of feet to several miles whereas many lenticular beds can be traced only for a few hundred feet. The general zone contains more individual conglomerate beds in the Whiskytown quadrangle than in the Behemotosh Mountain quadrangle. These individual sandstone and conglomerate beds reach their maximum thicknesses near the head of Spring Creek.

The conglomerate beds are most commonly 10 to 20 feet thick, but they range from several feet to as much as 100 feet in thickness. There is no apparent relationship between the thickness of the bed and the size of the constituent pebbles. The thicker beds are generally composed of several conglomerate layers separated by siliceous gritty sandstone and minor shale which together make a persistent, mappable unit. The conglomerate beds within the general zone are also separated by shale, which locally is extremely contorted. As the conglomerate beds are much more resistant to erosion than the shale, they form conspicuous, bold outcrops. The large areas of conglomerate in the Whiskytown quadrangle, such as that at the head of Spring Creek, are in part dip slopes.

The contacts between conglomerate and shale beds are sharp and smooth, although channeling is present at the top of a few conglomerate beds . In no place was a transition zone observed between conglomerate and shale, although there are transition zones between conglomerate and grit and sandstone.

The matrix of the conglomerate beds is hard and siliceous, and the rock commonly breaks smoothly across the pebbles. Some of the finer beds, composed mainly of closely packed chert pebbles, have a siliceous cement with scarcely any sandy matrix. These finer beds are commonly cut by thin quartz veinlets that pass through the pebbles. The coarser conglomerate beds, which are poorly sorted, commonly have 25 to 50 percent dark-sandy or gritty matrix material.

The conglomerate is composed mainly of black and light-gray or white chert pebbles, but it also contains pebbles of dark-gray, tan, and red chert, banded chert, vein quartz, shale, sandstone, and limestone in a dark- or light-colored siliceous, sandy, or gritty matrix. The pebbles range in outline from angular to subrounded, and within most beds they are fairly uniform in size; they constitute from 50 to more than 90 percent of the rock. Most of the pebbles are $\frac{1}{4}$ to $1\frac{1}{2}$ inches across, although fragments as much as 1 foot in diameter are present in a few beds.

Sphericity of the pebbles is poor, although most have well-rounded corners; large fragments are more rounded than the smaller ones. The length of the shale fragments commonly is four or five times greater than the thickness. The chert pebbles have an oblate shape in most beds and are well oriented (fig. 22). In other beds the pebbles are equidimensional, more angular, and consist of several lithologic types. Pebbles of limestone and sandstone are most abundant in the coarser conglomerate beds. The limestone pebbles are weathered out at the surface, giving a vuggy appearance to the rock. Many of the limestone fragments contain corals.

The conglomerate of the Bragdon formation as a whole probably should be classed mainly as a polymictic conglomerate, even though most of the beds are composed largely of chert pebbles, because they contain metastable fragments such as limestone and shale, and the interstitial material of most beds is a graywacke. Many oligomictic conglomerate rocks are present, however; they consist almost entirely of chert pebbles in a light-colored matrix, and generally show much better

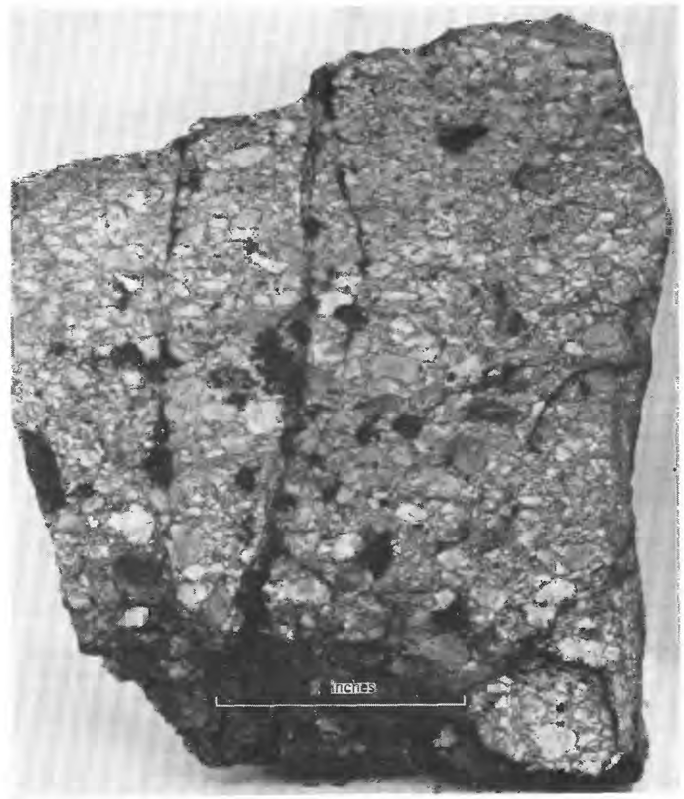

FIGURE 22.—Chert conglomerate of the Bragdon formation. Fragments are subrounded and well oriented and almost all are white, gray, or tan chert. Smaller fragments are more angular than the larger ones.

sorting and rounding than the polymictic conglomerate. The large pebbles or boulders are all contained in the polymictic conglomerate beds.

Few fragments of definitely igneous rock are present in the conglomerate beds. Several pebbles of rhyolite containing feldspar phenocrysts were found in the conglomerate on the west side of the Sacramento arm of Shasta Lake. The writers identified rhyolite, greenstone, and fragments of quartz and feldspar phenocrysts in thin sections of rocks collected by C. M. Gilbert, from the divide north of Blue Mountain on the road from Trinity Center to Volmer, north of the West Shasta district.

Diller (1906, p. 3) states that most of the Bragdon was derived from the Kennett, and the following is quoted from him:

> Fossils collected from pebbles in the conglomerate at a large number of localities throughout the Bragdon area were referred to Professor Schuchert and Dr. Girty, and all of them so far as determinable, with one possible but doubtful exception found on Bailey Creek, were reported as Devonian, like those already known in the Kennett region, and show clearly that the conglomerate is later than the middle Devonian.

The limestone and dark-chert pebbles are similar to the chert and limestone of the Kennett, but the light-gray and red chert and the sandstone pebbles in the conglomerate of the Bragdon are unlike anything the writers observed in the Kennett in this area, and they were probably derived from a banded chert. They may have been derived from the Chanchelulla formation of Hinds (1932) southwest of the West Shasta district; the rocks of Hinds' Chanchelulla, which apparently lie under rocks correlated with the Copley greenstone, are described as containing thinly bedded and banded gray and less commonly red and green cherts (Hinds, 1933, p. 85).

North of Backbone Creek two black siliceous shale bands each about 100 feet thick are interbedded with conglomerate and shale of the Bragdon, but this is the only locality at which siliceous shale was found in the Bragdon. Lithologically these beds are similar to the black siliceous shale of the Kennett.

A small band of amygdaloidal greenstone is interbedded in the shale of the Bragdon along the northwest edge of the mapped area, about 2,500 feet west of the junction of Beartrap and Backbone Creeks. Exposures are poor in this area, but the greenstone is apparently a flow about 150 feet thick that is interbedded in the shale of the Bragdon. The flow is a dense dark amygdaloidal greenstone that resembles the Copley greenstone.

The relatively thin, even bedding of the shale and siltstone, the repeated alternate bedding of clay- and silt-sized material, the sparsity of fossils, and the absence of ripple marks all suggest that the Bragdon formation originated as an offshore deposit on a slope below wave base (clino environment of Rich) (Rich, 1950, p. 717–741; 1951, p. 1–20). However, the widespread but lenticular conglomerate and sandstone in the Bragdon indicates either that deposition was partly under shelf conditions, or that conglomerate beds were spread from the shelf area onto the upper part of the slope during periods of storm. The latter assumption seems most probable as the conglomerate and sandstone beds are lenticular and erratic in vertical distribution. The slope was probably gentle, so no syngenetic sliding occured in the shale.

Interbed movement and low-angle thrust faults of small displacement are common at the base of the Bragdon. These are particularly prominent along Backbone Creek due to movement in the incompetent shale in the Bragdon above the competent Balaklala rhyolite; some of the thrust faults in this area may be related to movement along the steep fault in Backbone Creek.

AGE

The Bragdon formation is noteworthy because of the paucity of fossils. No fossils were found in 22 square miles of Bragdon that was mapped in the West Shasta district; however, outside this area some fossil localities are known. Hershey (1904, p. 347–360) regarded the Bragdon as being of Jurassic age. Diller (1906, p. 3; 1905, p. 379–387) proved the Bragdon to be of Paleozoic age, and he has the following report on the age of the Bragdon from fossil localities in the northern part of the Redding 30-minute quadrangle about 12 miles north of the West Shasta area.

> In the sandstones and shales fossils were found at half a dozen localities. An important occurrence is upon the divide southwest of Nawtawakit or High Mountain, where the sandstones, conformably interbedded with characteristic Bragdon conglomerate, contain shells which Dr. Girty reports as "Paleozoic and without much doubt early Carboniferous, related to the Baird." The fossils, among which is a large "*Spirifer* of the *striatus* type," occur in several beds well exposed and are undoubtedly of the Bragdon horizon.
>
> Perhaps the most important locality is beside the railroad 1½ miles northeast of Lamoine, where fossils were found in the sandstone adjoining the Bragdon conglomerate. From this locality Dr. Girty reports *Schizodus* sp., *Loxonema* sp., *Pleurotomaria?* sp., and *Straparollus* aff. *S. Luxus*. There is no room for doubt that these fossils belong to the Bragdon and are not derived from an older formation, and Dr. Girty remarks that if this be admitted "no other conclusion is possible than that the Bragdon is a Paleozoic formation. Indeed, it is clearly safe to say that the horizon is not later than Baird, for the local faunas have many points of resemblance with that of the Baird and none at all with those of the overlying Carboniferous formations."

CHICO FORMATION

Shale, sandstone, and conglomerate of the Chico formation crop out in the southeastern part of the Shasta Dam quadrangle. The continuation of this exposure of the Chico south and east of the Shasta Dam quadrangle was mapped by Diller (1906, p. 6). A small area of conglomerate of the Chico occurs in the southeast corner of the Igo quadrangle.

The Chico formation in the Shasta Dam quadrangle is composed of tan to brown well-bedded shale, sandy shale, sandstone, and some conglomerate. The rock is well cemented, and the shale breaks along bedding planes. The sandstone is locally feldspathic, and at these places contains angular to subangular fragments of feldspar. The conglomerate is made up of subangular to subrounded pebbles of shale, rhyolite, and greenstone as much as 6 inches across in a sandy matrix. Lenses of forebedded sandstone are found in the conglomeratic facies. The shale ranges in composition from coarse-bedded brown sandy shale to thin-bedded tan mudstone.

The contact between the Chico formation and the underlying Copley greenstone is exposed at several places in the Shasta Dam quadrangle. The Chico rests unconformably on all the older rocks, but no basal conglomerate is present. Most of the exposures of the base of the Chico in the Shasta Dam quadrangle show that shale of the Chico formation rests on soft, deeply weathered Copley greenstone.

The maximum dip of the Chico rocks in the Shasta Dam quadrangle is 15°. Dips range from 8° to 15°, averaging about 10° SE. This dip appears to be uniform throughout the small area where the Chico is exposed; no evidence was seen of local deformation.

No fossils were found in the Chico of the Shasta Dam quadrangle, but in this area the formation was mapped as Chico by Diller and it apparently correlates with part of the Upper Cretaceous Chico formation mapped by Diller (1906, p. 6) and Hinds (1933, p. 112–114) in nearby areas.

RED BLUFF FORMATION

Erosion remnants of the Red Bluff formation are along Clear Creek in the southeastern part of the Igo quadrangle and in the southeastern part of the Shasta Dam quadrangle. The formation is widely distributed in the north end of the Sacramento Valley east and south of the copper-zinc district. Anderson (1933, p. 238–239) reports that the coarser gravels are found near Redding at the head of the valley and that the deposit contains more sand near Red Bluff, 30 miles to the south. The Shasta copper-zinc district lies in the foothills from which part of the gravel and sand of the Red Bluff were derived, however, and the gravel and sand of the Red Bluff formation extended in embayments up canyons in these foothills and formed a veneer on stream-cut surfaces that slope gently up toward the foothills. Recent erosion has removed the Red Bluff from most of the canyons and has exhumed topography of an older cycle in most of the valley of the foothills.

The Red Bluff formation is composed principally of well-rounded boulders, many of which are elongate, flattened, oblate, or discoidal, set in an iron-stained matrix of sand and some clay. Fragments range in size from fine gravel to boulders several feet long; the average boulder is 6 to 12 inches across. The formation as a whole is poorly sorted, but lenses of well-bedded sand and sandy clay are common at some places. These lenses range from a few inches to 50 feet or more in length. Foreset beds are present in some of the sandy lenses.

The formation is well cemented in most parts of the district; river cliffs 100 feet high rise vertically, and except in the weathered zone much of the formation is hard and impervious. In other parts of the district, as at the Piety Hill (or Hardscrapple) mine (Logan, 1926, p. 186) about 1 mile northeast of Igo, the gravel is poorly cemented; and hydraulic methods were used to recover the gold at the base of the Red Bluff formation.

The formation is distinguished from the Recent gravel by its degree of cementation and iron staining and by its position which commonly is well above the present streams. In some canyons in the foothills, however, it is difficult to distinguish between the gravel beds of the Red Bluff formation and those that were formed on small terraces that were cut during the incising of the present streams. In such areas, the gravel beds are shown on the map with the symbol for the Red Bluff formation followed by a question mark. Anderson (1933, p. 238–239) reports that:

> The age of the Red Bluff formation is usually assigned to the Pleistocene although no paleontological evidence is at hand to substantiate this. The basis of the age determination appears to be the distinctive red color of the matrix as contrasted to the dominant grays of the Recent alluvial deposits and the fact that the Red Bluff formation is unconformable above the Tehama and Tuscan formations. Since the present streams have frequently cut their canyons as deep as 300 feet (Iron Canyon) below the Red Bluff, it does not seem desirable to place the Red Bluff in the Recent.

RECENT DEPOSITS

The main areas of Recent deposits are shown on the geologic map of the district (pl. 1). No attempt was made to delineate small areas of alluvium or shallow but extensive mantle soil and rock where sufficient outcrops were exposed to permit the mapping of the underlying rocks. The areas of soil and rock mantle and alluvium shown on the map are those that are extensive enough to mask bedrock geology and that make it impossible to extend geologic contacts across covered areas with any degree of certainty.

Three types of Recent deposits are distinguished on the geologic maps. The pattern for alluvium, Qal, includes residual soil and residual-debris mantle of several types that is essentially in place or has been moved only a few feet or tens of feet by such processes as solifluction. Material that has moved appreciably from its place of origin under the influence of gravity is distinguished from the mantle by the symbol Qls. This comprises landslides, slump, rockslides, debris flows, and mudflows. Deposits of sand and gravel in the present streams have the symbol Qg.

Little residual soil remains on the higher slopes, but in places these slopes are covered with rock debris or with a thin layer of soil and rock debris. Alluvium and surface mantle, for the most part, are limited to valleys and to the low foothills, particularly where the underlying rock is greenstone. In these areas, the mantle ranges from well-developed soil that is suitable for agriculture in the valleys to mixed soil and rock debris on hillsides that is suitable at some places for vineyards or orchards. Soil, except in alluvial valleys, is reddish and iron stained. Areas in the higher foothills and mountains that are marked Qal are covered with a mixture of coarse and fine rock debris that contains enough soil in the interstices of the rocks at many places to support a dense growth of chaparral. These areas are particularly common on north- or east-facing slopes. Little or no soil has formed where the bedrock is rhyolite, but north-facing slopes even on rhyolite will support a brush cover. An exception to the lack of soil on rhyolite and the lack of soil at higher altitudes occurs on the coarse-phenocryst Balaklala rhyolite. Here considerable areas of mixed soil and rock debris are formed, partly because the coarse-phenocryst rhyolite weathers more easily than other varieties of rhyolite and partly because some of this rock is at or near the altitude of the old Klamath surface, described under "Physiographic features," and was deeply weathered during an earlier cycle.

Landslides, and earth and rock masses that have moved by flowage, cover considerable areas; the larger slides measure several thousand feet across. These larger slides are true landslides in the upper part, but the middle and lower parts have moved largely as mudflows or earthflows. Therefore the upper part may be composed of large slide blocks that are only slightly displaced from each other, while the lower part is a jumble of soil and rock debris. The Balaklala Angle Station gossan landslide apparently is an example where both types of movement have occurred (p. 107). Some of the old landslides have not been reactivated recently, but others, for example, the landslide at the head of Motion Creek, which was active in 1950–51, dam the streams and form mudflows that destroy roads and bridges. Even old landslides that appear to be stabilized move to some extent during the rainy season. The old landslide on which the mine plant of the Mountain Copper Co. was built many years ago has moved enough to crack some of the concrete building foundations.

Recent sand and gravels, Qg, were of commercial importance to the district as a source of placer gold. Nearly all the deposits of sand and gravel in large and small streams have been placered at least once in the search for gold, and along streams such as Whisky Creek, much of the gravel was neatly stacked by the miners by hand in order to reach bedrock. The channels of the Sacramento River and Clear Creek have been dredged along parts of their courses. The alluvium in the present streams ranges from deposits that contain boulders 10 feet across, as along Brandy Creek or Eagle Creek, to broad sand bars that occur along parts of Clear Creek and the Sacramento River. Most of the deposits of sand and gravel in the West Shasta district are too small or too difficult to reach to be of interest as commercial sources of sand and gravel.

INTRUSIVE ROCKS

MULE MOUNTAIN STOCK

The Mule Mountain stock (Hinds, 1933, p. 105), which consists of trondhjemite and albite granite, and some quartz diorite, underlies most of the southwestern quarter of the Shasta Dam quadrangle and the southeastern quarter of the Whiskytown quadrangle. This stock includes the hornblende-quartz diorite of Diller (1906, p. 8) and part of the alaskite porphyry of Graton (1909, p. 81). The stock has an oval-shaped outcrop pattern that is 10 miles long from north to south and is 5 miles wide (pl. 1). The outcrop pattern centers about the town of Shasta, and the stock includes Mule and Democrat Mountains. Good exposures of the stock may be seen in the road cuts along U. S. Highway 299W near the town of Shasta, along the Iron Mountain road near Spring Creek, and in Clear Creek canyon west of Mule Mountain. Dikes of albite granite are common in the Copley greenstone near the contact with the stock.

The Mule Mountain stock is deeply weathered, and road cuts more than 50 feet high near Shasta are entirely in white crumbly, disintegrated albite granite. Unweathered albite granite is exposed in some of the deep canyons, for example, Clear Creek and Spring Creek canyons.

The stock is heterogeneous; it consists of a generally leucocratic, holocrystalline granitoid rock that ranges in composition from albite granite to hornblende-quartz diorite, but is here generally referred to as albite granite. Essential minerals are quartz, plagioclase, and epidote.

Most of the stock is a granitoid rock of the composition of trondhjemite, and is a leucocratic soda-rich quartz diorite that contains sodic oligoclase and quartz, and 10 to 15 percent interstitial epidote and chlorite, and that has an average grain size of about 2 millimeters. This phase is considered to be the composition of the original, unaltered intrusive.

In places the rock has been extremely reworked by late magmatic solutions and is an albite granite. The altered rock has a pseudoporphyritic texture containing quartz "phenocrysts" 4–6 millimeters in diameter in a white groundmass of albite and quartz that averages 2 millimeters in grain size and that contains 5 to 10 percent epidote in clusters from 2 to 6 millimeters in diameter. Large parts of the Mule Mountain stock were altered by late magmatic solutions that added silica and soda. The altered parts of the stock have gradational contacts to the unaltered or slightly altered trondhjemite. The principal bodies of altered rock are outlined on the geologic map (pl. 1). Quartz "phenocrysts" are limited mainly to the altered areas, although some of the trondhjemite also contains added quartz as pseudophenocrysts.

Chemically and mineralogically the equigranular trondhjemite and the white pseudoporphyritic albite granite are similar. They are composed mainly of quartz, albitic plagioclase, epidote, and chlorite. The commonly equigranular trondhjemite contains slightly more CaO, FeO, and Fe_2O_3 and slightly less Na_2O and SiO_2 than the porphyritic albite granite (table 4). In the field the difference is more apparent, as the pseudoporphyritic albite granite is whiter and looks more siliceous than the equigranular trondhjemite. The epidote is finely disseminated throughout the equigranular trondhjemite, but is regrouped in clusters ranging from 2 to 6 millimeters in diameter in the albite granite.

TABLE 4.—*Analyses of albite granite and quartz diorite from the Mule Mountain stock*

[Samples 1, 4, 5, by U. S. Geol. Survey rapid-analysis method. Analysts, S. M. Berthold and E. A. Nygaard. Samples 2, 3, 6, analyst, M. K. Carron]

	1	2	3	4	5	6
SiO_2	67 6	72.74	73.48	75.6	75.8	76.48
Al_2O_3	14.8	14.14	13.63	13.4	13.9	13.09
Fe_2O_3	1.6	1.26	2.08	2.0	.92	.64
FeO	3.2	1.82	.73	.52	.44	.79
CaO	5.1	2.12	2.60	1.2	1.9	1.70
MgO	2.1	.63	.60	.79	1.2	.90
K_2O	.86	.75	.51	1.2	.43	.51
Na_2O	3.2	4.12	3.78	3.5	4.4	4.46
H_2O-		.13	.74			.22
H_2O+		1.53	1.83			1.18
TiO_2	.36	.26	.28	.22	.20	.25
MnO	.10	.05	.02	.02	.02	.02
P_2O_5	.03			.02	.01	
Ignition loss [1]	1.3			2.2	1.4	
	101	99.55	100.28	101	101	100.24

Norms

[Computed by the CIPW system]

	1	2	3	4	5	6
Quartz	30.18	39.66	43.08	46.68	42.48	43.40
Orthoclase	5.00	4.45	2.78	7.23	2.78	2.78
Albite	27.25	34.48	31.96	29.87	37.20	37.72
Anorthite	23.35	10.56	10.79	5.84	9.45	8.34
Corundum		2.75	2.24	4.69	2.75	2.14
Diopside	1.44					
Hypersthene	8.63	3.58	2.11	2.64	3.96	3.01
Magnetite	2.32	1.96	2.78	1.16	.93	.93
Ilmenite	.61	.46	.47	.46	.30	.46
Hematite				1.28	.32	
	98.78	97.90	96.21	99.85	100.17	98.78

[1] Includes gain due to oxidation of FeO.

1. Hornblende-quartz diorite, sec. 36, T. 33 N., R. 6 W.
2-3. Trondhjemite from Whiskytown quadrangle, California.
4. "Porphyritic" albite granite with epidote from sec. 6, T. 32 N., R. 5 W.
5. Siliceous, "porphyritic" albite granite from Highway 299W, 0.25 mile east of branch road to South Fork Lookout station.
6. Siliceous, "porphyritic" albite granite from Whiskytown quadrangle, California.

Intrusive breccias are abundant in the northeastern part of the stock (fig. 23). They formed where the albite granite of the Mule Mountain stock shattered and partly absorbed Copley greenstone. The fragments of greenstone range in size from tiny relicts to as much as 7 feet in diameter. Many are blocky and irregular in outline and look like they were formed by breaking up

FIGURE 23.—Intrusive breccia near Spring Creek. Reworked Copley greenstone fragments in albite granite matrix.

of an originally massive rock, but possibly 25 percent of the fragments are elongate and oriented, and were derived from foliated greenstone. Some are corroded until only wisps of the original fragment can be seen.

There is some correlation between the size of a fragment and the amount of recrystallization and granitic material that has been added. Many large fragments are little-altered Copley greenstone that show amygdaloidal and pillow structures. Smaller fragments are reworked to a granular rock containing feldspar, hornblende, and quartz porphyroblasts as much as 1 centimeter in diameter in a fine-grained groundmass. In Spring Creek the fragments of reworked greenstone average about 3 by 4 inches across, and hornblende porphyroblasts 1 to 2 millimeters in diameter have been formed in a fine-grained groundmass that has a granoblastic texture.

The matrix between the fragments in the intrusive breccia is coarse-grained pseudoporphyritic albite granite. Where the fragments are small and have been extensively reworked, the matrix of albite granite has assimilated the greenstone and is more basic and contains unaltered hornblende crystals. The rock in the resistant knob of hornblende-quartz diorite that lies on the west edge of the intrusive breccia area on the Iron Mountain road, where the road crosses Spring Creek, is an example of contamination by absorbed material from the greenstone. The knob is a hard massive dark-gray granitoid hornblende-quartz diorite. It contains about 2 percent dark fine-grained xenoliths. These xenoliths have euhedral porphyroblasts of hornblende 1 to 2 millimeters long that are identical with the hornblende in the enclosing massive hornblende-quart diorite. Throughout the hornblende-quartz diorite are many tiny areas of mafic minerals that have a texture similar to that of the xenoliths, and are the remnants of former fragments. The hornblende in the intrusive matrix was derived from assimilation of basic xenoliths, and the resultant rock is more basic than the original magma. This knob of hornblende-quartz diorite, which has the appearance of being the least altered variety of the Mule Mountain stock, is actually more basic than the original magma because of contamination by reworking and incorporation of much greenstone.

The Mule Mountain stock is cut by many veins of quartz and some dikes of aplite. The quartz veins range in thickness from 2 inches to several feet. They contain small amounts of pyrite, chalcopyrite, and calcite, and some are auriferous and have been mined for their gold content. Aplite dikes are common in the trondhjemite and albite granite on Mule Mountain and in the greenstone near the contact with albite granite. They are commonly less than 1 foot thick and have sharp walls. Some of the aplite in dikes has quartz phenocrysts and resembles Balaklala rhyolite; the aplite dikes can be distinguished by their sugary groundmass.

Inclusions.—Porphyritic Balaklala rhyolite and greenstone inclusions locally are abundant in the Mule Mountain stock. Inclusions of Balaklala rhyolite are inconspicuous as they are the same color and have the same mineralogy as the trondhjemite and albite granite. However, along the western and northwestern contact between the Mule Mountain stock and Balaklala rhyolite many inclusions of porphyritic rhyolite in albite granite can be recognized by their aphanitic groundmass, whereas the albite granite has a granitoid groundmass. Greenstone inclusions are more abundant than those of rhyolite. In addition to the intrusive breccia described above, many large inclusions of schistose Copley greenstone are exposed in road cuts on U. S. Highway 299W near Shasta and along the road to the Forest Service lookout station on South Fork Mountain. Some of these inclusions are several hundred feet long and only 6 to 8 feet thick; at first glance they resemble mafic dikes, but amygdules and pillow structures are preserved in the larger inclusions, some of which are probably roof pendants.

The Mule Mountain stock generally has a sharp contact with the Copley greenstone and the Balaklala rhyolite. Balaklala rhyolite is unaltered where it is in contact with trondhjemite or albite granite because of the similarity in mineralogic character and chemical composition; the contact metamorphic halo in the greenstone surrounding the intrusive is more pronounced. Near Matheson, and on the west side of Slick Rock Creek south of Iron Mountain, the greenstone has been metamorphosed for several hundred feet from the contact of albite granite. Within 50 feet of the intrusive the chloritic Copley greenstone is recrystallized to a fine-grained rock containing mainly plagioclase and hornblende. This grades outward into greenstone with lumps and knots of epidote and finally into a foliated chloritic greenstone. Within the metamorphic halo the foliated character of the greenstone is destroyed.

The contact of the Mule Mountain stock is generally parallel to the foliation in the intruded rock, but locally cuts across foliation. Apparently the Copley greenstone was already foliated when the albite granite was intruded. However, some diastrophism occurred after the intrusion of albite granite, as its borders are locally sheared parallel to the regional foliation. Therefore the intrusion is believed to be syntectonic with the Nevadan orogeny. The albite granite is older than the Shasta Bally batholith, as it is intruded northeast of Igo by the batholith (pl. 1).

PETROGRAPHIC DESCRIPTION

Trondhjemite.—About 65 percent of the Mule Mountain stock consists of a hypidiomorphic granitoid rock composed of quartz and plagioclase, and 10 to 15 percent mafic minerals that consist only of green biotite, epidote, and chlorite. Accessory minerals are apatite, sphene, and magnetite. Hornblende was not observed in this rock. Much of the rock is equigranular and averages 2 millimeters in grain size, but in places rounded quartz masses as much as 8 or 9 millimeters in diameter that resemble phenocrysts are abundant. The distribution of the mafic minerals is a characteristic feature of the rock. These minerals are very fine grained and are interstitial to plagioclase, being present in thin stringers and lenses that wrap around the plagioclase.

The rock is composed of 60 to 65 percent plagioclase, which is subhedral and shows both Carlsbad and albite twinning. The plagioclase is strongly altered to sericite, and a few cores are saussuritized. Most of the plagioclase is sodic oligoclase ranging in composition from $Ab_{90}An_{10}$ to $Ab_{86}An_{14}$. Zoning is inconspicuous except in a few grains that have a slightly more calcic core.

The mafic minerals include epidote, chlorite, and green biotite listed in decreasing order of abundance. Epidote is present in saussuritic cores of plagioclase and as an interstitial mineral associated with chlorite and green biotite. Chlorite occurs in thin interstitial veinlets; it has a distinct anomalous blue interference color. Green biotite is present in very small quantities; it is a mineral of late magmatic origin and is associated with sericite in fractures in quartz.

Quartz, a late mineral, occurs as anhedral grains that average about 2 millimeters in diameter, and constitutes about 20 to 30 percent of the rock. It replaces plagioclase and fills the spaces between these grains, but also locally forms pseudophenocrysts. These quartz pseudophenocrysts are aggregates of quartz grains that replace the plagioclase (fig. 24).

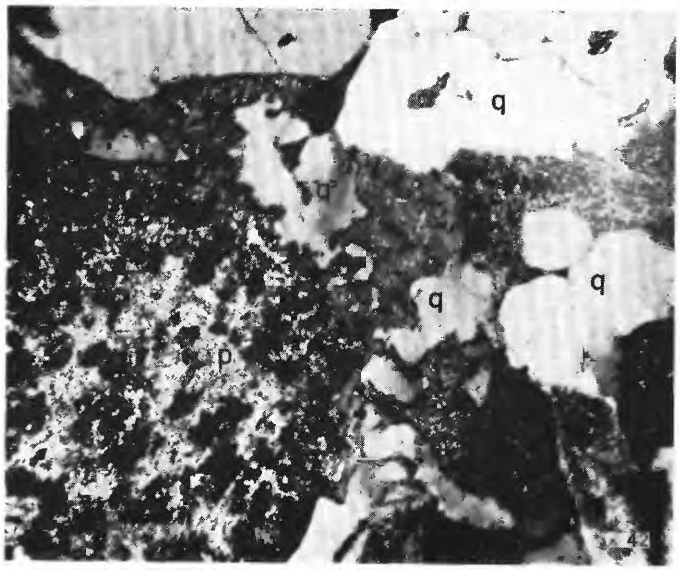

FIGURE 24.—Photomicrograph of albite granite showing early stage in the formation of a pseudophenocryst of quartz (*q*) that is interpreted as having replaced plagioclase (*p*).

"*Porphyritic*" *albite granite*.—This granite is a white, hypidiomorphic granular rock containing rounded quartz pseudophenocrysts 6 to 10 millimeters in diameter in a granitoid groundmass that averages 1½ millimeters in grain size. It is similar to the albite granite described by Gilluly at Sparta, Oreg. (Gilluly, 1933) and to the albite granites associated with spilitic suites described by Dewey and Flett (1911, p. 202–209, 241–248), and by Harwood and Wade (1909, p. 549–554). The rock contains mainly quartz and albite and less than 10 percent epidote and chlorite. Epidote occurs in clusters commonly 2–6 millimeters in diameter in contrast to the thin veinlets of epidote in the largely equigranular trondhjemite.

The "porphyritic" albite granite constitutes about one-third of the Mule Mountain stock. It appears as large irregular bodies throughout the trondhjemite mass where late magmatic solutions rich in soda and silica replaced the original trondhjemite. Contacts between the two types of rocks are gradational.

Under the microscope the rock is seen to contain mainly quartz, albite, and epidote, and small amounts of chlorite, zoisite, biotite, sericite, apatite, and opaque minerals. Quartz and albite make up more than 90 percent of the rock. Albite is in subhedral grains and is extremely corroded by quartz. Albite ranges in composition from $Ab_{96}An_4$ to $Ab_{89}An_{11}$. The plagioclase rarely contains any epidote or clinozoisite, but some of the albite is strongly altered to sericite.

Quartz is one of the last minerals to crystallize. It is interstitial to albite, and where the large pseudophenocrysts formed, it corroded the albite as shown by many relicts of albite through the quartz. Both quartz and albite show strong cataclastic structures. Mortar structure is common, and all the quartz has undulatory extinction. The quartz is full of tiny rounded fluid inclusions.

Large areas of myrmekite replace grains of quartz and albite in albite granite. Several distinct intergrowth textures are present (Sederholm, 1916; Gilluly, 1933, p. 71–72; and Alling, 1936, p. 163–172). The most common pattern has tiny, corroded, hook-shaped masses of albite in a sea of clear quartz. The quartz and albite are in optical continuity over areas of several square millimeters. The intergrowth seems to replace earlier quartz and albite and is interpreted as a simultaneous recrystallization caused by late magmatic solutions. Another less common intergrowth texture has slender irregular rods of albite in quartz, and both minerals are in optical continuity. The rods of albite grade into unreplaced albite crystals, and are oriented parallel to cleavage planes of the unreplaced albite. This texture is due to replacement of albite by quartz along cleavage planes. The third type of intergrowth is myrmekite. The edges of the myrmekitic intergrowths are coarser than the centers and grade into a normal granitoid texture. Albite in the myrmekite has the composition $Ab_{96}An_4$, whereas the subhedral albite crystals have a composition about $Ab_{90}An_{10}$. Apparently the last hydrothermal solutions were the most sodic and formed the intergrowth textures.

The mafic minerals include green biotite, epidote, and chlorite. Biotite is interstitial to quartz and albite, and is in clusters of small plumose growths intergrown with sericite. The biotite is either a late primary mineral or an early hydrothermal mineral. Epidote is the principal mafic mineral. It is found in clusters as much as 6 millimeters in diameter that are made up of individual crystals as much as 2 millimeters in length. The epidote is much coarser than in the equigranular trondhjemite and only in a few slides were tiny epidote grains seen in the plagioclase crystals. Evidently the epidote was recrystallized by deuteric solutions into these coarsely crystalline knots. The epidote content was decreased from about 20 percent in the equigranular trondhjemite to less than 10 percent in the silicified "porphyritic" albite granite.

Light green chlorite commonly surrounds epidote and thins out from epidote clusters into veinlets along grain boundaries of quartz and albite. Inclusions of opaque minerals that are altered to leucoxene are abundant in the chlorite. Euhedral apatite and sphene crystals occur in minor quantities as accessory minerals.

Hornblende-quartz diorite.—This rock is localized in a small body several hundred feet in diameter in the north end of the Mule Mountain stock in Spring Creek at the crossing of the Iron Mountain road. The hornblende-quartz diorite grades to the north and east into an intrusive breccia in Copley greenstone and to the south and west into trondhjemite. It is a dark gray hard equigranular granitoid rock, formed by assimilation of greenstone by trondhjemite. The average grain size is 2 millimeters. The rock consists mainly of quartz, plagioclase, hornblende, chlorite, and epidote and some sphene, apatite, and magnetite. Fine-grained dark-gray xenoliths of altered greenstone make up about 2 percent of the body. They range in size from tiny wisps to ellipsoidal or irregular fragments as much as 6 inches in diameter. The xenoliths have hazy borders with the hornblende-quartz diorite.

Subhedral grains of plagioclase that are extremely altered to epidote, clinozoisite, and sericite constitute about 60 percent of the rock. Albite twinning is common but is inconspicuous owing to saussuritic alteration. Thin rims of clear albite-oligoclase on the altered plagioclase are common. The original plagioclase was probably much more calcic, but no relicts of it were observed in the saussurite.

The hornblende-quartz diorite contains about 25 percent quartz. Quartz is present as anhedral aclastic grains that are interstitial to plagioclase. Quartz has corroded plagioclase as shown by many irregular relicts of plagioclase (fig. 25).

Hornblende and chlorite make up about 10 percent of the rock. Hornblende is in subhedral grains that are pleochroic from light to dark green. Chlorite and epidote are pseudomorphic after hornblende.

Aplite.—Aplite dikes that have sharp walls and are usually less than a foot thick cut the albite granite on Mule Mountain, and similar dikes cut the Copley greenstone near the albite granite contact. Some of the aplite contains euhedral quartz phenocrysts, commonly 2 millimeters in diameter, in a fine-grained sugary groundmass of quartz, albite, minor chlorite, sphene, epidote, and leucoxene. The groundmass of the aplite has a microgranitoid texture. Myrmekitic and micrographic intergrowths of albite and quartz are abundant in the aplite.

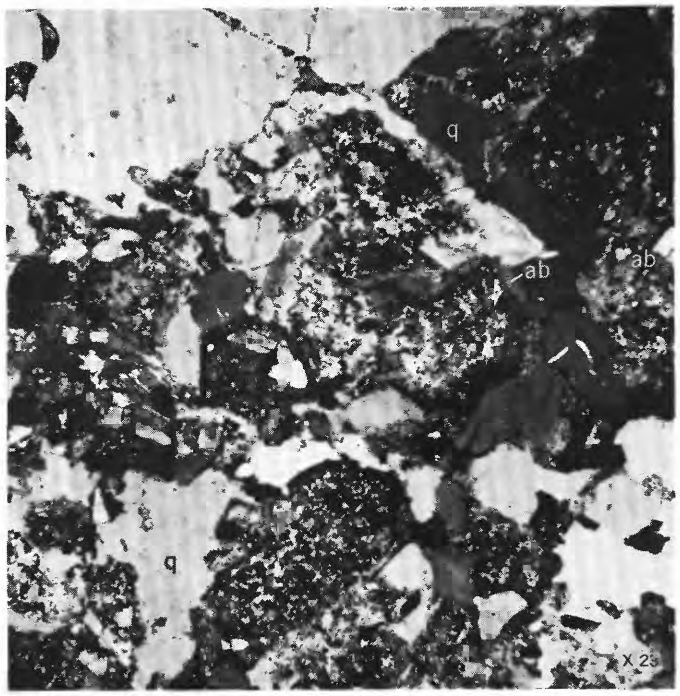

FIGURE 25.—Photomicrograph of hornblende-quartz diorite, a varietal type of the Mule Mountain stock. Plagioclase has cores with strong saussuritic alteration and some clear rims of albite (*ab*). Some quartz (*q*) has replaced the plagioclase. Crossed nicols, ×14.

AGE

The Mule Mountain stock is dated as Late Jurassic in age, although the youngest rocks that are intruded by the Mule Mountain stock are the Balaklala rhyolite of Middle Devonian age. Diller correlates the Mule Mountain stock with a small plug of albite granite (hornblende-quartz diorite of Diller) that intrudes Bully Hill rhyolite of Triassic age in the Redding quadrangle 1½ miles southeast of Oak Run and 6 miles southeast of the Afterthought mine (Diller, 1906, p. 8). The Mule Mountain stock, in turn, is intruded by satellitic bodies of the Shasta Bally batholith in the southeastern part of the Igo quadrangle (pl. 1).

The Mule Mountain stock was syntectonic with the Nevadan orogeny of Late Jurassic or Early Cretaceous age. This stock was intruded after the orogeny foliated and regionally metamorphosed the Copley greenstone, but early enough for the later phases to affect the albite granite of the stock itself. The albite granite is elongated with the regional foliation and its elongate shape was apparently determined by regional structures formed during orogeny, but massive albite granite in some places cuts directly across foliated Copley greenstone. The albite granite locally metamorphosed the Copley greenstone to amphibolite and epidote amphibolite, and if the latter rocks had been present during the main part of the orogeny, they should have been reduced by retrograde metamorphism to chlorite-albite schists similar to the mineralogy of the Copley elsewhere in the district. The albite granite is itself locally sheared or sheeted parallel to its contact; the writers believe this is due to movements in the closing phases of the Nevadan orogeny, as the areas of foliation near the borders of the stock are too widespread to be accounted for by post-orogenic local movements.

SHASTA BALLY BATHOLITH

DISTRIBUTION AND GENERAL FEATURES

The Shasta Bally batholith is the largest pluton in the Redding area, and it underlies most of the western half of the Igo quadrangle. Several small satellitic bodies crop out to the east between the Shasta Bally batholith and the Mule Mountain stock (pl. 1). Regionally the batholith crops out for a distance of 30 miles in a N. 20° W. direction, mostly beyond the mapped area, and has a maximum width of 10 miles.

Most of the batholith is a biotite-hornblende-quartz diorite, but it ranges in composition near the edge and in satellite bodies from gabbro to granodiorite. The color ranges from dark gray in the gabbro to nearly white or light gray in the granodiorite. The batholith is a single intrusive; there is no evidence of multiple intrusion of the rock types that differ slightly from each other. The rock is here called a biotite-quartz diorite even though hornblende is a common constituent, as hornblende is rarely visible megascopically in the main body of the intrusive. The principal minerals are quartz, feldspar, biotite, and hornblende. The rock has a granitoid texture and the average grain size is about 2 millimeters. Porphyritic textures are rare, in contrast to the prominent pseudoporphyritic textures of the Mule Mountain stock.

The main part of the batholith intrudes Copley greenstone in the mapped area, but the satellitic plug along Clear Creek in the southern part of the Igo quadrangle intrudes the Mule Mountain stock of albite granite also. In the southern part of this quadrangle the border of the batholith is darker than the interior, and much of the border phase is hornblende diorite. It has a sharp contact with Copley greenstone, which is recrystallized to a fine-grained amphibolite, but the hornblende diorite grades into biotite-quartz diorite toward the interior. In the northern part of the Igo quadrangle near Brandy Creek the contact between granitic rock and greenstone is less distinct; massive quartz diorite, which is the predominant rock along this part of the contact, grades outward to a migmatite that has interlayered dark and light bands. The migmatite zone is described under "Igneous metamorphism."

The border of the batholith commonly has a planar structure due to primary foliation parallel to the contact and at a few localities alined hornblende gives the rock a linear element, but the interior is mostly massive. The foliated quartz diorite is slightly coarser grained than the interior of the batholith and contains more mafic minerals, averaging about 40 percent hornblende and biotite.

The batholith was intruded in the form of a large schlieren arch, and its top was probably only a short distance above the peaks from Shasta Bally to Grouse Mountain. Primary foliation, shown by alinement of minerals and inclusions and by segregation of minerals in bands of different composition, is limited to the borders of the batholith and to the peaks and high ridges in the central part. In the east part of the batholith the primary foliation strikes N. 20° W. parallel to the contact and dips 60° to 70° E. Westward toward the center the foliation becomes flatter until it is nearly horizontal on the peaks between Shasta Bally and Grouse Mountain, which are probably near the top of the intrusive. On the west side of the intrusive, outside of the mapped area, the foliation dips steeply west.

Inclusions.—Near the border of the intrusive, sporadic small dark spindle- to pancake-shaped amphibolitic inclusions 1 to 3 inches in diameter and one-half inch thick are found. The inclusions are alined parallel to the primary foliation of the quartz diorite. They are much finer grained than the quartz diorite in which they are enclosed and have a crystalloblastic texture. These inclusions are probably reworked fragments from the Copley greenstone.

PETROGRAPHIC DESCRIPTION

Although the Shasta Bally batholith and the small satellite bodies of granitic rock are all part of one intrusive, which is uniform in composition throughout most of the body, several varietal types have been recognized in the field and in thin sections. The varietal types, particularly those in the Clear Creek plug, are described in detail as some of them show the effect on the quartz diorite magma of assimilation of greenstone. The varietal types are granodiorite, hornblende diorite, and gabbro; these are limited to the border of the batholith or to satellite plugs.

Biotite-quartz diorite.—Biotite-quartz diorite constitutes about 90 percent of the batholith. It is an equigranular rock averaging about 2 millimeters in grain size. Essential minerals are quartz, plagioclase, biotite, and hornblende, and less common constituents and accessory minerals include orthoclase, epidote, chlorite, kaolinite, apatite, magnetite, zircon, and allanite.

Plagioclase constitutes 45 to 50 percent of the rock. It occurs as subhedral chunky grains that have prominent albite twinning and commonly show Carlsbad and pericline twinning. Zoning is prominent, and cores are commonly $Ab_{68}An_{32}$ to $Ab_{60}An_{40}$ and rims $Ab_{75}An_{25}$ to $Ab_{70}An_{30}$.

Most of the quartz diorite contains 20 to 25 percent quartz as anhedral grains that are commonly interstitial to plagioclase. Hornblende and biotite constitute 15 to 25 percent of the rock; normally biotite is slightly more abundant than hornblende. Hornblende forms euhedral crystals that are pleochroic from light to dark green. Anhedral grains of biotite have the following pleochroism: X=colorless; Y=Z=dark-brown to reddish-brown; it is in part altered to chlorite.

A maximum of 5 percent orthoclase may occur as tiny, anhedral interstitial grains. It is more abundant in the quartz diorite in which biotite is the predominant mafic mineral. Common accessory minerals are sphene, apatite, zircon, and magnetite. They are most common as euhedral inclusions in biotite. Allanite is found rarely as an accessory mineral.

Granodiorite.—Small areas of granodiorite that crop out in the southern part of the Igo quadrangle are probably small intrusive bodies that are differentiates of the quartz diorite, but the contact relationship is not exposed. The granodiorite is a light-gray fine-grained rock that contains mainly quartz, orthoclase, and plagioclase and less than 10 percent biotite and 1 percent hornblende. Apatite, sphene, zircon, and magnetite are common accessory minerals.

The granodiorite contains 45–60 percent plagioclase and 10–15 percent orthoclase. Twinning and zoning of the plagioclase are similar to those of the quartz diorite. Plagioclase has cores that range in composition from $Ab_{72}An_{28}$ to $Ab_{63}An_{37}$ and rims of $Ab_{75}An_{25}$. Orthoclase is usually interstitial to plagioclase. Both microperthite and antiperthite are present. In the microperthite thin stringers of oligoclase are included in orthoclase and probably formed by exsolution. In the antiperthite, cores of plagioclase are nearly half replaced by thin veinlets, and a ramifying network, of orthoclase stringers whereas the rims contain no orthoclase.

Myrmekite containing tiny stringers and hook-shaped forms of quartz in orthoclase, although not observed in the quartz diorite, is common in the granodiorite.

Hornblende diorite.—This rock is concentrated near the borders of the intrusive and in the small satellitic bodies. It differs from the quartz diorite mainly in that it contains a more calcic plagioclase, contains less than 5 percent quartz, and contains more hornblende than biotite.

The hornblende diorite is an equigranular, medium coarse-grained, dark-gray rock that contains mainly plagioclase and hornblende and small amounts of augite, orthoclase, quartz, epidote, sphene, magnetite, apatite, and zircon. Plagioclase constitutes 50 to 60 percent of the rock and is andesine-labradorite that is zoned from cores of $Ab_{48}An_{52}$ to rims of $Ab_{58}An_{42}$. Albite twinning is prominent while pericline and Carlsbad twinning are less common. Hornblende, as euhedral crystals 3 to 4 millimeters in diameter that poikilitically enclose plagioclase, magnetite, and apatite constitutes as much as 40 percent of the rock. The hornblende is strongly pleochroic from light to dark green. Augite forms the cores of a few hornblende crystals. In places quartz and orthoclase are present in small amounts as interstitial small grains. Sphene, apatite, zircon, and magnetite are common accessory minerals.

The hornblende diorite and the quartz diorite probably are part of a single intrusive. This intrusive assimilated part of the Copley greenstone as it was intruded, reworking the chloritic Copley greenstone to hornblende and plagioclase.

Gabbro.—Coarse-grained hornblende gabbro occurs as small dikes and irregular bodies near the margin and in the roof rock of the Clear Creek plug in the Igo quadrangle, which is a satellitic body of the Shasta Bally batholith. The biotite-quartz diorite in the interior of the Clear Creek plug, as locally in the Shasta Bally batholith, grades outward through hornblende-quartz diorite to hornblende diorite near the border. The contact of the hornblende diorite with the albite granite is sharp, whereas the contact with Copley greenstone is gradational and the gradation extends into gabbro that was formed in situ by metasomatism of the Copley greenstone. This metasomatism is described under "Metamorphism related to the Shasta Bally batholith."

AGE

The Shasta Bally batholith is Late Jurassic or Early Cretaceous in age. It was not affected by the Nevadan orogeny and cuts directly across foliation formed during the Nevadan orogeny of Late Jurassic or Early Cretaceous age. Also the Shasta Bally batholith has metamorphosed the Copley greenstone to amphibolite, gneiss, and migmatite, but no retrograde metamorphism has been superposed on the altered zones by later orogeny.

The Shasta Bally batholith is overlain nonconformably by beds of the middle part of the Paskenta formation of Anderson (1902) and Horsetown formation of Lower Cretaceous age west of the West Shasta district (Hinds, 1934, p. 182–192). Hinds states that

> Lower Cretaceous fossils present at many horizons show that the strata in contact with the [Shasta Bally] batholith range from Middle Paskenta (Valanginian) on the western side to Lower Horsetown (Hauterivian) on the eastern according to information furnished me by Dr. F. M. Anderson [(1933)]

On the basis of geologic and paleontologic evidence the intrusion of the Shasta Bally batholith thus occurred after the Nevadan orogeny of Jurassic age but before the deposition of the Lower Cretaceous strata which nonconformably overlie the batholith.

A 50-pound sample of the Shasta Bally batholith was collected for age determination by the Larsen method. David Gottfried of the U. S. Geological Survey laboratory (1954, oral communication) reports the age of the intrusion as 97 million years.

MINOR INTRUSIVE BODIES

Small intrusive bodies are abundant in the Copley greenstone and the Balaklala rhyolite, and to a lesser extent in the Kennett and Bragdon formations and in the Shasta Bally and Mule Mountain plutons. These minor intrusions were emplaced before and after the Nevadan orogeny. The intrusions of pre-orogenic age include amphibolite, diabase, and nonporphyritic and porphyritic rhyolite; the younger intrusions include lamprophyre, andesite, diorite porphyry, dacite porphyry, quartz latite porphyry, and aplite.

ROCKS INTRUDED BEFORE THE NEVADAN OROGENY

Rhyolite.—Minor intrusives are preponderant in small stocks, plugs, dikes, and sills of Balaklala-type rhyolite in Copley greenstone and Balaklala rhyolite. Many small irregular stocks, plugs, and dikes of porphyritic and nonporphyritic rhyolite that were feeders for the flows of Balaklala age cut the underlying Copley and the older parts of the Balaklala. Other small rhyolitic dikes and sills cut the Copley throughout the mapped area. However, rhyolitic flows also occur in the Copley near the top of the formation, and Balaklala rhyolite is infolded in the Copley. In the absence of bedding, it is difficult to differentiate the form in which the rhyolite occurs, and all bodies of rhyolite that are lithologically identical with Balaklala rhyolite are given that symbol. Bodies of intrusive rhyolite, being largely feeders for the overlying flows, are described with the Balaklala rhyolite.

Hornblendite.—A few hornblendite and serpentinized hornblendic dikes or sills are in the Copley in the Igo quadrangle and in the southern part of the Shasta Dam quadrangle. They range in thickness from a fraction of an inch to several hundred feet.

Some hornblendite dikes are in the southern part of the Shasta Dam and Whiskytown quadrangles. They are dark-green porphyritic mafic dikes 50–60 feet thick that have phenocrysts of hornblende and augite as much as 1 centimeter in diameter in an aphanitic chloritic matrix. Under the microscope they are seen to contain mainly hornblende and small amounts of augite, olivine, and opaque minerals. The mafic minerals are in part altered to epidote, antigorite, and talc. The dikes or sills have chilled borders against Copley greenstone, and some are cut by lamprophyre dikes.

Diabase.—Dikes and sills of diabase cut all rocks in the district older than the Shasta Bally batholith. The intrusive bodies range from a few feet to about 200 feet in thickness and are as much as 5,600 in length. They are fine-grained massive dark-gray to greenish-gray rocks that are locally finely porphyritic. Phenocrysts where unaltered are plagioclase and hornblende, but commonly these are altered to epidote and chlorite.

The diabases are all altered to some extent, although megascopically they look like unaltered rocks. They are composed of plagioclase, hornblende, epidote, chlorite, calcite, quartz, biotite, and magnetite and have a relict ophitic texture. Plagioclase ranges in composition from albite ($Ab_{96}An_4$) to labradorite ($Ab_{41}An_{59}$). Where unaltered, it is labradorite and occurs as clear lath-shaped crystals that have prominent albite twinning in a groundmass of chlorite and epidote. Most of the plagioclase has saussuritic alteration and is cloudy, massive albite containing inclusions of epidote. Mafic minerals in the diabase are mainly chlorite and epidote and have some relict cores of hornblende and minor secondary biotite. Augite, which was probably originally present, is completely altered. Quartz is present in minor quantities as tiny interstitial grains. Most of the quartz appears to be primary, although some may be released silica. Calcite is in irregular patches and veinlets in plagioclase and filling fractures in the rock.

A few diabase intrusive bodies associated with keratophyre in the central part of the district are albite diabase.

The diabase may be of several ages. It has all undergone some alteration, probably during regional metamorphism. The albite diabase, which is intrusive into keratophyres of the Copley greenstone, is probably of Devonian age, although the evidence is not conclusive. It may be related to an albitic suite of spilite, keratophyre, and soda rhyolite that is of Devonian age.

Some diabase dikes and sills cut the Kennett and Bragdon formations and are post-Mississippian in age. They appear to be similar to the diabase dikes cutting the Copley in many places in the eastern part of the Shasta Dam quadrangle.

The dikes occur as sharp-walled straight bodies that maintain fairly uniform widths in competent rocks, but they are less regular in shale. Figure 26 shows the mashing that accompanies intrusion at some places in the shale.

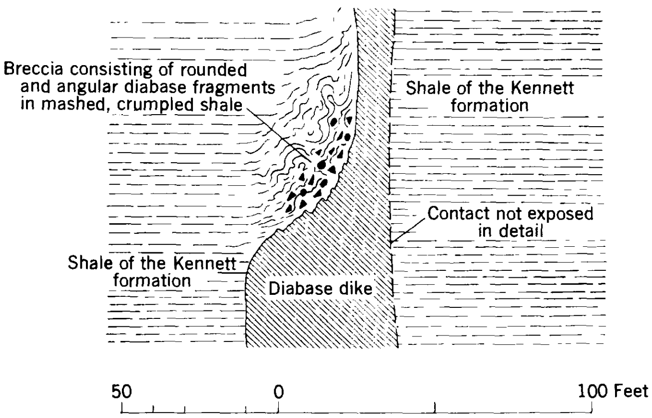

FIGURE 26.—Sketch of intrusive breccia along diabase dike in shale of the Kennett formation.

ROCKS INTRUDED AFTER THE NEVADAN OROGENY

Most of the minor post-Nevadan intrusive bodies are felsic dikes and sills related to the Shasta Bally batholith and lamprophyre dikes that cut the Mule Mountain stock of Jurassic(?) age. The felsic dikes include diorite porphyry, dacite porphyry, quartz latite porphyry, and aplite. Two groups of diorite porphyry and dacite porphyry dikes and sills are recognized in the field. The first group has large, conspicuous, white, zoned plagioclase phenocrysts and locally is called Birdseye porphyry; the second group has a much less pronounced porphyritic texture and is part of a group of intrusive dikes and sills that range in composition and texture from diorite porphyry to aplite. This group has a much more restricted areal distribution than the Birdseye porphyry and is localized within a few miles of the contact of the Shasta Bally batholith. The rocks of second group are here referred to under felsic dikes.

Birdseye porphyry.—This light-gray to buff-colored conspicuously porphyritic rock contains phenocrysts of plagioclase, hornblende, and biotite in a fine-grained to aphanitic groundmass.. The composition ranges from diorite porphyry to dacite porphyry, and large and small dikes and sills cut all the rocks in the district except the biotite-quartz diorite of the Shasta Bally batholith. The name Birdseye porphyry was applied by the local miners to these porphyritic dikes, presumably because the centers of the zoned feldspar phenocrysts weather out and somewhat resemble the pupil of an eye. Diller (1906, p. 7) called the dikes dacite porphyry; Ferguson (1915, p. 245) called a similar dike at the Mad Mule mine in the Iron Mountain quadrangle a diorite porphyry. The term "Birdseye porphyry" as applied locally will be used throughout this report.

The most prominent areas of Birdseye porphyry are on Mad Mule Mountain, at the Uncle Sam mine, and in the Behemotosh Mountain quadrangle northeast of Backbone Creek. At the Mad Mule mine a dike of Birdseye porphyry, striking east-west and dipping 40° to 60° N., is exposed for more than 1,000 feet along the strike. It is reported by Ferguson (1915, p. 245) to have a maximum thickness of about 150 feet. The dike is at the contact between shale of the Bragdon formation to the north and Balaklala rhyolite to the south, but its south contact may be a fault. It contains large euhedral plagioclase phenocrysts as much as 1 centimeter long in a gray aphanitic groundmass.

About 2,500 feet to the northwest, at the Bright Star mine northwest of Mad Mule Mountain, and extending to the southwest, are two irregular areas of Birdseye porphyry that may be large sills, although the outcrops are not sufficient to be certain of contact relationships. Both outcrops are 400 to 600 feet wide; one is 1,700 feet long and the other 2,000 feet long in a northeasterly direction. The rock is a dacite porphyry containing euhedral platy phenocrysts of plagioclase as much as 1 centimeter long in a gray fine-grained groundmass of plagioclase and hornblende.

Another large area of Birdseye porphyry is northeast of Backbone Creek where a vertical dike 100 feet thick striking N. 60° W. is exposed for a length of 6,200 feet. The dike is intruded along a strong fault zone several hundred feet wide in shales of the Bragdon formation and has enclosed two large fault slivers. One is a sliver of probable Kennett limestone; the other is crumpled shale of the Bragdon. The porphyry has euhedral plagioclase phenocrysts 1 centimeter long in a gray aphanitic groundmass.

Many smaller Birdseye porphyry dikes are exposed throughout the district; most are about 10 feet thick and are too small to show on the geologic map of the district. In all the dikes plagioclase forms conspicuous large euhedral phenocrysts; one dike located 2,000 feet northwest of the Friday-Lowden adit contains unoriented plagioclase phenocrysts 2.5 by 2.5 by 0.5 centimeters.

The Birdseye porphyry generally contains about 20 to 30 percent plagioclase phenocrysts that range from 5 to 8 millimeters in length, and contains less than 5 percent small, inconspicuous hornblende phenocrysts in a dark-gray microgranitoid groundmass of plagioclase, quartz, hornblende, epidote, chlorite, apatite, magnetite, and carbonate. There is a distinct hiatus between size of phenocryst and groundmass, with no seriate texture between them. Some of the Birdseye porphyry contains quartz phenocrysts and as much as 20 percent quartz in the groundmass. The rock is called a dacite porphyry if it contains more than 5 percent quartz in the groundmass. Diorite porphyry contains less than 5 percent quartz and has a granular groundmass.

Under the microscope the Birdseye porphyry is seen to have markedly zoned plagioclase phenocrysts; oscillatory zoning is common. The plagioclase in the diorite porphyry is zoned from cores of $Ab_{52}An_{48}$ to rims of $Ab_{58}An_{42}$; in the dacite porphyry the plagioclase is zoned from cores of $Ab_{60}An_{40}$ to rims of $Ab_{64}An_{36}$ and is slightly altered to sericite, epidote, carbonate, and kaolinite. Alteration in places is more intense along certain zones in the plagioclase, and the central part of the crystals weathers more rapidly than the rims.

Hornblende is the most common mafic phenocryst, although a little green biotite is present in some. Hornblende occurs as green euhedral crystals that are strongly pleochroic. They are commonly 2–3 millimeters long and are much less conspicuous than the large white plagioclase phenocrysts. Hornblende is less commonly altered to green biotite or epidote.

The groundmass consists predominantly of plagioclase, quartz, hornblende, and green biotite and some orthoclase, carbonate, epidote, sphene, apatite, and magnetite, and has a micrograniitoid texture. Plagioclase in the groundmass usually is much more altered to sericite or kaolinite than the plagioclase phenocrysts.

The Birdseye porphyry commonly is an unaltered, unsheared rock. Some parts of it, however, have been sheared by faulting and are hydrothermally altered. Plagioclase is partly altered to sericite, and small pyrite cubes may be disseminated through the altered rock as reported by Ferguson (1915, p. 245) at the Mad Mule mine.

The Birdseye porphyry is probably Late Jurassic or Early Cretaceous in age. As far as is known the dikes are localized within about 20 miles of the contact of the Late Jurassic or Early Cretaceous granitoid rocks of the Klamath Mountains. They intrude the Mule Mountain stock of albite granite, which is Jurassic(?) in age. They are in turn cut by lamprophyre dikes associated with the Shasta Bally batholith.

Felsic dike rocks.—Many small dikes and sills localized in the Copley greenstone are within 3 miles of the eastern contact of the Shasta Bally batholith. They are mostly quartz latite porphyries, but range in composition and texture from diorite porphyry to aplite. A cluster of these dikes or sills is localized in the southern part of the Igo quadrangle between Igo and Mule Mountain. They strike north to northwest and dip 45° to 75° NE. conformable with foliation in the Copley. The bodies are 10 to 50 feet thick and are as much as 3,000 feet long. These dikes apparently are not related to the many porphyritic rhyolite dikes in the same area, which are composed mainly of quartz and albite and are similar to the Balaklala rhyolite. The dikes are not differentiated on the surface map and are all shown as felsic dikes. The dikes are of probable Late Jurassic or Early Cretaceous age; they have the same minerals although in different proportions as the Shasta Bally batholith and are probably related to it.

Andesite porphyry.—Some dikes of andesite porphyry, less than 50 feet thick and several hundred feet long, are present at Merry Mountain and Grizzly Gulch in the Whiskytown quadrangle, at the Reid mine in the Old Diggings district in the Shasta Dam quadrangle, and near Brandy Creek in the Igo quadrangle. The dikes are all too small to be shown on the geologic map of the district. They are dark-greenish-gray porphyritic rocks containing phenocrysts of hornblende as much as 6 millimeters long and phenocrysts of plagioclase as much as 2 millimeters long in a very fine grained groundmass. The dikes cut the Copley greenstone at Merry Mountain, in Grizzly Gulch, and at the Reid mine, and cut albite granite at Brandy Creek.

The andesite porphyry contains 20 to 30 percent phenocrysts of hornblende and about 10 percent plagioclase in a groundmass of plagioclase, hornblende, biotite, chlorite, epidote, magnetite, and apatite. Hornblende occurs as subhedral green crystals that have poor alinement. They have the following pleochroism: X=light yellowish green, Y=green, Z=dark green. Plagioclase consists of subhedral grains as much as 2 millimeters across. They are andesine and are commonly zoned from cores of $Ab_{52}An_{48}$ to rims of $Ab_{57}An_{43}$, and are twinning mainly by the Carlsbad law and somewhat by broad albite twinning. Most of the plagioclase is clear and unaltered although some cores are slightly cloudy due to alteration to zoisite, calcite, and kaolinite. Green biotite, a secondary mineral, occurs in veinlets that cut hornblende and plagioclase. Magnetite and to a lesser extent apatite are common accessory minerals.

The andesite porphyry has been sheared in places, but not metamorphosed; the hornblende is only slightly altered to chlorite, and the plagioclase is somewhat altered to kaolinite and zoisite.

The age of the dikes is not known with certainty. They cut the albite granite, which is Jurassic(?) in age. They are unmetamorphosed so are probably younger than the Late Jurassic or Early Cretaceous orogeny.

Lamprophyres.—The dikes classified as lamprophyres are all extremely fine grained hard dark-gray to black rocks with a slight greenish cast and resemble basalt. They occur as dikes as much as 50 feet thick that intrude Copley greenstone, Balaklala rhyolite, and albite granite. Some contain small megascopic phenocrysts of hornblende and less commonly plagioclase. Most of the lamprophyres occur in albite granite on the west side of the Mule Mountain stock in the vicinity of Mule Mountain, or they intrude the Copley near the albite granite contact. They cut dikes of Birdseye porphyry in the Copley. Although they appear to be unaltered rocks in the hand specimen, in thin section they are seen to be altered.

The lamprophyres contain plagioclase, biotite, epidote, chlorite, and carbonate, and some quartz or amphibole may be present. Plagioclase ranges in composition from andesine ($Ab_{59}An_{41}$) in the least altered dikes to albite ($Ab_{91}An_{9}$) in the most altered. The more calcic plagioclase occurs in laths from 0.2 to 1 millimeter long and has a pilotaxitic texture. Albite and Carlsbad twinning is common in the andesine, and broadly spaced albite twinning is present in albite. The plagioclase is more massive and has many inclusions of carbonate, epidote, and zoisite in the more altered lamprophyres.

Biotite is the most common mafic mineral in the least altered lamprophyres and may constitute as much as 20 percent of the rock. It is rarely found as phenocrysts, but occurs as tiny flakes 0.1 to 0.2 millimeters long, and is pleochroic from X=very light green to Y=Z=dark-olive green. Biotite is in part altered to chlorite even in the least altered rocks, and is completely altered to chlorite in some. Actinolite is present in some of the lamprophyres. It has an acicular habit; needles as much as 0.5 millimeter long are in subparallel arrangement. The actinolite is pleochroic from light to dark green, and probably is of secondary origin.

The least altered lamprophyres contain biotite and 5 to 10 percent epidote as phenocrysts as much as 1 millimeter long. The amount of epidote and zoisite increases as the degree of alteration increases and the lime content of the plagioclase decreases. The epidote appears to be pseudomorphous after amphibole.

The most altered lamprophyres are composed only of albite, chlorite, carbonate, epidote, and secondary quartz. The original calcic plagioclase has altered to albite and contains inclusions of epidote, zoisite, and carbonate. The mafic minerals are altered to chlorite and epidote. Some pyrite occurs in a few dikes.

The lamprophyres are probably Late Jurassic or Early Cretaceous in age and may be younger than the Shasta Bally batholith. They cut Birdseye porphyry, which is younger than the Nevadan orogeny and which is probably related to the Shasta Bally batholith. The lamprophyres cut albite granite, which is Jurassic(?) in age.

STRUCTURE

GENERAL FEATURES

The principal structural feature of the rocks of the West Shasta copper-zinc district is a broad central anticline and flanking synclines that trend a little east of north through the district (fig. 27). The anticline is gently arched along its axis and has a culmination

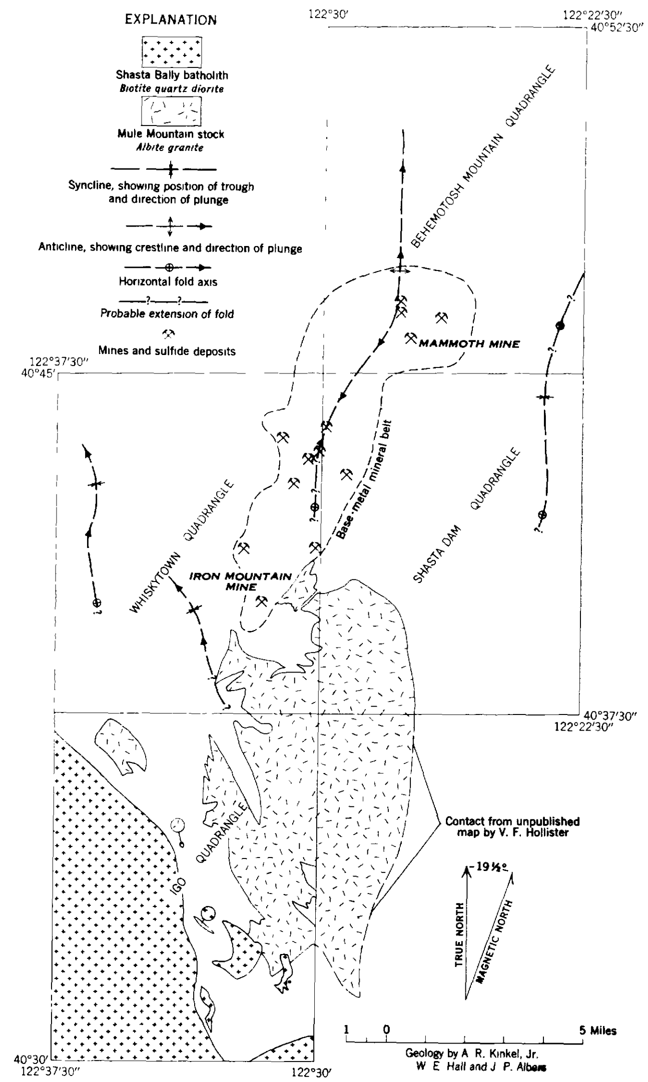

FIGURE 27.—Map of West Shasta copper-zinc district showing relationship of main folds to the mineral belt and main intrusive rocks.

over the northern part of the mining district. It plunges north at a low angle near the north edge of the mapped area, and plunges slightly to the south in the central part of the district; the southern part of the folded structure is interrupted by the Mule Mountain stock of albite granite. The synclines lie east and west of the central part of the mining district. The trough of the eastern syncline is horizontal; the western syncline plunges north at the north edge of the mapped area, but its southward extension (and a parallel syncline) is cut out and distorted by the Mule Mountain stock and Shasta Bally batholith.

The trough of the eastern syncline is marked by the erosion remnants of shale of the Kennett formation at an altitude of about 1,500 feet along the east side of the Shasta Dam quadrangle north of Central Valley. The crest of the central anticline west of this syncline trends

N. 20° E. across Mammoth Butte and Iron Mountain, and the horizon of the base of the Kennett in the eastern syncline is estimated to have been at an altitude of about 5,000 feet at the crest of the central anticline. Westward from the crest to the trough of the western syncline along the head of Whisky Creek the base of the Kennett drops to an altitude of 1,550 feet. Thus the crest of the central anticline at its highest point rises 3,500 feet above the troughs of the flanking synclines, although part of this difference in altitude is due to the thick section of Balaklala rhyolite in the central part of the area.

Although the folding is on such a broad scale that the average dip of the flanks of the central anticlinorium is not more than 20°, dips on bedding in individual folds of the anticlinorium range from horizontal to 90°. The rocks are strongly folded in some areas adjacent to areas of moderately folded rocks; local crumpling is particularly intense where there is a great difference in the competence of adjacent rocks, as between conglomerate and shale, or where thin-bedded tuff of the Copley greenstone is interlayered with massive flows, if these beds are in zones of deformation. The irregularity in the amount and degree of folding is apparently due largely to the lenticularity of flows and small intrusive rocks and the consequent abrupt differences in competence. Although the small folds generally parallel the trend of the anticlinorium, local folds may be at variance with the regional pattern where regional stresses were transformed by a body of heterogeneous rock into many local stresses of varying intensity and direction.

Strong foliation is present in many parts of the district, and is largely coincident with areas of folded rocks. Where the rocks are not folded, foliation is confined to local zones of shearing. Foliation ranges from schist and gneiss, through moderately folded rocks that contain fracture cleavage, to flat-lying volcanic and sedimentary rocks that show only minor interbed movement or bedding-plane cleavage. Bands of strongly foliated rocks a few tens of feet to several hundreds of feet in width occur adjacent to rocks that have only weak fracture cleavage, and the alternation of strongly and weakly foliated rocks is one of the characteristic features of the district. The explanation for this lies in many local geologic features, including differences in competence and composition of adjacent beds, buttressing effects, and the location of intrusive masses.

In parts of the district, foliation is parallel to bedding where this can be determined by interlayered shale or tuff; in other parts of the district it crosscuts the bedding. However, where foliation is most intense, even compositional differences are obscured, or metamorphic differentiation has formed banded rocks that simulate bedded structures. In such strongly foliated rocks it is rarely possible to determine the relationship between foliation and bedding.

The distribution of strongly foliated rocks is related in part to the location of large intrusive masses, but some of it shows no apparent relationship to these rocks. A steep foliation, that has some regularity, is poorly defined throughout much of the district. Foliation is warped, folded, and crumpled in many limited areas (pls. 1 and 3). It is impossible to explain the differences in the amount of folding, crumpling, and foliation entirely by regional compressive stresses that acted on heterogeneous rocks, even if buttressing is considered to be effective locally. The changes in strike and dip of bedding and foliation, and local crumpled areas, suggest that the intrusions crumpled the rocks locally during their emplacement. The albite granite of the Mule Mountain stock was intruded into rocks that were already foliated, as it cuts across the foliation at some places and crumples the preexisting foliation at others. The younger biotite-quartz diorite of the Shasta Bally batholith crumpled the rocks locally, but formed a zone of amphibolite, gneiss, and migmatite from the greenstone along its border; secondly planar structures in the recrystallized greenstone are mostly parallel to the contact, and possible preexisting structures are destroyed in such areas.

Faults are very common in the district, and where the rocks are well exposed on the surface or in underground workings a great many can be located. In most of the district, however, they are difficult to locate because of poor exposures along fault zones, and unless distinctive rocks are offset, faults undoubtedly would be missed. Many of the faults at the Iron Mountain mine, for instance, could not have been mapped at the surface if they had not first been located in underground workings and projected to the surface.

There are two preferred directions of N. 20°–55° W. and N. 60°–80° E. for some of the main faults. Faults in the Bragdon formation in the Whiskytown quadrangle trend N. 20°–45° W. They are steep faults that have the southeast side offset downward relative to the southwest. Their extension from the sedimentary rocks into the greenstone can rarely be traced because the greenstone along the projection of these faults is sheared and contains few distinctive units. The large fault that extends N. 55° W. across the Igo quadrangle is poorly exposed along most of its length and in the western part of the quadrangle is about parallel to foliation. Some of the faults that trend N. 20°–45° W. are occupied by quartz veins. These veins are often continuous for several hundreds of feet to a thousand feet or more in contrast to unoriented veins, which tend to be less extensive or regular. Quartz veins that strike N. 20°–45° W. are most prevalent in the Shasta Dam and Behemotosh Mountain quadrangles.

Steeply dipping faults that trend N. 60°–80° E. are exposed in many mines of the district, and as described under "Ore deposits," are economically important as they apparently control ore deposition. These faults are mineralized at many places and may have acted as channels for ore-bearing solutions. Faults that strike N. 60°–80° E. occur in the Iron Mountain, Stowell, Balaklala, Mammoth, and Golinsky mines. On all of these faults, the north side is downthrown relative to the south, and the offset ranges from 200 to 700 feet where the amount is known.

STRUCTURE OF INDIVIDUAL AREAS

THE MINERAL BELT

The mineral belt in the West Shasta copper-zinc district includes the areas of base-metal sulfide mineral deposits, which extend from about 2 miles north of the Mammoth mine to about 1 mile south of the Iron Mountain mine (fig. 27). It has no reference to the distribution of gold deposits, which extend over an area much larger than the base-metal deposits. The mineral belt is bounded on the east by the main body of Copley greenstone, where the Balaklala rhyolite has been removed by erosion, and on the west by the overlying shale of the Bragdon formation. It may extend farther west than shown on figure 27, but no information is available on such an extension because of the thick cover of younger shale beds.

Most bedding and volcanic-flow contacts in the rocks of the mineral belt have gentle dips, generally less than 30°, and although the rocks are warped into many broad folds and domical structures, they are not strongly folded except in local areas. The sedimentary rocks of the Bragdon formation, which are continuous around the nose of the central anticline north of the Mammoth mine, indicate a plunge of probably less than 10° where the Balaklala rhyolite dips under the Bragdon. The plunge of the Balaklala rhyolite north of the Mammoth mine may be even less than that in the Bragdon north of the northwestward-trending fault in Backbone Creek; the fault cuts across the nose of the anticline, and the rocks south of the fault appear to be plunging less steeply than those north of the fault. The Balaklala and Copley formations are not exposed again along the strike of the anticline for about 8 miles to the north, where a window of these rocks is exposed on Dog Creek.

Irregular gentle warping, shown by bedding in tuff in the Balaklala rhyolite, or by the contact between the upper and middle units of the Balaklala, is the characteristic structure along the crest of the central anticline from the Sutro, Golinsky, and Mammoth mines in the north to the Stowell mine in the south. Basin-shaped warps are present at the Balaklala and the Shasta King mines, but the Mammoth and probably the Keystone mines are on arched or domical structures. Other basin-shaped structures, in addition to warps due to folding, occur where thick lenticular deposits, either pyroclastic layers or flows, were deposited in a primary sedimentary basin. Between the Mammoth and the Sutro mines this type of basin is filled with coarse pyroclastic rocks and tuff.

South of the Stowell mine the broad central anticline is more closely compressed and passes into a series of folds which extend through the Iron Mountain mine area. Here the flanks of the folds dip as steeply as 60°–70°. These folds, in turn, are cut off by the Mule Mountain stock of albite granite.

On the east flank of the central anticline the beds dip gently east from the Golinsky mine, and the Balaklala rhyolite is overlain by the Kennett formation. However, south of this area and east of the Spread Eagle and Sugarloaf prospects and the Iron Mountain mine, the rocks are strongly folded into many smaller, less regular anticlines and synclines. On the west flank of the central anticline, the dips are regular and gentle, and the Balaklala rocks are overlain by the Kennett and the Bragdon formations. Here again, as the southern part of the anticline is approached, the broad structure disappears into smaller, less regular folds.

Many of the rocks in the mineral belt are weakly foliated or sheeted, but some are strongly foliated in local bands. Difference in competence and the lenticularity of most of the flows and pyroclastic rocks have greatly affected the localization and distribution of the weaker types of foliation and sheeting; bands of strong foliation cut through all rock types. Two types of foliation are present: one is a steeply dipping planar structure that ranges from rather widely spaced jointing or closely spaced sheeting to fracture cleavage that has films of alined secondary minerals along closely spaced planes. The rock between cleavage planes in this type of planar structure is unaffected; flow cleavage is common locally along zones of intense movement, but these zones are rare in the mineral belt.

The second type of foliation common in the mineral belt is bedding-plane foliation formed by flexural-slip during folding. Even though most of the folding is gentle, bedding-plane foliation is locally well developed, probably because of the great difference in competence in the layered rocks. This type of foliation forms in zones that contain tuff beds and volcanic breccia that were less competent than the flows, and adjustment between flows was taken up along the flow contacts rather than within the flows.

Planar structures are not evenly distributed in the rocks. A thick flow, such as the coarse-phenocryst rhyolite, acts as a competent body and is foliated much less than the underlying series of lenticular flows and pyroclastic rocks. Steep foliation is more common in some areas on the axes than on the flanks of folds. Thus, under some conditions, a body of foliated rock has a linear element, that is, certain flows are foliated, more strongly along fold axes, and form an elongate body that is limited above and below by less foliated flows and which plunges with the fold axes.

KENNETT AND BRAGDON FORMATIONS

The Kennett and Bragdon sedimentary rocks are characterized by alternate competent and incompetent layers, which almost entirely controlled their reaction to stresses. The formations are composed principally of thin bedded shale and siltstone, which acted as incompetent units, but they also contain beds of chert and siliceous shale, limestone, and many beds of sandstone and conglomerate, which formed competent layers. The common form of disturbance is a broad folding of all the sedimentary rocks, accompanied at many localities by mashing and crumpling and much interbed slippage in incompetent layers, particularly where they are interbedded with competent layers. The broad folds can be traced by following the beds of conglomerate and sandstone, which do not reflect the crumpling and mashing of the shale; attitudes of shale beds between conglomerate or sandstone are unreliable indicators of structure at most localities because of local crumpling. Fold axes in shale between competent layers commonly are unoriented because the competent unit is broken into blocks that have been mashed into the adjoining incompetent layers. Boudinage structure occurs along some conglomerate and sandstone beds, particularly where these are overlain and underlain by a considerable thickness of shale.

Isoclinal folds are rare, but close folds are common, having amplitudes of a few feet to several hundred feet and dips on the flanks that range from 30° to 60°.

In common with the other parts of the district, the distribution of folds and crumples in the sedimentary rock is erratic. Rocks that are strongly folded and cut by many small faults are overlain and underlain by rocks that have regular bedding for a considerable distance, and the areas of strong disturbance have little continuity, that is, they can seldom be followed in one direction for any great distance. To some extent this is due to the difference in reaction to stress between a thick sequence of shale that contains no conglomerate, as in the shale above and below the main conglomerate zone, and a sequence in which shale and conglomerate alternate. However, part of the main conglomerate zone is only weakly folded, and the interbedded shale beds in these places show only minor interbed slippage; localized stresses apparently account for the strong disturbance of the sedimentary rocks in some areas.

The Kennett formation, throughout much of the district, contains a thin-bedded black siliceous shale interlayered with chert at the base of the formation. This lower part of the Kennett is intensely crumpled in many exposures. It rests on rhyolite or greenstone, both of which were apparently more competent rocks than the siliceous shale, and interbed movement that occurred between the Kennett and the underlying formations was taken up largely in the siliceous shale. Considerable movement must have taken place to account for the crumpling in the basal Kennett, but the contact between the thick beds of underlying lava and overlying shale would be a plane of strong disturbance during folding.

Although cleavage is parallel to bedding in most of the shale of the Bragdon formation, locally a steep slaty cleavage is formed where the shale has been closely folded. This steep slaty cleavage may strike parallel to bedding but have a steeper dip, or it may be transverse cleavage at any angle to the bedding. Pencil structure, formed by the intersection of slaty cleavage and bedding, is common in the crests of these minor folds; the slaty cleavage and pencil structures locally obscure the bedding. The steep slaty cleavage and the pencil structure rarely can be followed along strike in the field because they are commonly prominent only in one series of beds and extend along the axes of plunging folds. The structures may be entirely absent in overlying or underlying beds of different competence, or only fracture cleavage or bedding-plane foliation may be present in these beds. Flexural-slip folding, in which interbed movement forms bedding-plane cleavage (that is sometimes cut by poorly defined fracture cleavage) in the incompetent beds and strong steep joints in the competent beds, can be seen at many places. Figure 28 is a sketch of a road cut in shale and sandstone of the Bragdon that illustrates this type of folding.

The pebbles in many of the conglomerate beds are elongate and are alined parallel to the bedding. This is apparently a primary feature, as elongate pebbles occur in areas where there is no other evidence of disturbance of the sediments. However, at a few localities where folding has been intense, the pebbles are stretched, and their long dimension is not parallel to

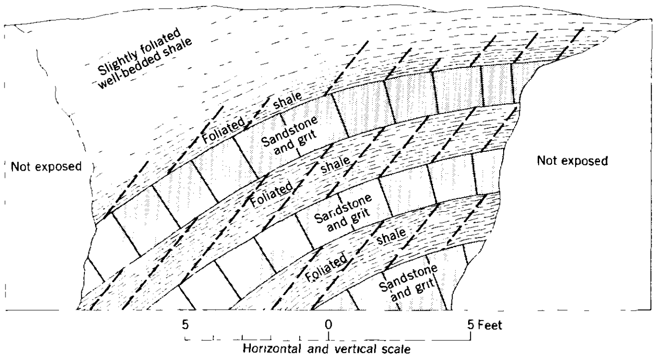

FIGURE 28.—Cross section of shale and sandstone of the Bragdon formation in road cut. Joints formed in competent standstone beds at right angles to bedding. Bedding-plane foliation and weak fracture cleavage formed in shale beds.

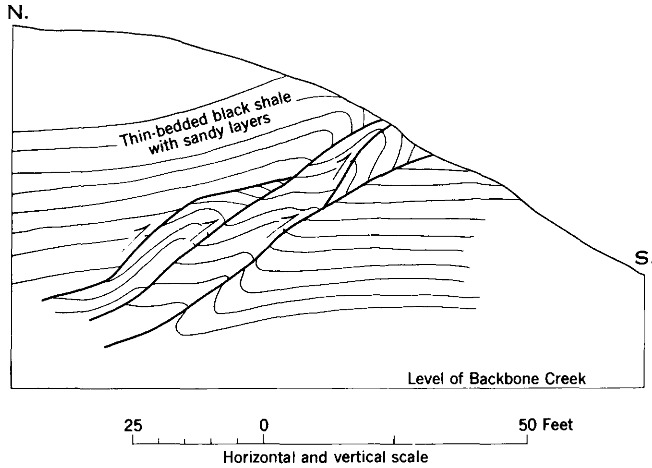

FIGURE 29.—Small thrust fault in shale of the Bragdon formation in Backbone Creek.

the bedding. Rock flowage has occurred at these localities, but without the formation of foliation in the conglomerate; no foliated conglomerates occur in the district.

The Kennett and Bragdon formations are cut by steep normal and flat thrust faults. The normal faults trend N. 20°–40° W. and N. 60°–80° E., which is the same direction as those in the Copley and Balaklala rocks. The northeast and northwest sides are downthrown, as shown by the repetition of the main conglomerate beds. Fault zones are rarely exposed because the mashed shale beds do not crop out, but can be traced by a combination of exposures in canyons, slickensided float, abrupt termination of conglomerate beds, and the pattern on aerial photographs. At some places the faults cause a mashing of the shale for several hundred feet; at others the disturbance of the shale is very slight. Where the faults are exposed, as in deep canyons, the brecciated zone ranges from a few feet to 50 feet in width and consists of jet black shiny slickensided lenticles of foliated shale. Very little clay gouge is present. The faults having the widest zones of mashing are those that form the contact between the sediments and the rhyolite in Backbone Creek and between the greenstone and the sedimentary rocks just east of Shirttail Peak.

Low-angle thrust faults are common in some localities in the sediments. They are prominent along Backbone Creek and in the area east and south of Shirttail Peak. The dip of these faults ranges from nearly horizontal to 40°; they tend to cross bedding at a low angle for a few hundred feet, then turn parallel to the bedding and disappear in bedding-plane movement. Figure 29 shows a thrust fault near the junction of Backbone and Fall Creeks.

The faults that cut the sediments are not mineralized as a rule, although a few are accompanied by quartz veins that contain sulfides, as at the Gladstone and American gold mines, or are occupied by dikes of Birdseye porphyry (dacite porphyry) that is hydrothermally altered.

COPLEY GREENSTONE IN THE IGO AND WHISKYTOWN QUADRANGLES

The structural features in the Copley greenstone in the western part of the district were formed in part by regional stresses and in part by the intrusion of the two large plutons in the Igo quadrangle. Deformation in the greenstone consists largely of broad open folds away from the intrusions, but adjacent to them the greenstone beds are altered to amphibolite schist, gneiss, and migmatite. Local zones of shearing, ranging from tight fault zones to wide bands of chlorite schist, are common throughout the greenstone.

Most of the Copley greenstone in the Whiskytown quadrangle north of Clear Creek and west of Whisky Gulch contains gentle open folds, and weak foliation. Individual flows, such as the layer of pillow lava and pyroclastic rocks east of Grizzly Gulch, can be traced continuously for several thousand feet, and the flows have gentle dips. The rocks are not sheared except where fault zones cut through the gently dipping lavas; the flow contacts, and the pyroclastic and tuff beds, preserve their original characteristics. South and southeast of this area the greenstone becomes more deformed as the intrusive masses are approached, until near the intrusion the greenstone has been altered and strongly deformed.

In the eastern part of the Whiskytown quadrangle a foliated structure was present in the Copley greenstone before the intrusion of the Mule Mountain stock. In this area unfoliated albite granite cuts across foliated greenstone and in places crumples the foliation. The small plug of albite granite in the northeastern part of the Igo quadrangle also appears to have crumpled the foliation, as if by a shouldering action, along the north edge of the intrusion (pl. 3).

The Shasta Bally batholith of biotite-quartz diorite formed migmatite, gneiss, and amphibolite schist from the greenstone for a considerable distance from its contact. Foliation in the Copley greenstone generally parallels the contact of the batholith, but it is not at all certain how much of this foliation was formed by the intrusion of the batholith. At some places gneissic structure probably follows a previous foliation in the greenstone (which may or may not be parallel to the bedding), and at other places it follows primary bedding structures.

Primary structures in the Copley greenstone are largely destroyed by recrystallization in the gneissic zone, but a few beds of black shale occur in the recrystallized greenstone in Brandy and Boulder Creeks in the Igo quadrangle and the gneissic structure is parallel to the bedding at these points. Gneissic structure, marked by differing degrees of recrystallization, is also parallel to the bedding where the batholith intrudes shale of the Bragdon formation east of Buckhorn Summit on U. S. Highway 299W. However, along the edge of the batholith north of Andrews Creek, and at the Greenhorn mine west of the Igo quadrangle, the intrusive contact cuts across the gneissic banding that follows earlier foliation in the greenstone. Recrystallization tended to work out along primary structures in the wall rocks and along the outer edge of the gneissic zone, where it has extended in long tongues into the greenstone along lines of earlier foliation. It is not possible to determine at most places whether the earlier foliation is parallel to bedding or not. Recrystallization tended to follow and emphasize any structure that was in the wall rocks at the time of the intrusion of the batholith, but it is also possible that locally the parallelism between the planar structures in the intrusive mass and in the wall rocks was formed by frictional drag along the walls of the intrusion.

The small plug of biotite-quartz diorite in Clear Creek in the southeastern part of the Igo quadrangle crumpled the foliation of the Copley greenstone along its northern border, but in places also cuts across the foliation. The screen of greenstone between the Mule Mountain stock and the Shasta Bally batholith, being affected by both intrusions, contains no relict primary structures. This greenstone is a strongly foliated chlorite or amphibolite schist near the Mule Mountain stock and an amphibolite gneiss near the Shasta Bally batholith. At most places the foliation in the screen is parallel to the contacts of the intrusions, but in a few places is discordant to them at the immediate contact.

SHASTA BALLY BATHOLITH

Only a small part of the Shasta Bally batholith is included in the Igo quadrangle, but reconnaissance across most of the batholith has shown that the structures in the mapped area are typical. Reconnaissance trips were made across the batholith at Bully Choop Mountain west of the Whiskytown quadrangle, and across parts of the Trinity Alps to the northwest. The batholith is at least 20 miles long in a northwesterly direction, and as much as 9 miles wide. In contrast to the Mule Mountain stock, the Shasta Bally batholith contains primary planar and linear structures. Primary foliation is common along much of its contact, but linear structures are limited to alined hornblende prisms on the southern part of the contact in the Igo quadrangle, where more hornblende is present than elsewhere in the intrusion.

Foliation in the intrusive mass is invariably parallel to the contact. It is a primary flow structure that shows a banding of oriented light and dark minerals. Along the contact of the intrusion, there is no difficulty in the field in distinguishing between rock that was molten and is flow banded, and altered greenstone in which much granitic material was added to the greenstone (migmatite zones), although mineralogically the two may be similar. Primary structures are more common near the edge of the intrusion than in the interior, but they are formed in the interior in the parts that are topographically high, and thus near the roof of the intrusion. Flow banding is regular and uniform at most localities (pl. 2), and dips steeply to the east on the eastern side of the intrusion, parallel to the contact. In the central part of the batholith, in the headwaters of Eagle Creek and north of Bully Choop Mountain to the west, the dip of the flow banding ranges from 10° to horizontal. West of Bully Choop Mountain along the west edge of the intrusion, only a small amount of flow banding was observed, but all of it dips steeply to the west. It thus appears that the Shasta Bally batholith forms a schlieren arch, and that the top lies only a short distance above the present erosion surface. The exposures of many large and small bodies of the same rock for many miles to the northwest along the strike of the Shasta Bally batholith, and the wide metamorphic halo that parallels this zone suggest that the intrusions are only the top of a much larger underlying mass.

Alined hornblende laths give a linear character to the rock at a few places. The distribution of the laths is so limited and erratic, however, that no general alinement can be traced except in the area north of the town of Igo. In this area the alinement of hornblende laths is roughly S. 30° E.; the plunge is 45° SE.

Jointing is common in some parts of the intrusion, but absent in others. Too little of the batholith is exposed in the mapped area to justify a detailed study of the jointing, but no marked pattern was recognized. A few aplite dikes cut the biotite-quartz diorite, but they do not appear to follow a regional joint pattern.

Faults are not common in the Shasta Bally batholith in the area included in the Igo quadrangle. A few faults can be traced for several hundreds of feet in the Silver Falls mine northwest of Igo and at other prospects in that area, but faults generally are rare in all parts of the batholith, either in the mapped area or to the west and north.

MULE MOUNTAIN STOCK

The massive granitoid Mule Mountain stock contains no planar structures that are related to its intrusion. The stock is slightly foliated, but it transgresses the foliation of the enclosing greenstone at many places. At a few localities foliation extends from the wall rocks into the intrusion and decreases in intensity. It seems probable that late orogenic stresses followed lines of earler foliation, and extended it from the wall rocks into the intrusive mass. Local schistose bands range in width from a few feet to several tens of feet at a few places, but these are fault zones. Some areas of autobrecciation occur locally in which angular to rounded blocks of albite granite are embedded in a matrix of the same material. They are probably intrusive breccias. Much of the rock in the outcrop is shattered and mashed to some extent, and cataclastic structures are commonly seen in thin sections.

Xenoliths of the intruded rocks are common in the Mule Mountain stock, particularly along its northeastern border. These are of two types: elongate schlieren of greenstone that range from large bodies to mere wisps of mafic minerals, and angular to rounded fragments of greenstone formed by brecciation and penetration of the wall rocks by the intrusion. These contact breccias are described under albite granite. Their occurrence well into the intrusion near its northeast border (pl. 1), and large masses of greenstone in the intrusive body south of the South Fork Mountain in the Whiskytown quadrangle, suggest that some of this material occurs as roof pendants, and that in these areas the top of the stock was not far above the present erosion surface.

The small plug of albite granite in the northeastern corner of the Igo quadrangle contains a planar structure that is most strongly developed in the southwestern half of the plug, where the albite granite is a leucocratic gneiss. The gneissic structure is parallel to and continuous with the gneissic structure in the surrounding greenstone. It was formed by recrystallization under stress due to the intrusion of the Shasta Bally batholith, and contains material added from the batholith.

LOCATION OF VOLCANIC CENTERS

Two centers of eruption of the rhyolitic rocks are known, and several other localities are regarded as probable centers. None have been located that might have been the source of the mafic rocks of the Copley greenstone.

The best exposure of an eruptive center, which was the source of the cumulo dome of coarse-phenocryst rhyolite and its associated pyroclastic rocks, is in the vicinity of the Uncle Sam gold mine. Erosion of part of this center below the level of its surface flows has exposed a broad volcanic neck that is complex in structure. The wall rocks of this volcanic neck are principally nonporphyritic rhyolite, whereas the central part exposed in the long adit of the Uncle Sam mine is composed mainly of coarse-phenocryst rhyolite, but bodies of nonporphyritic rhyolite were also found.

Most of the coarse-phenocryst rhyolite in the main vent is identical in appearance with the same rock that occurs as a cumulo dome, except that it contains chlorite and has been silicified. In the immediate vicinity of the main vent, coarse-phenocryst rhyolite occurs as vertical dikes, lens-shaped masses, and stockworks in the shattered walls. Many xenoliths of nonporphyritic rhyolite are present, but some of the bodies of nonporphyritic rhyolite probably represent wall rock that is in place, the coarse-phenocryst rhyolite having been intruded on both sides of such bodies. The vent must be several thousand feet across, although some of this area is occupied by shattered wall rock and screens of wall rock and flows cover the northern half of the vent. Excellent exposures of the south side indicate that the coarse-phenocryst rhyolite forced its way into the rocks in several large and small coalescing strands, rather than as a single body. Much of the wall rock was shattered, probably by explosive activity at the time of the earliest extrusion of the coarse-phenocryst pyroclastic material that underlies the main dome of coarse-phenocryst rhyolite.

A second vent, the outline of which is below the level of Shasta Lake except on the south side, is exposed in the northeast corner of the Shasta Dam quadrangle. A jumble of several types of rhyolitic material was thrown for several thousand feet from the second vent. To the south the fragments decrease in size and quantity, until about 1 mile distant only a few rhyolitic fragments can be found in a matrix of greenstone tuff.

Distribution of rhyolitic flows and pyroclastic beds indicate the presence of other vents in the district. Rhyolitic flows probably did not extend far from their source, nor did relatively thin beds and stubby lenses of coarse pyroclastic material originate far from their present position. A vent probably existed near the southeast corner of the Igo quadrangle because of the concentration of thin rhyolitic sills, flows, and minor pyroclastic beds in that area. A group of radially arranged rhyolitic dikes on Copley Mountain suggests that this may be a deeply eroded vent. The group of rhyolitic flows and pyroclastic beds east of the Sacramento River in the south central part of the Shasta Dam quadrangle suggest a vent in that vicinity, because many of the pyroclastic beds contain coarse fragments and many of the flows are thin. It seems probable, for the same reasons, that vents were present near Merry Mountain, near the small plug of albite granite in the northwest corner of the Igo quadrangle, and near Whiskytown.

In addition to a main vent at the Uncle Sam mine, there were probably several along what is now the mineral belt. The interlayering of many lenticular flows and pyroclastic beds, and the impossibility of determining more than a generalized stratigraphic sequence for the different types of flows makes it probable that the volcanic material came from many vents. In addition, some of the bodies of pyroclastic material have rounded outlines and may represent breccia pipes or breccia-clogged volcanic necks.

PHYSIOGRAPHIC FEATURES

The West Shasta district is in the southeastern part of the Klamath Mountains in the Pacific Border physiographic province as described by Fenneman (1931). Two landscape forms stand out prominently in the district. One is the deeply dissected mountains and foothills that cover most of the district; the other is the broad, flat terrace that lies about 200 feet above the Sacramento River. Traces of early topographic cycles are found in the higher mountainous area, but few traces of them remain in the foothills.

A surface with low or moderate relief existed in Pliocene or Miocene time in most of the area now occupied by the Klamath Mountains (Diller, 1894, p. 404, 1902, p. 9; Fenneman, 1931, p. 465–471; Hershey, 1904, p. 423–474; Lawson, 1894, p. 241–271) it is here referred to as the old Klamath surface. Uplift of the area into a high plateau in Pliocene or early Quaternary time renewed the cutting power of the streams, which deeply dissected the old surface. Viewed from the present valleys, no indication of the old surface can be seen in the West Shasta district, but from points of vantage, such as Shirttail Peak or Mad Mule Mountain in the Whiskytown quadrangle, an old surface can be inferred from the general accordance in altitude of ridges and peaks. As Cotton (1949, p. 138) points out:

> Accordance of summit levels does not indicate with certainty the destruction of an uplift plane, but the former existence of such a surface may usually be inferred with a fair degree of confidence where accordance is well marked with a restored surface either plane or domed.

Figure 30, a view from Shasta Dam of the mountains in the West Shasta district, shows the general accordance in the altitude of the peaks.

The old Klamath surface was probably a surface of moderate relief, as the accordance of the altitudes of ridges and peaks, though reasonably close, is not close enough to imply a level plateau before it was dissected. This surface sloped easterly. The highest part of the peneplain over the Salmon, Yollo Bally, and Trinity Mountains west of the Sacramento River is at an altitude ranging from 6,000 to 7,000 feet. In the Shasta mineral belt to the east of this mountainous areas, there is a good accordance of many ridges and peaks at an altitude of about 4,000 feet, although a few peaks are above this general level. The relatively level ridge that extends north from Iron Mountain, and many other ridges north and northwest of Iron Mountain have an altitude of about 4,000 feet, but a few peaks lie above this level, for example those east of Shirttail Peak, which have altitudes between 4,000 and 4,600 feet. These probably are monadnocks on the old surface, rather than a higher surface, as there is little summit accordance above 4,000 feet.

It is possible that part of the old Klamath surface may be an exhumed surface of low relief that was present at the base of the Cretaceous rocks. These are considerably softer than the pre-Cretaceous rocks, and scattered outcrops north and west of the West Shasta district suggest that Cretaceous rocks may have occurred at about the level of the old Klamath surface in part of this area. Isolated rounded stream cobbles, commonly of rock types foreign to the West Shasta district, are found on many of the highest ridges and peaks; they may be residual from gravel of Cretaceous age.

The levels of a few terraces below 4,000 feet are discernible. They are remnants of erosion surfaces of low relief located on the sides of canyons and may represent a system of dissected terraces. At a few localities these terraces contain a thin veneer of gravel, but generally the surface is covered, and its shape is modified by slope wash. One of these terrace remnants occurs between the altitudes of 3,450 and 3,650 feet in the West Shasta district. The difference in altitude between the terrace above the Keystone mine at 3,450 feet, that northwest of Sugarloaf Mountain, indicated

FIGURE 30.—View looking westward from Shasta Dam showing general accordance of summit elevations and entrenched valleys on left. Hills are underlain mostly by Balaklala rhyolite.

by the ridge at 3,550 feet, and the ridge northwest of South Fork Mountain at about 3,100 feet may be due to valley floors on separate drainage systems, or to different distances from the headwaters in the same drainage basin. An old valley floor remains at a few places that probably represents a terrace at an altitude of about 2,250–2,300 feet.

The old valley floor of the Sacramento River, in which the present river is entrenched, is a broad level valley several hundred feet above the present river bed in the vicinity of Redding, but it grades into the foothills a short distance to the north. The broad floor of the old valley, at an altitude of about 725 feet, slopes gently to the southeast and is a striking feature of the landscape. The area near the foothills is a terrace cut into the bedrock by the Sacramento River and is bordered by pediments and by fans formed by tributaries. A few miles downstream from the foothills the bedrock is veneered with gravels of the Red Bluff formation, which thickens toward the main part of the valley. The Recent streams are so deeply entrenched into the old valley floor where it was composed of gravel of the Red Bluff formation that in some places only remnants of the floor remain.

Traces of older landscapes generally are preserved only on the higher slopes in the mountainous area, but locally the interfluves between the deeply incised present tributaries are smooth. Headward erosion of tributaries and the destruction of interfluves by slumping is rejuvenating the topography of much of the district. Also, some of the ridges and peaks of moderate altitude below the old Klamath surface have subdued and rounded forms and a gentle summit convexity, indicating a mature topography. Traces of this mature topography are found mainly above 2,500 feet, and it appears probable that in addition to the Klamath surface, a later mature surface formed in parts of the district when the main streams were below an altitude of 2,000 to 2,500 feet (allowing for uplift since that time).

In some parts of the district, lower slopes, generally less than 2,000–2,500 feet in altitude, have accumulated rock debris and soil as much as 100 feet thick. The unstratified deposit is composed of rock fragments as much as 1 foot in diameter in a matrix of red soil. This unusually thick mantle was formed by solifluction and rainwash from the higher slopes, many of which now are bare. The eroded material aggraded lower slopes, and the increased runoff on the denuded upper slopes accelerated the process. Steep-walled gullies, as much as 75 feet deep along the road to the Mammoth mine, are now working headward in the mantle of wash and are restoring graded slopes in continuity with the upper slopes.

Landslides are a common feature of the district because the walls of gullies and the canyon slopes along most of the streams are oversteepened. They range from small mudflows and slump along gullies to large slides and earth flows that are several thousand feet across. Slumping occurs along oversteepened drainage areas where the soil and subsoil cover is deep or where the rocks are soft and deeply weathered. Slump and solifluction supply much debris to the streams during the rainy season; larger slumped areas cause temporary dams across the streams, and the breakdown of the loose damming material causes disastrous floods in some canyons.

The geologic map (pl. 1) shows the larger landslides. They are generally shallow, and sliding apparently has taken place on several spoon-shaped surfaces in the upper part of the slide and as a mudflow in the lower part. There are gradations between extensive solifluction, debris slides, and true slides that are bounded by sharply defined planes of slippage. The tops of the slides do not have a backward slope, as erosion along scarps and aggrading of the flattened top forms a flat area that slopes with the topography but is flatter than the hill slope. Gullies cut in the older slides show that the jumbled material is unstratified, or poorly stratified. The thickness of some of the larger slides is known either from drill holes, as at Iron Mountain, or from deep gullies that have exposed the bottom of the slide. They are unusually shallow; it is doubtful if the thickness of the larger slides exceeds 200 to 300 feet.

The recognition of landslides is of importance in the district because much ill-advised prospecting has been done in material that is not in place, and because the topographic flats at the top of landslides that may still be active provide what appear to be attractive plant sites in the generally steep topography. Not all flat or gently sloping areas are due to landslides; the recognition of landslides may be difficult unless recent gullies expose the jumbled material that composes the slide. Flat or gently sloping areas that suggest landslide tops occur at many places, but some topographic flats are remnants of older landscapes, or are due to differential erosion (fig. 31). The canyon side of an old terrace level, where it has been oversteepened by Recent stream erosion, is a favorite place for a landslide.

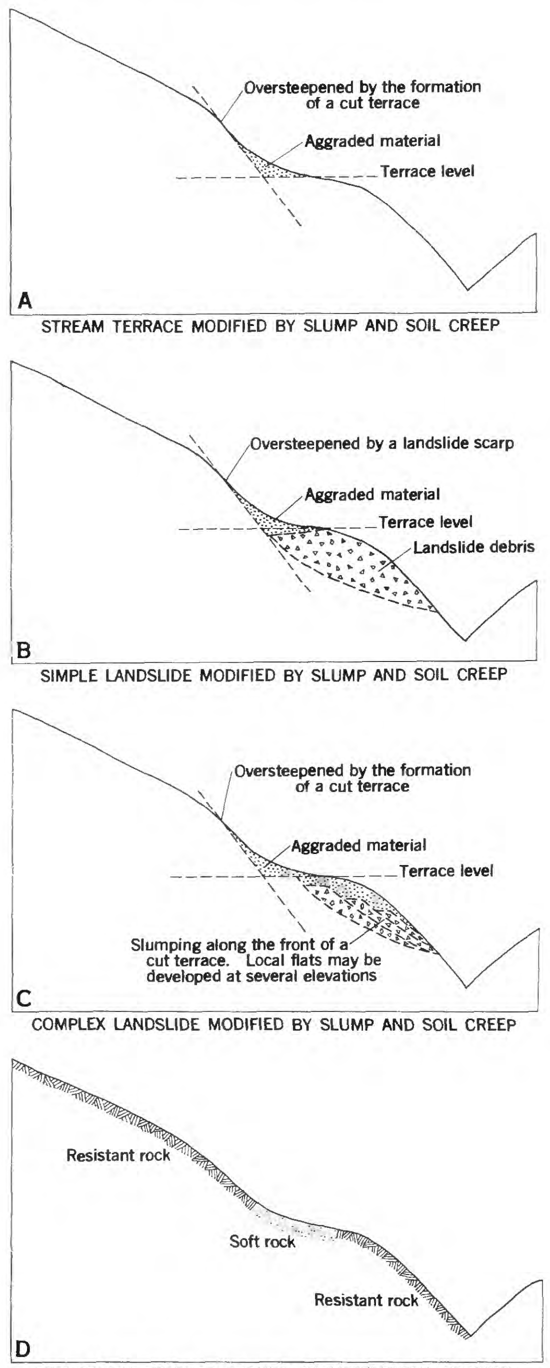

FIGURE 31.—Formation of topographic flats. Four processes, all of which are represented in the West Shasta copper-zinc district.

The present dendritic drainage pattern appears to be superposed from the old Klamath surface onto the older rocks, and except in local areas and where streams follow well developed faulting or sheeting in the older rocks, the streams are not controlled by structures in the underlying rocks (fig. 32). However, some of the streams have become adjusted to the underlying rock structure as the more resistant rhyolite and the plutons underlie much of the higher topography and the less resistant greenstone occurs mainly in the lower areas. Most of the streams are cutting narrow V-shaped canyons at the present time, but where bodies of harder rocks have impeded downcutting, a few streams have reaches with a low gradient and locally some gravel fill. In a region where there are a few large permanent

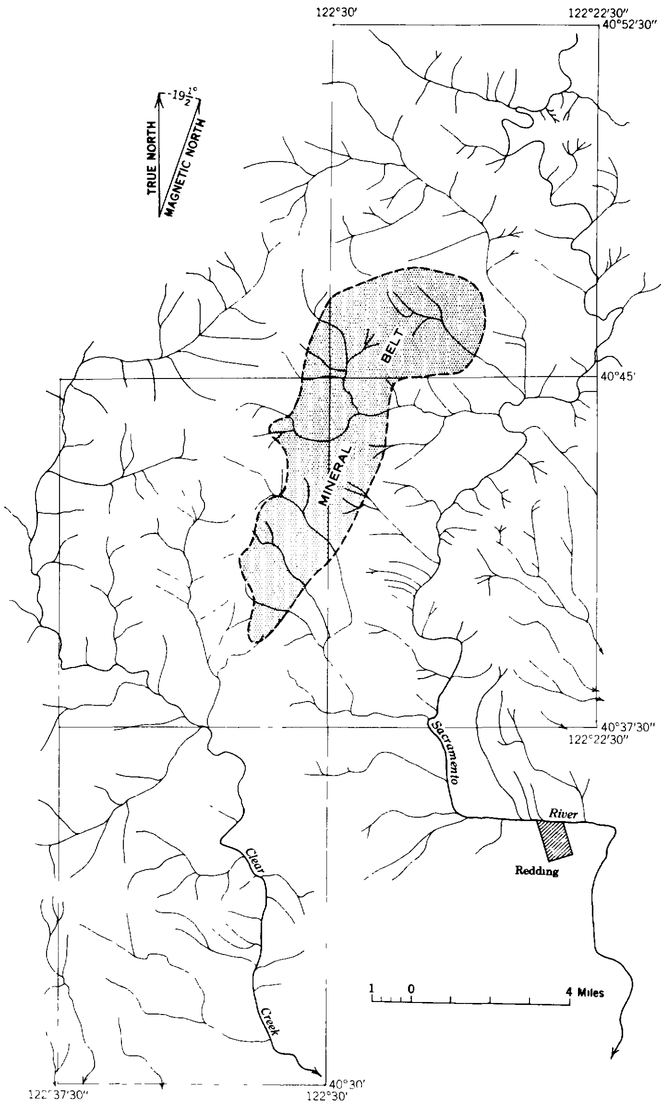

FIGURE 32.—Drainage pattern, West Shasta copper-zinc district.

streams but many small streams that run only in the rainy season, the difference in topographic expression between the two types is noticeable. The large streams at some places cut more rapidly than the tributaries, and the tributaries have discordant junctions along many of the permanent streams. Hanging junctions of this type are common along Clear Creek below Whiskytown and along some sections of Backbone and Whisky Creeks. Main streams such as Clear Creek in the Igo quadrangle occupy narrow, youthful valleys that are entrenched in broad older valleys.

Small patches of gravel of the Red Bluff formation occur on the mature slopes of the broad old valley of Clear Creek. The scattered remnants of this soft and easily eroded formation suggest that the broad valley may be an exhumed surface formed during a much earlier cycle. The grade of Clear Creek and slope of the canyon walls are complex where the stream crosses rocks that differ in their resistance to erosion; sections of the stream have a low gradient, and at a few places the stream is braided and is aggrading, but at other places rapid downcutting is taking place.

The intermittent streams form narrow canyons that have a steep gradient and oversteepened walls. Minor deposits of gravel, which were important sources of placer gold, have collected where the gradient has not been sufficient to transport large boulders. In these areas a small reach of the stream has been aggraded, but most of the intermittent streams during the rainy season consist of rapids and waterfalls in narrow, V-shaped, brush-clogged valleys. Gully erosion is widespread; a large part of the gullying was initiated or greatly accelerated by the removal of the timber for mining purposes and the destruction of timber, brush, and grass cover by fumes from heap roasting and smelters during the early mining operations.

The Sacramento River near Redding is entrenched as much as 200 feet below the level of the broad valley, at an altitude of about 725 feet, that was formed at an earlier stage. North of Redding the broad valley disappears where the Sacramento River enters the mountainous area, and in most of the West Shasta district, and to the north, the river occupies a deep narrow canyon. The canyon at Shasta Dam afforded a good location for a dam because above this point the present river valley widened at many places and contained bottom land along reaches with a low gradient that allowed a large area for water storage.

GEOLOGIC HISTORY

The geologic record in the West Shasta district from Middle Devonian time through part of the Mississippian appears to be one of continuous sedimentation and volcanic activity. The deposition of a great thickness of marine sediments (Hinds, 1933, p. 81–86), exposed in Trinity County and probably present under the rocks exposed in the West Shasta district, was interrupted in the Middle Devonian by the outpouring of mafic and siliceous lava on the ocean floor. Sedimentation was resumed in the Middle or Late Devonian, and the sedimentary sequence continued into the Mississippian. In the West Shasta district, the upper part of the Mississippian is eroded, and only the younger rocks of Cretaceous and Recent age crop out. In nearby areas, however, the record of post-Mississippian sedimentation and volcanic activity is more complete. Rocks of Pennsylvanian and early Permian age are absent, but late Permian, Triassic, and Lower Jurassic sedimentary and volcanic rocks occur immediately to the east of the West Shasta district (Diller, 1906, p. 10).

Orogeny occurred in the Late Jurassic or Early Cretaceous, accompanied by the intrusions of batholiths; a major unconformity separates the rocks of Paleozoic age from those of Late Cretaceous age and a major nonconformity separates the Shasta Bally batholith of Late Jurassic or Early Cretaceous age from the overlying Lower Cretaceous Paskenta and Horsetown formations of Hinds (1933, p. 113) that crop out southwest of the West Shasta district. Another erosional gap in the sedimentary record occurs between the Upper Cretaceous and the Eocene rocks. Sedimentation in the Eocene and Pliocene was probably partly in oceanic bays and partly in fresh water in the southern part of the area, but volcanic highlands were formed in the northern part. Pliocene rocks in the Redding area consist largely of subaerial lava flows and mudflows (Hinds, 1933, p. 116).

The oldest formation in the West Shasta district is the Copley greenstone of probable Middle Devonian age. The Copley is probably underlain by rocks that crop out west of the Shasta Bally batholith; these are the Chanchelulla formation of Hinds (1932), the Salmon hornblende schist, and the Abrams mica-schist that are described by Hinds (1933, p. 78) as underlying the Copley, and thus are of pre-Middle Devonian age. The latter formations contain a large amount of sedimentary material that was probably deposited in oceanic basins, but they contain some mafic lavas that are interbedded with the sedimentary material.

The mafic flows have a sporadic distribution in the formations that underlie the Copley indicating that volcanic activity began slowly at isolated points during the sedimentary cycle preceding the Copley and gained in intensity and areal extent until mafic flows and pyroclastic material constituted almost the entire deposit over a large area.

The Copley has a large areal extent as compared with its thickness. It is composed of many mafic flows and bodies of pyroclastic material, none of which themselves are widespread. The formation covers an area of at least 1,000 square miles. Its limited thickness, the presence of pillow lavas, water-deposited tuff and shaly tuff, the minor local folding in shaly layers that was caused by syngenetic sliding, and the lack of widespread beds of tuff, all suggest that the formation was built up under water on the underlying sedimentary formations. The Copley flows and pyroclastic rocks must have been derived from many widely scattered vents, as the individual flows can rarely be traced for more than a few thousand feet, and the bodies of pyroclastic rock tend to be stubby lenses that cover irregular areas of very limited horizontal extent.

Rhyolitic lava and pyroclastic material began to erupt at a few centers in the widespread area of mafic flows in the latter part of the main period of deposition of the Copley, and in the upper part of the Copley rhyolitic material is fairly common in widely scattered localities. In the West Shasta district centers of eruption of the rhyolitic rocks (the Balaklala rhyolite) formed at about the end of deposition of the Copley, and although there was an overlap in some areas, the beginning of the main period of rhyolitic eruptions marks the end of the major eruptions of mafic flows. A well-defined transition zone between mafic and rhyolitic flows is present at many places; it is marked by coarse, unsorted, rhyolitic and mafic bombs and fragments in a matrix of Copley greenstone type fragmental lava.

The Balaklala rhyolite was also deposited at or near sea level. It is composed of a series of flows, coarse and fine pyroclastic material, and tuff layers, some of which are water deposited; individual flows are less widespread than those of the Copley, probably because the rhyolitic flows were more viscous. Underwater deposition of most of the Balaklala rhyolites is indicated by: water-deposited tuff, the absence of widespread tuff beds, rhyolitic flows immediately overlying pillow lavas, local folding that resulted from syngenetic sliding in the water-deposited tuff beds, a fish plate found in the tuff of the upper part of the Balaklala, and the thick, stubby, irregularly outlined bodies of coarse pyroclastic material. However, waterworn fragments in volcanic breccia and accretionary lapilli tuff indicate that part of the volcanic pile extend above sea level, probably as a group of volcanic islands.

The break between mafic and silicic flows was not abrupt, as a few mafic, Copley-type flows persist into the lower part of the rhyolitic sequence. Several volcanic centers that were the source of the rhyolitic flows and pyroclastic beds of the Balaklala rhyolite have been located, and many others were probably present. The Balaklala rhyolite formed a broad, elongate, volcanic dome about 3,500 feet thick, 16 miles long, and 3 miles wide. The wide areal extent of these siliceous, rhyolitic flows as compared with their thickness is probably due to their eruption from many vents in the early part of the eruptive cycle. The rhyolite in the upper part came largely from a main, centrally located vent.

It seems reasonable to assume that the volcanic complex of Copley and Balaklala rocks did not build a volcanic pile on the ancient sea floor that was equal in height to the total thickness of the two formations (estimated at 7,000–10,000 feet). Compaction of the underlying sediments under load, subsidence due to the outpouring of lava, and isostatic adjustment probably reduced the absolute elevation of the lava. A broad underwater dome or sea ridge of moderate relief along which volcanic islands were scattered must have remained, however, because of the distribution and character of the overlying Kennett formation.

The Balaklala rhyolite is overlain conformably by the Kennett formation of Middle Devonian age. The gradation between the two formations is so complete that the boundary between the formations at some places has been drawn arbitrarily where shale predominates over tuff. Where the Kennett formation was deposited on rhyolitic flows, the contact is sharp, although the volcanic activity did not end abruptly; a few rhyolitic flows and many beds of tuff occur in the lower part of the Kennett, interbedded with shale. The lower part of the Kennett in most areas is composed of black siliceous and cherty shale that contain some radiolaria, and it seems probable that some of the silica was derived from volcanic activity. The middle part of the Kennett is composed of black shale, and the upper part is a bed of coral limestone 250 feet thick. Corals and broken coral debris indicate deposition in relatively shallow water. This is substantiated by the distribution of the Kennett, which was deposited near the outer edge of the volcanic highland, but which is not believed to have covered some of the central parts of the volcanic pile.

Diller (1906, p. 3) believed that an unconformity was present between the Kennett formation and the overlying Bragdon formation because of the boulders of fossiliferous limestone of the Kennett in the conglomerate of the Bragdon, the absence of the Kennett in some areas where the Bragdon rests directly on the underlying Balaklala or Copley rocks, and the difference in the thickness of the Kennett in several parts of the district. Boulders derived from the Kennett indicate erosion of that formation. However, the lack of any evidence of a stratigraphic break between the Kennett and the Bragdon in the West Shasta area leads the writers to conclude that the Kennett was warped following deposition, and that in some area outside of this district the Kennett was raised above sea level and subjected to erosion. The West Shasta area remained under water throughout Kennett and Bragdon time and received continuous sedimentation except where volcanic islands projected above water level during Kennett time.

The range in the thickness of the Kennett is not as great as Diller and others supposed, as described under "Kennett formation," and the difference in thickness is best explained by the deposition of the Kennett against a volcanic pile that had moderate relief, and that in places extended above sea level. It seems more probable that the Kennett did not completely cover local seamounts on the volcanic pile than that it was eroded from these points by subaerial erosion, although erosion by currents, syngenetic sliding, differential compaction, and limited deposition on areas of higher relief probably played a part in forming a deposit of unequal thickness.

The Bragdon formation of Mississippian age, which overlies the Kennett, is a thick deposit of well-bedded shale and conglomerate that was deposited in water of moderate depth. No basal conglomerate is in the Bragdon, which again suggests that there was continuity of sedimentation between the Kennett and the Bragdon in this district.

Following the deposition of the Bragdon a great thickness of sedimentary rocks was deposited, but they have been eroded from the West Shasta district. They have been mapped by Albers and Robertson [3] in the East Shasta area, and by Diller (1906, p. 4–6). These deposits include beds of shale, sandstone, limestone, and volcanic material that range in age from Mississippian to Recent; major gaps in the sedimentary cycle occur between the Brock shale of Triassic age and the Modin formation of Jurassic age, between the Potem formation of Jurassic age and the Chico formation of Late Cretaceous age, and between the Chico formation and the formations of Tertiary age (Diller, 1906, p. 4–6).

Orogeny in this district has been dated as Late Jurassic or Early Cretaceous from several lines of evidence that are given under "Metamorphism," "Mule Mountain stock," and "Shasta Bally batholith." Two plutons, the older of albite granite and the younger of biotite-quartz diorite, were intruded into the rocks of the district; they are both dated as of Late Jurassic or Early Cretaceous age (Hinds, 1934, p. 182–192), but the orogeny and igneous intrusion deformed the rocks into a series of broad folds that contain local areas of close folding. Rocks at many places were foliated, and local shear zones and faults were formed. All the rocks older than the Shasta series were deformed by the orogeny, but the Shasta series was deposited nonconformably on the Shasta Bally batholith.

ROCK ALTERATION

All the rocks older than Late Jurassic have been altered from their original character to some extent by dynamic and igneous metamorphism, hydrothermal alteration, and by weathering. Some of the alteration was described under "Geologic formations" in describing the appearance of the rocks. This is especially true of the description of the Copley greenstone and in describing the contact relationship between the Copley

[3] Albers, J. P., and Robertson, J. F., Geology and ore deposits of the East Shasta copper-zinc district, Shasta County, Calif. U. S. Geol. Survey. Prof. Paper. [In preparation.]

greenstone and the Shasta Bally batholith and the Mule Mountain stock. Also some alteration is described under "Ore deposits" in the description of the hydrothermal alteration related to the ore deposits. The purpose of this section is to summarize the processes that have affected the rocks since their crystallization or cementation, and to give a detailed description of the igneous metamorphism along the edge of the Shasta Bally batholith.

Although the types of metamorphism can at some places be related to particular processes, it is not possible to separate the different types in all instances. Thus hydrothermal metamorphism is superposed on dynamic and igneous metamorphism, and the effect is dependent on stability relationship between the solutions and the different end products of dynamic or igneous metamorphism. Along the borders of the Shasta Bally batholith, igneous metamorphism is superposed on the products of dynamic metamorphism. Likewise, no line can be drawn between late magmatic and hydrothermal alteration in the albite granite. The changes in the composition and the mineral assemblage that occur in the greenstone at its contact with the Shasta Bally batholith are due largely to igneous metamorphism. Much material has been added from the intrusion, and the mineral assemblage at some places is determined almost entirely by the amount of material that has been introduced from the intrusive mass. The mode of introduction, however, includes much more than metasomatism, for it ranges from direct additions of molten material in the migmatites to minor solutions carrying only the constituents of feldspar and quartz or solutions that probably had only a catalytic effect in hastening mineral changes. Thermal and dynamic metamorphism also effected changes in the mineral assemblage.

Deformation accompanying intrusion and the concomitant cataclastic structures that were produced aided the penetration of solutions from the igneous mass over a much larger area than could be effected by purely thermal metamorphism. The heating of the pore solutions and the cataclastic effects due to orogeny and syntectonic intrusions allowed the activated pore solutions to migrate and effect recrystallization. In many parts of the district, the increase of permeability due to dynamic metamorphism has locally influenced other types of metamorphism. It cannot be assumed that different rocks underwent the same degree of metamorphism because they are associated in the field. Whether igneous metamorphism was carried to completion or not, in many localities, depended on a difference in the ease of access of solutions, which is in turn dependent on both primary and secondary structures. Some areas are more foliated than others and during igneous metamorphism solutions worked out along the more foliated bands and emphasized this foliation by mimetic crystallization.

The principal mineralogic changes that are attributable to metamorphism are the formation of minerals of the green schist facies caused by both dynamic and igneous metamorphism, and the formation of amphibolite, gneiss, and migmatite by igneous metamorphism.

DYNAMIC METAMORPHISM

Dynamic metamorphism is defined by Turner (1948, p. 5) as the structural and mineralogic reconstitution of the rocks during deformation caused by orogenic crustal movements. This type of metamorphism has affected all the rocks older than the biotite-quartz diorite of the Shasta Bally batholith.

The effectiveness of dynamic metamorphism varies widely throughout the district. Bands in which the rocks are well foliated are separated by areas in which the rocks are massive, and bands of foliated rock lens out along strike. The preservation of original textures, the lack of cataclastic structures, and the absence of sutured mineral contacts in most of the rocks show the lack of over-all shearing and crushing. The foliation at most localities has a regional trend, although it diverges from this trend in local areas. The alternation of foliated and massive rocks at some places suggests that zones of foliation reflect deep crustal breaks that are represented in the rocks now exposed at the surface by narrow zones of strongly foliated rocks or faults, or by wider zones of less strongly foliated rocks. Recrystallization of primary rock minerals is strongest where movement was limited to narrow zones; where movement was absorbed over a wide zone, oriented minerals are commonly limited to closely spaced planes, and the rock between the planes does not contain oriented minerals.

The grade of dynamic metamorphism decreases from the western part of the Whiskytown quadrangle eastward to the eastern part of the Shasta Dam quadrangle. In the central and western parts of the area the Copley was altered to minerals of the green schist facies, which involves the alteration of the mafic minerals to epidote and chlorite and the alteration of plagioclase to albite and small amounts of epidote, zoisite, and carbonates. The green schist facies is widespread, although both foliated rocks and minerals of the facies are more common in the western than in the eastern part of the district.

Although the Balaklala rhyolite and the sedimentary rocks were folded and locally sheared, they were much less affected mineralogically by orogenic stresses than the Copley. The principal alterations in the Balaklala rhyolite are the occurrence of sericite, hydromica and chlorite in some of the rhyolite.

Sericitization is common through much of the lower and middle units but not in the upper unit of the Balaklala. Some of this alteration may be an early phase of the period of ore deposition, although much of it is widespread and has no apparent spatial relationship to massive sulfide deposits. Where the rhyolite has been foliated and has closely spaced sheeting or schistosity, sericite is formed along the planes of movement. Movement continued after the formation of the sericite, however, as bands of sericite schist are locally crenulated; deformation outlasted recrystallization.

Dynamic stresses possibly caused some of the minor chloritic alteration which gives the rhyolite a light-green color. Feldspar was replaced by chlorite along cleavage planes, and veinlets and irregular patches of chlorite were formed through the groundmass. The chloritic alteration is most intense in the few dacitic flows of the Balaklala, although it is present to some extent in all the Balaklala. Where the chloritic alteration is prevalent, as on South Fork Mountain in the Whiskytown quadrangle, the resulting product is difficult to distinguish in the field from Copley greenstone.

Most of the Balaklala rhyolite is poorly foliated. Fracture cleavage predominates, and ranges from a poorly defined sheeting having little or no alinement of minerals to a quartz-sericite schist in strong shear zones. Bedding-plane foliation is present to a lesser extent and is localized in thin tuffaceous beds and along flow contacts.

Orogenic stresses have had little effect upon the sedimentary rocks except for the formation of local bedding-plane cleavage, fracture cleavage in folds in shale beds, and crumpling in the shale near competent conglomerate beds.

IGNEOUS METAMORPHISM

Igneous metamorphism includes all the mineralogic and textural changes induced in a solid rock by the intrusion of a plutonic body. This includes the metamorphic changes brought about by the heat of the intrusion and those that are caused by fluids—either gaseous or hydrothermal—given off by the intrusive mass during its emplacement and cooling. Both alteration within the igneous rock and of the surrounding invaded rock are included.

Metamorphism of the invaded rocks has been caused by three different intrusive masses—the plugs of coarse-phenocryst rhyolite near the Uncle Sam mine, the Mule Mountain stock of albite granite, and the Shasta Bally batholith of biotite-quartz diorite.

METAMORPHISM RELATED TO PLUGS OF BALAKLALA RHYOLITE

The only alteration definitely related to the Balaklala rhyolite is the intense silicification and quartz-talc alteration around the plugs of coarse-phenocryst rhyolite in the vicinity of the Uncle Sam and Clipper gold mines. The alteration of the plugs and the surrounding nonporphyritic and medium-phenocryst Balaklala rhyolite is most intense in brecciated areas. Apparently preintrusive explosions, possibly of phreatic origin, shattered the rhyolitic rocks, and the coarse-phenocryst rhyolite magma quietly welled up in some of the shattered zones. In these zones, and in associated shattered zones that do not contain magma, late magmatic solutions entered and almost completely replaced the rhyolite, forming irregular areas of highly silicified rock that are conspicuous owing to their resistant character and to their white color.

Although the parent rock contains very little magnesia the solutions must have been rich in it, as local areas of quartz-talc alteration are present along the margins of the plug as well as in the silicified zones. Small lenses and pockets of talc as much as 6 inches long occur near the margins of the silicified coarse-phenocryst rhyolite plug in a road cut 300 feet south of the Uncle Sam gold mine adit, and on the road between the Uncle Sam and Clipper mines. Also, quartz-talc schist occurs near the margin of the plug along the road from the Uncle Sam gold mine to the North Fork of Squaw Creek. Strong cataclastic structures are present in the rhyolite near the margins of the plug, where the rock is replaced by quartz and talc (fig. 33). The magnesia may have been derived from the underlying Copley greenstone.

METAMORPHISM RELATED TO THE MULE MOUNTAIN STOCK

The wall rock of the Mule Mountain stock has been altered in only a few areas, and to a limited extent, but this stock has been widely altered by deuteric or hydrothermal (late magmatic) solutions, and by assimilation of the wall rocks.

In places the Copley greenstone is recrystallized to amphibolite and epidote amphibolite as much as 200 feet from the albite granite contacts. This alteration destroys the texture of the original rock; the altered rock consists of hornblende, epidote, and andesine, but there has been little or no change in chemical composition. The amphibolite is formed by recrystallization of the Copley caused by heat and solutions from the intrusion and has little or no gain or loss of material. On weathered surfaces the recrystallized greenstone is difficult to recognize from the green schist facies of the

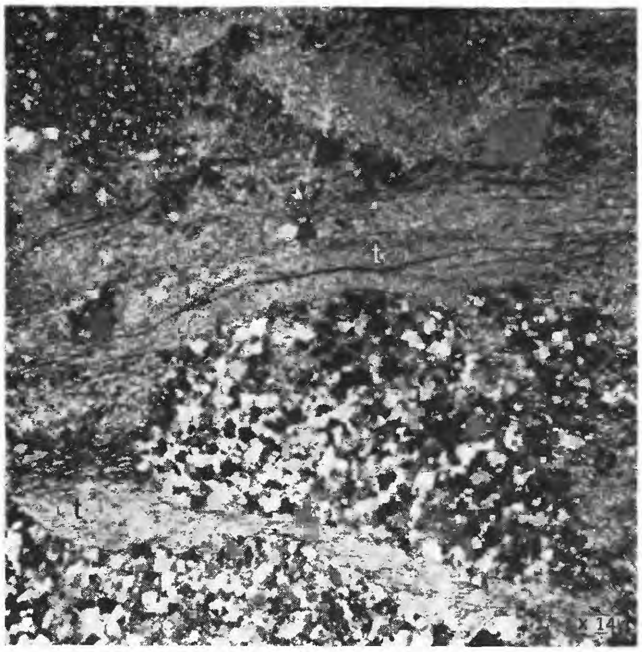

FIGURE 33.—Photomicrograph of extremely sheared coarse-phenocryst rhyolite from the margin of the rhyolite plug near the Uncle Sam mine. The rhyolite is almost completely replaced by quartz (*q*) and talc (*t*), forming a quartz-talc schist. Crossed nicols, × 14.

normal greenstone, but the more granular texture of the amphibolite is readily recognizable on unweathered surfaces.

Near the intrusion of the albite granite the Copley greenstone in some places is silicified and is cut by abundant veinlets of quartz, forming a stockwork of greenstone fragments separated by quartz veinlets.

Much of the alteration related to the Mule Mountain stock is endomorphic. This includes propylitization, albitization, silicification, and changes in the composition of the intrusion where wall rocks were assimilated. Propylitization, which involves the alteration of the mafic minerals to epidote and chlorite and the alteration of the more calcic plagioclase to albite, epidote, zoisite, and carbonates, is confined to the mafic hornblende-quartz diorite part of the Mule Mountain stock. The alteration is only partly completed; most of the plagioclase is altered, but some relict hornblende remains.

Silicification and albitization occur together in irregular areas throughout the leucocratic phases of the stock and are prevalent in brecciated areas within the stock. The alteration is believed to be largely by deuteric solutions. The altered rock chemically is only slightly more sodic and silicic than the unaltered trondhjemitic intrusive mass, but its megascopic appearance and texture are much different. The alteration has caused a bleaching of the rock, a coarsening of grain size, and the formation of a pseudoporphyritic texture containing quartz grains in clusters a quarter of an inch in diameter in a finer grained granitoid matrix. Quartz was the last mineral to crystallize in the unaltered albite granite, so it is difficult to distinguish magmatic quartz from hydrothermal quartz. Thin veinlets of quartz cut the rock and replace plagioclase along grain boundaries. The plagioclase in the bleached, "porphyritic" albite granite is clear and is more sodic than the plagioclase in the unaltered trondhjemite.

METAMORPHISM RELATED TO THE SHASTA BALLY BATHOLITH

The Shasta Bally batholith has had a considerable effect upon the Copley greenstone, which is metamorphosed to amphibolite, hornblende gneiss, and epidote amphibolite as much as 4 miles from the eastern contact of the batholith, and which is altered to gabbro in the borders of the Clear Creek plug of biotite-quartz diorite. The grade of metamorphism decreases eastward away from the contact, but the Copley is altered to amphibolite and hornblende gneiss as far as 4,000 feet from the batholith. The gneiss and the amphibolite grade eastward to epidote amphibolite, farther out to greenstone consisting of albite, chlorite, epidote, and quartz (green schist facies).

Amphibolite facies.—Two distinct rock types are present in the amphibolite facies, which extends as far as 4,000 feet east of the contact of the Shasta Bally batholith. One type is an amphibolite that has been essentially unmodified chemically by the biotite-quartz diorite, and the other is a banded dark quartz-hornblende gneiss or migmatite that, in part, has much added material.

The amphibolite is a megascopically finely crystalline dark-greenish-gray rock that has a uniform foliation striking about N. 20° W. and dipping mostly 60°–70° E. It makes up 75 percent or more of the amphibolite facies and is cut by many thin veinlets of epidote, or quartz and epidote. Locally quartz augen 2 to 3 inches long are abundant. The contact between the biotite-quartz diorite and the amphibolite is sharp, and locally the biotite-quartz diorite transgresses the foliation of the amphibolite.

Under the microscope the amphibolite is seen to be composed of hornblende, biotite, plagioclase, calcite, chlorite, orthoclase, epidote, quartz, and magnetite, and has a crystalloblastic texture. Hornblende and plagioclase make up more than 90 percent of the rock. Hornblende occurs as green, anhedral, acicular crystals averaging 1 millimeter in length and 0.3 millimeter in width in a matrix of clear plagioclase. The acicular crystals are unoriented in the plane of foliation. The hornblende has the following optical properties: optic sign negative; $2V \cong 70°$; indices $n_\alpha = 1.650 \pm 0.003$, $n_\beta = 1.664 \pm 0.003$; $n_\gamma = 1.675 \pm 0.003$. These optical properties correspond to green hornblende. The plagioclase is clear and has very little albite twinning. Its composition ranges from oligoclase to calcic andesine ($Ab_{68}An_{32}$–$Ab_{50}An_{50}$). Biotite, orthoclase, and quartz are present in minor quantities as small interstitial grains.

Locally porphyroblasts of plagioclase or hornblende as much as a quarter of an inch in diameter have formed. They are particularly abundant in the amphibolitic zone at the north end of the small plug on Clear Creek in the Igo quadrangle, but they are rare in the amphibolitic zone along the margin of the Shasta Bally batholith. The porphyroblasts are found in the Copley mainly where it forms a thin cover or roof over the intrusive mass.

The migmatite, which is the second type of rock in the amphibolite facies, is best exposed in the Igo quadrangle in the Brandy Creek area and near the small plug on Clear Creek. In the Brandy Creek area, where gneissic structure of the migmatite parallels the intrusive contact, three lithologic zones are recognized in the transition from Copley greenstone to biotite-quartz diorite. The outermost zone is a dark hornblende gneiss that is extremely well banded and has alternating dark and light bands one-eighth inch to 1 inch thick. The bands at first glance give the appearance of being regular and straight, but in detail they are lenticular, and none can be traced for more than 10 feet. The dark hornblende gneiss, which has had little added material, grades toward the biotite-quartz diorite contact to leucocratic gneiss that has only 5 to 10 percent mafic minerals, and has had much added material. The leucocratic gneiss grades to foliated biotite-quartz diorite within the border of the batholith. The first two units constitute a migmatite zone between Copley greenstone and the Shasta Bally batholith.

The outermost zone, of dark hornblende gneiss, consists of hornblende, plagioclase, quartz, chlorite, biotite, epidote, orthoclase, magnetite, apatite, and sphene, listed in order of decreasing abundance. It has well-formed cataclastic structures now partly obscured by recrystallization. The dark bands contain more than 50 percent hornblende and a lesser amount of biotite and chlorite. The light bands contain mostly quartz and plagioclase ($Ab_{50}An_{50}$) and some orthoclase. Mortar structures and augen are common in localized areas, and there is a marked lineation shown by acicular hornblende crystals and by plagioclase and quartz augen. The grain size averages about 0.2 millimeter.

The augen in the dark bands in the hornblende gneiss, as in the amphibolite, are probably formed by metamorphic differentiation. The augen contain the same minerals as the amphibolite, but they have a greater concentration of the light minerals. Although the acicular amphibole crystals bend around some of the quartz augen, in most of them the amphibole needles project into the augen with no deflection but in much less concentration. Some of the amphibole crystals within the augen have a tendency to be oriented toward the center.

The mineralogic character of the light bands in the hornblende gneiss are identical to the dark bands, but they have a concentration of light-colored minerals. There is no change in grain size between the dark and light bands, and they have interlocking grains at the contact. The plagioclase has the same composition ($Ab_{50}An_{50}$) in both bands. Apparently the light and dark bands and the augen both formed by metamorphic differentiation (Eskola, 1932, p. 68–77; McCallien, 1934, p. 11–27; Read, 1933, p. 317–328; Stillwell, 1918, p. 354–396; Turner, 1941, p. 1–16, 1948, p. 137–147) in the hornblende gneiss, with possibly small addition of material, in contrast to the leucocratic gneiss in which much granitic material has been added.

In a few places there has been lit-par-lit injection of fine-grained hornblende diorite, quartz latite porphyry, and aplite dikes parallel to the banding of the gneiss. These injected bands have sharp contacts with the dark and light bands of the hornblende gneiss and have many crosscutting relationships with the hornblende gneiss.

Toward the contact of the Shasta Bally batholith the dark hornblende gneiss in the migmatite zone grades into the second zone, in which the color is lighter and the banding is much less pronounced. This leucocratic gneiss contains 80 to 90 percent quartz and plagioclase, and small amounts of orthoclase, hornblende, biotite, epidote, muscovite, magnetite, apatite, and sphene. Alinement of hornblende, biotite, and chlorite, and quartz-plagioclase augen impart a pronounced lineation to the rock, although the mafic minerals are uniformly distributed through the rock in contrast to the concentration of mafic minerals in bands in the hornblende gneiss. The grain size is generally coarser in the leucocratic gneiss and ranges from 1 to 2 millimeters across where the rock is not sheared. Adjoining bands of quartz grains may have very different grain sizes, but all the grains in a single lens or band are nearly the same size, as described by Sander (1938, p. 29). Cataclastic structures are much more strongly developed in the leucocratic gneiss than in the dark hornblende gneiss. Mortar structure, porphyroclasts, and local formation of mylonite are common. Cataclastic structures are mostly healed by quartz. Orthoclase and muscovite are late minerals, and in part fill fractures that cut across both foliated and cataclastic structures, which are parallel.

Plagioclase in the leucocratic gneiss has a composition of $Ab_{62}An_{38}$, which is about the same as that of the cores of the plagioclase in the biotite-quartz diorite (p. 49), in contrast to the more calcic plagioclase ($Ab_{50}An_{50}$) in the hornblende gneiss. The plagioclase thus becomes less calcic with the decrease in proportion of mafic minerals. Also, in the gradation from dark hornblende gneiss to leucocratic gneiss, hornblende decreases and quartz increases. Cataclastic structures increase, and much of the leucocratic gneiss is a mylonite. Chlorite, epidote, and some of the amphibole in the mylonitic bands are late minerals and wrap around quartz and plagioclase porphyroclasts.

The leucocratic gneiss zone is exposed for a width of 100 feet in Brandy Creek, and grades to hornblende gneiss away from the batholith. Toward the batholith there is a covered zone for 125 feet; then an outcrop of coarse-grained biotite-quartz diorite showing primary foliation. The foliated biotite-quartz diorite contains about 40 percent biotite and hornblende, whereas the leucocratic gneiss contains less than 10 percent mafic minerals, most of which are biotite, chlorite, and epidote. About 1,000 feet north of Brandy Creek the contact between leucocratic gneiss and foliated biotite-quartz diorite is sharp and conformable.

Epidote amphibolite facies.—The amphibolitic rocks grade to the east, away from the biotite-quartz diorite, to greenstone containing epidote and fibrous amphibole, and finally to chloritic lavas. The greenstone containing fibrous hornblende are light to dark green hard, locally massive rocks and are epidote amphibolites. They are usually lighter in color and are harder and more massive than the chloritic lava. Locally they are cut by abundant veinlets of quartz and epidote or have a lumpy appearance due to formation of knots of epidote the size of a fist. Quartz augen 2 to 3 inches long are common.

Under the microscope the epidote amphibolite is seen to contain plagioclase, fibrous amphibole, chlorite, epidote, zoisite, calcite, quartz, pyrite, and magnetite. Plagioclase is slightly more calcic than that in the chloritic lava, but it is more sodic than that in the amphibolitic facies. It ranges in composition in the epidote amphibolite from $Ab_{90}An_{10}$ to $Ab_{72}An_{28}$. Plagioclase in the chloritic lava ranges from $Ab_{97}An_3$ to $Ab_{78}An_{22}$; that in the amphibolite facies from $Ab_{68}An_{32}$ to $Ab_{50}An_{50}$. The amphibole is fibrous to acicular and grains are commonly less than 1 millimeter long. Locally the amphibole is oriented, but more commonly it is unoriented in plagioclase that may either be massive or have a felted texture. The amphibole is colorless to light green and has the following optical properties: Optic sign negative $2V \cong 80°$; $Z \wedge c = 14°$; and indices of refraction $n\alpha = 1{,}603 \pm 0.003$; $n_\beta = 1.615 \pm 0.003$; $n\gamma = 1.625 \pm 0.003$. These optical properties fit tremolite most closely. Epidote is present in clusters commonly 1 to 2 millimeters but as much as 3 inches in diameter, in veinlets with quartz cutting the rock, and as amygdule fillings. Quartz occurs as amygdule fillings and as veinlets cutting the rock. The amount of amphibole decreases to the east away from the biotite-quartz diorite contact, and the amount of chlorite increases toward the east. The texture changes from crystalloblastic in the amphibolite and epidote amphibolite to pilotaxitic in the green schist facies.

The Copley greenstone has been altered by hydrothermal emanations from the biotite-quartz diorite and to a lesser extent by lit-par-lit injection. The amphibolite was formed by recrystallization of chloritic lava—aided by heat and fluids from the biotite-quartz diorite intrusive mass—with essentially no change in chemical composition. The amphibolite consists mainly of andesine plagioclase and hornblende, although it contains a little interstitial orthoclase and quartz, which suggests that there has been a little potash and silica metasomatism.

In some places, particularly along the contact of the Clear Creek plug, intrusive breccia formed at the contact between biotite-quartz diorite and amphibolite or migmatite. The fragments in the intrusive breccia include all the rock types in the amphibolite or migmatite zone.

The leucocratic gneiss could not have been formed by lit-par-lit injection and assimilation of Copley greenstone as it is more felsic than either the biotite-quartz diorite or the Copley greenstone. Some lit-par-lit injection is recognized, but it is on a small scale. The dikes injected into the gneiss are generally parallel to the gneissic structure, but they have sharp contacts with the gneiss and some have crosscutting relationship.

The migmatite was probably formed by fluids given off ahead of the intrusion that reworked and replaced the Copley greenstone. This conclusion is the same as that of Sederholm (1926, p. 131) and Stark (1935, p. 1) in their descriptions of migmatites from Finland and Colorado. The replacement of Copley is nearly complete in the leucocratic gneiss zone, and there are all gradations in the replacement from dark hornblende gneiss to leucocratic gneiss. The metasomatism was done by solutions rich in silica, soda, and potash to form a gneiss that is more felsic than either the host rock or the parent magma.

The crushing of the migmatite was protoclastic, similar to that described by Waters and Krauskopf at the border of the Coville batholith (Waters and Krauskopf, 1941, p. 1412). Although the migmatite probably formed slightly before the intrusion of the Shasta Bally batholith, the minerals in the migmatite are the same as those in the biotite-quartz diorite. The migmatite was crushed by the intrusion of the batholith, and the cataclastic structures were healed by silica-rich solutions from the batholith. Quartz and plagioclase are in part crushed and drawn out, but these cataclastic structures are transgressed and healed by quartz and plagioclase which are similar in appearance to the crushed minerals, and are about the same age.

The contact of migmatite and amphibolite with the Shasta Bally batholith is generally sharp, although in a few places intrusive breccias were formed with a matrix of biotite-quartz diorite. The intrusive breccias are described under "Shasta Bally batholith." Ptygmatic folding occurs in the migmatite at a few places, but generally the wall rocks of the intrusion were not plastic.

The sharp, in part transgressive, contact between biotite-quartz diorite and migmatite and amphibolite, and the incorporation of fragments of migmatite and amphibolite in the intrusive breccias at the border of the intrusion, indicate that these rocks were formed prior to the intrusion. The intimate relationship between crushing of the migmatite, healing and transgression of cataclastic structure by minerals identical to those within the intrusion, and the local lit-par-lit injection indicate the approximate consanguinity of the migmatite and biotite-quartz diorite intrusion. An explanation for the local transgressive relationship and for the great range in the width of the amphibolite, gneiss, and migmatite zones might be, as Durrell (1940, p. 108) suggested for the granitic rocks of the Visalia quadrangle, that the magma continued to advance while cooling so that the contact aureole formed at the period of highest temperature would be locally destroyed by the advancing magma.

Where the Clear Creek plug is in contact with the Copley, the greenstone has been reconstituted to form the rock units listed below. These changes in the greenstone are well exposed along the eastern contact and in the roof of the plug in Clear Creek canyon. The rock units that are exposed in Clear Creek, listed roughly in paragenetic sequence, are:

1. Greenstone recrystallized to fine-grained amphibolite with little added material.
2. Hybrid rock containing fine-grained, recrystallized greenstone with porphyroblasts of hornblende and stringers and lenses of hornblendite.
3. Fine-grained hornblende gabbro containing porphyroblasts of hornblende a quarter of an inch in diameter.
4. Dikes and veinlets of coarse-grained hornblendite.
5. Coarse-grained, nodular hornblende gabbro containing 40 to 50 percent rounded hornblende crystals three-eighths of an inch in diameter.
6. Banded gabbro and gabbro showing "unconformities."
7. Intrusive breccia containing fragments of units 1–6 in a matrix of hornblende-quartz diorite and hornblende diorite.

The first alteration of the greenstone was recrystallization to a fine-grained amphibolite and epidote amphibolite, and the formation of knots of epidote. This alteration was a simple recrystallization of the greenstone to amphibolite containing hornblende and andesine or labradorite, and little or no material was added. The alteration is similar to that along the southern margin of the Shasta Bally batholith. The amphibolite ranges from several hundred to a thousand feet in width along the northeast contact of the plug.

Near the intrusion, porphyroblasts of hornblende that average about 5 millimeters in diameter were formed along fractures in the amphibolite (fig. 34). Further recrystallization resulted in formation of lenses and veinlets of hornblende gabbro, which locally were mobilized and have intruded amphibolite. In places the amphibolite has recrystallized to form a fine-grained nodular gabbro containing 20–25 percent of spherical aggregates of hornblende crystals in a matrix of hornblende and labradorite that average about 0.3 millimeter.

This fine-grained nodular hornblende gabbro is intruded by coarse-grained nodular hornblende gabbro composed of 50 percent spherical aggregates of hornblende crystals 5–10 millimeters across and interstitial labradorite averaging 1 millimeter in length. The nodular gabbro has the same mineralogy as the nodular hornblende gabbro, and apparently resulted from the further recrystallization and some remelting, which

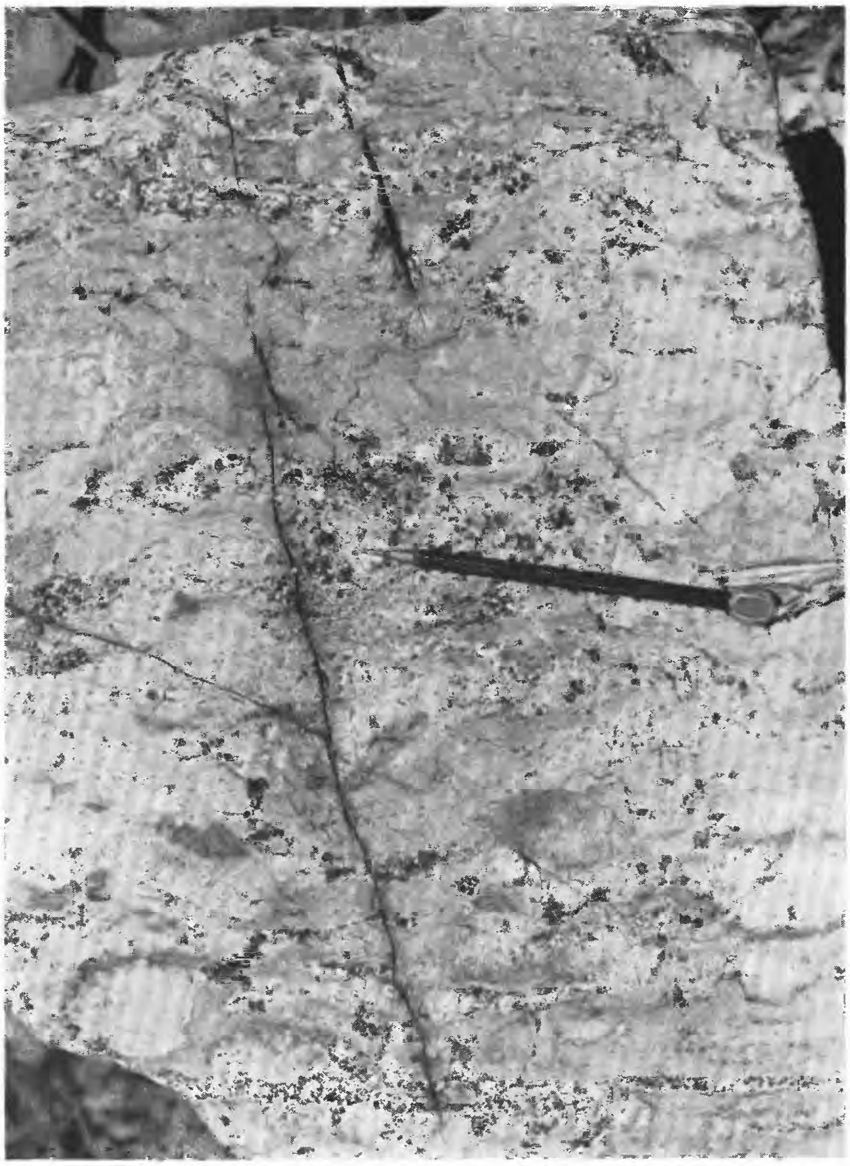

FIGURE 34.—Formation of porphyroblasts of hornblende in hybrid Copley greenstone near contact with the Clear Creek plug. The coarser stringers have the composition of hornblende gabbro.

caused mobilization of the matrix and intrusion of the nodular gabbro as a thick melt or mush. The spherical aggregates of hornblende crystals in the nodular hornblende gabbro are recrystallized to large single hornblende crystals or are party recrystallized to an aggregate of several large ones. This melt or mush of nodular gabbro filled fractures in the nodular hornblende gabbro and amphibolite, and also formed small plugs of nodular gabbro in the amphibolite. Some small local pegmatitic lenses in the nodular gabbro have euhedral hornblende crystals 1 to 2 inches long in coarse plagioclase. These mafic pegmatitic lenses are erratically distributed through the nodular gabbro.

Some of the hornblende gabbro has a well-formed banding (fig. 35). Dark bands 1–2 inches thick, that are predominantly of hornblende, are separated by light bands a quarter of an inch thick that are predominantly of plagioclase. The grain size averages about 2½ millimeters in diameter. The banding is not uniform, but bends in all directions.

The banding in the gabbro may be due either to separation of felsic and mafic minerals during the emplacement and crystallization of the gabbro, or to the formation of the banded gabbro in place by recrystallization of a banded migmatite similar to the one at Brandy Creek described under "Igneous metamorphism." The latter seems more likely to the writers, as the banded gabbro has some relicts locally of fine-grained hornblende gabbro similar to the matrix of much of the recrystallized greenstone.

FIGURE 35.—Fragment of banded hornblende gabbro cut by hornblendite dikelet. The banded gabbro is a large fragment in an intrusive breccia with a matrix of hornblende diorite.

In a few places the bands in the hornblende gabbro show several "unconformities" (fig. 36). The "unconformities" were probably formed while the gabbro was still in a plastic state. Eroding of the banded gabbro by the magma and the concomitant fragmentation and rotation of the banded gabbro by stresses from the intrusion probably caused these "angular unconformities" and formed the intrusive breccias. The banded gabbro is intruded by the fine-grained hornblende-quartz diorite of the Clear Creek plug. There has been some lit-par-lit injection of hornblende-quartz diorite bands 1 to 2 inches thick parallel to the layering in the gabbro, and some of the hornblende-quartz diorite bands cut across the layering in the banded gabbro.

FIGURE 36.—Fragment of hornblende gabbro showing "angular unconformities" cut by a dikelet of hornblendite. Scale shown by 50-cent piece.

Dikelets of hornblendite cut the hybrid rocks and the gabbro. They range from small lenses several inches long to dikelets as much as 2 inches thick and several tens of feet long. The hornblendite dikelets occur only in the hybrid zones about the Clear Creek plug and in Brandy Creek near the margin of the Shasta Bally batholith. These dikelets could be one of the mobilized products formed by the recrystallization of the greenstone to hornblende gabbro and separation of the hornblende and plagioclase either by gravity or by squeezing.

Intrusive breccias that contain fragments of amphibolite, banded gabbro, hybrid greenstone, hornblendite, and hornblende gabbro in a matrix of hornblende-quartz diorite (fig. 37) occur at the margin of the Clear Creek plug. The fragments of the hybrid rock and gabbro in the intrusive breccias date the formation of the gabbro as slightly earlier than the intrusion of the plug. Also some bands of hornblende-quartz diorite similar to the rock in the margin of the plug cut the banded gabbro. The spatial relationship of the gabbro and hybrid rocks to the margin of the plug dates their formation just before or during the emplacement of the plug.

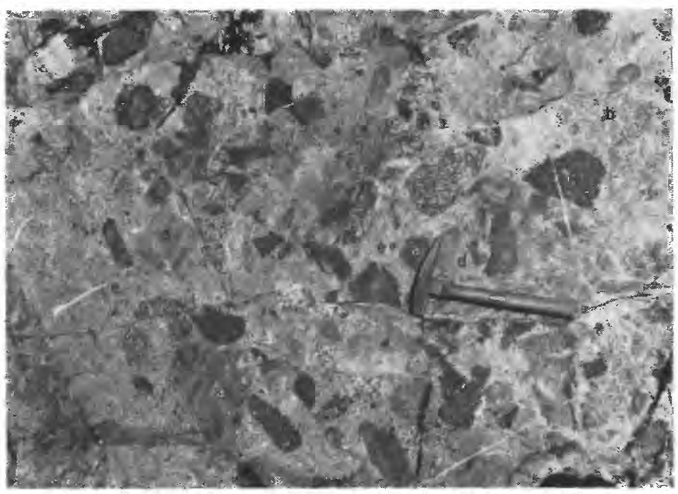

FIGURE 37.—Intrusive breccia containing fragments of gabbro, hornblendite, and Copley greenstone in a matrix of hornblende diorite; greenstone (*a*), gabbro (*b*), hornblendite (*c*), silicified greenstone (*d*), hornblende diorite matrix (*e*).

The ground was under strong stress at the time of the formation of the hybrid rocks and of the nodular gabbro, probably due to the force of the intrusion, and the amphibolite and hybrid rocks were shattered and contorted in many places. This shattering was probably augmented by the buttressing effect of the Mule

Mountain stock, as the shattered amphibolite and hybrid rock lie between the plug of biotite-quartz diorite and the earlier stock of albite granite; this shattering permitted thorough soaking of the greenstone by solutions emanating from the intrusion.

WEATHERING

Weathering has not been an important factor in the alteration of the rocks except in the Mule Mountain stock of albite granite and to a lesser extent in the Copley greenstone. The feldspar in the greenstone is altered to clay minerals at the surface, and weathering emphasizes the schistosity. Weathering has had little effect upon the nonporphyritic and medium-phenocryst Balaklala rhyolite, but the coarse-phenocryst Balaklala rhyolite at the surface has a "punky" look due to alteration of the feldspar to clay minerals. Both kaolinite and nontronite are present. Kaolinite is prevalent as an alteration of the albite phenocrysts and is abundant in the groundmass of the weathered rhyolites. Nontronite commonly forms cross-fiber veinlets in fractures in quartz phenocrysts. Feldspar phenocrysts in the porphyritic rhyolite are altered to clay minerals and blend with the groundmass and thus are less prominent than quartz phenocrysts.

The albite granite is deeply weathered. Road cuts that are 50 feet deep near the town of Shasta on U. S. Highway 299W are entirely in white crumbly, disintegrated albite granite. Mechanical disintegration has been more active than chemical decomposition in the weathering of the rock, and much of the feldspar in the disintegrated rock appears to be little altered. Because of the deep disintegration, unweathered exposures of albite granite are present only in the deep canyons as in Clear or Spring Creeks. The other intrusive and sedimentary rocks were very little affected by weathering.

The massive sulfide deposits have been oxidized to a gossan capping that ranges in thickness from a few feet at the Balaklala and Mammoth mines to a maximum of 400 feet along the footwall of the Camden fault at the Iron Mountain mine. The mineralogy of the oxidized sulfides is discussed under "Ore deposits."

ORIGIN OF THE ALBITE

The rocks in the West Shasta district are a typical spilitic suite as described by many writers including, among others, Benson (1915, p. 121–173), Dewey and Flett (1911, p. 202–209, 241–248), Gilluly (1935, p. 225–252, 336–352). Peach and others (1911), and Sundius (1930, p. 1–17). The rocks of the spilitic suite that occur in this area are spilite, keratophyre, soda-rich rhyolite, albite diabase, and albite granite.

Field and microscopic evidence indicates that the albite in the soda-rich rocks of the West Shasta district was formed at different times and by several processes; both primary and secondary albite are present. All of the evidence that was seen indicates that the albite in the sodic Balaklala rhyolite is primary. In the greenstone some of the albite is known to be secondary, and the evidence for a primary origin of the remainder is less conclusive than in the rhyolite. Albite in the albite granite is secondary, and the rock formed by the albitization of trondhjemite. Field evidence is conclusive that albitization of the trondhjemite is later and is unrelated in time to the formation of albite in the Balaklala rhyolite and Copley greenstone. The Copley greenstone was composed predominantly of albite and chlorite at the time of intrusion of the Mule Mountain stock, and it was altered by contact metamorphism to an amphibolite containing andesine and hornblende as much as 200 feet from the contact.

The plagioclase in the phenocrysts and in the groundmass of the Balaklala rhyolite is uniformly albite, as is the plagioclase in rhyolitic tuff, and no relicts of a more calcic plagioclase core are present. The composition of the albite in the phenocrysts and in the groundmass is about the same and ranges from $Ab_{96}An_4$ to $Ab_{92}An_8$. The albite in the groundmass has a felted texture in most of the rhyolite, although trachytic textures and spherulitic structures are present. Trachytic texture and fluidal structures are reflected by unaltered albite laths in the groundmass of the nonporphyritic rhyolite. The albite in the unaltered, unsheared rhyolite is in clear laths that have sharp albite and Carlsbad twin lines and commonly have no inclusions. Broken fragments of plagioclase in crystal tuff are uniformly albite.

Thin lenticular dacitic flows and dacitic tuff layers are interbedded in the rhyolite in some areas. The plagioclase in the dacite is unzoned oligoclase and has composition $Ab_{80}An_{20}$, but the plagioclase in the adjacent rhyolite is albite. It would be difficult to account for the difference in composition of the plagioclase in adjoining rocks by albitization that was so thorough and widespread that it removed all traces of an originally more calcic plagioclase in cubic miles of rhyolite.

At a few localities the rhyolite is cut by sharp-walled quartz-albite veinlets. The material in the veinlets is coarser grained than in the groundmass, but the texture and composition of the groundmass is unchanged. The albite in the veinlet is introduced, but there is no evidence that it is other than primary in the groundmass. The albite in the veinlets is $Ab_{96}An_4$, which is the composition of the most sodic albite in the groundmass.

Myrmekitic and micrographic textures are not uncommon in the porphyritic rhyolites. The time of origin of these intergrowths is not known with certainty, as micrographic intergrowths of quartz and albite are found mainly in isolated albite phenocrysts surrounded by a quartz-albite groundmass having a pilotaxitic texture, but they also occur localized near the grain boundaries of quartz phenocrysts and in isolated patches in the groundmass. A few micrographic intergrowths, which were observed near quartz-albite veinlets, were probably formed at the same time as the veinlets, but some euhedral albite crystals having a micrographic texture were deposited in crystal tuff; the micrographic intergrowth apparently formed before the crystal was incorporated in the tuff as nearby crystals of albite contain no myrmekite. No myrmekite is present in the matrix of tuff. Similar isolated euhedral albite phenocrysts that have micrographic intergrowths in porphyritic rhyolite apparently formed during crystallization of the rhyolite.

The evidence is not conclusive whether the albite in the keratophyre and spilite flows and tuff beds is of primary or secondary origin. The composition of the plagioclase ranges from $Ab_{97}An_3$ to $Ab_{78}An_{22}$. Most of the albite is cloudy due to inclusions of chlorite and, in places, epidote or zoisite, and the plagioclase is either equant or lath shaped. Secondary albite occurs where some veinlets of quartz and albite cut the keratophyres, amygdules are present partly filled with albite, and some of the feldspar has saussuritic alteration. However, it is doubtful that the original plagioclase was more calcic than oligoclase, as the percentage of epidote and zoisite inclusions is too small to account for albite or sodic oligoclase by saussuritic alteration of original andesine or labradorite, which composition would be expected if the rock were originally an andesite or basalt. Also, a few specimens of variolite are only slightly altered and have spherulitic growths of oligoclase of composition $Ab_{85}An_{15}$ and minor interstitial small anhedral grains of quartz.

Some keratophyres with pilotaxitic texture have clear laths of plagioclase that range in composition from $Ab_{90}An_{10}$ to $Ab_{85}An_{15}$. These plagioclase laths have both albite and Carlsbad twinning. A finely porphyritic spilite from near Shasta Dam has clear, unzoned plagioclase phenocrysts with Carlsbad and albite twinning in a groundmass of clear plagioclase laths. Both the phenocrysts and the groundmass have a composition of $Ab_{93}An_7$. The plagioclase in these rocks is probably original.

The albite granite of the Mule Mountain stock formed by albitization of trondhjemite by late magmatic solutions that penetrated large areas of the stock in the most sheared parts and added soda and silica. The albite in these areas is clear or is altered to sericite. In the parts of the stock least altered by late magmatic solutions, the albite is clouded by abundant inclusions of epidote, zoisite, and sericite; this cloudy albite is embayed and replaced by quartz and clear albite in the areas altered by late magmatic solutions. Micrographic and myrmekitic textures locally are common in these altered parts in the stock, but are rare in the unaltered parts of the stock.

Primary and secondary albite are both evident in the rocks of the West Shasta district. Secondary albite is indicated in crosscutting quartz albite veinlets, in amygdules filled with albite, and in micrographic and myrmekitic intergrowth controlled by grain boundaries. Evidence for the primary origin of most of the albite is less direct; it is indicated by the uniform composition and texture of the albite throughout wide areas of Balaklala rhyolite, the occurrence of oligoclase instead of albite in dacitic rock interlayered with rhyolite, the lack of replacement textures or crosscutting veinlets in most of the rhyolite, the secondary relationship of replacement albite where it is present, broken crystals of albite and of fragments of micrographic intergrowths in tuff where the albite apparently has been formed before its incorporation into the tuff, and preservation of primary textures and structures in rhyolite and greenstone. It is improbable that albitizing solutions derived from one rather limited source, such as the albite granite, could account for the complete alteration of all feldspar in a wide variety of rocks over a large area. The evidence favors the derivation of the albite from several, possibly related, sources. The primary albite is related to a soda-rich magma chamber that probably is related to a geosynclinal environment. Dikes and sills of rhyolite both earlier and later than the flows of Balaklala rhyolite contain albite of the same composition and texture as the rhyolite of the flows, which suggests that all came from the same magmatic source. The secondary albite is derived in part from soda-rich volcanic emanations and from late magmatic solutions. The albitic spilites, keratophyres, and soda rhyolites are volcanic flows that were predominantly of submarine origin, and it is probable that part of the soda content of some of these rocks was derived from sea water. Whether the magma chamber was contaminated by absorption of salt water, whether the intrusive rocks were contaminated by entrapped salt water in the wall rocks during the process of emplacement, or whether the intrusive rocks absorbed soda from sea water after extrusion on the sea floor is speculative.

A group of soda-rich, silicic rocks that may be related to a single intermittent or long-enduring magma chamber is an interesting feature in the West and East Shasta districts. The latter district includes the Afterthought, Bully Hill, and Rising Star mines about 14 miles east of the mines in the West Shasta district. The rocks are the Balaklala rhyolite, the albite granite, the Bully Hill rhyolite, and the associated sodic mafic rocks that occur in both districts. Igneous and sedimentary layered rocks formed in a geosynclinal environment that lasted from pre-Middle Devonian to Middle Jurassic time. The Balaklala rhyolite is of Middle Devonian age and the Bully Hill rhyolite is of probable Triassic age yet the two rocks are essentially identical in appearance and mineralogy. The albite granite was intruded in Late Jurassic or Early Cretaceous time. Probably a relationship is indicated here between a geosynclinal environment, sodic igneous rocks, and sodic emanations. Evidence may indicate a magma chamber of exceedingly long duration that contained rocks of an unusual and distinctive type. The albite granite may be an intrusion of the residual magma that remained after the extrusion of the Balaklala rhyolite and Bully Hill rhyolite.

BASE-METAL DEPOSITS

GENERAL FEATURES

The copper-zinc ore deposits of the West Shasta district occur as massive pyrite bodies that contain chalcopyrite and sphalerite, and some gold and silver. The most striking features of the ore bodies, which have a brassy metallic appearance, are the absence of megascopic gangue minerals and the sharp boundaries against barren or weakly pyritized wall rock. The insoluble material consists of quartz, sericite, clay minerals, and unreplaced rock, and ranges in content from as little as 3.5 to as much as 35.5 percent, probably averaging about 15 percent. The massive sulfide ore is commonly separated from barren wall rock by a thin selvage of gouge; gradational contacts between ore and wall rock are rare.

In a few areas, usually near or adjacent to bodies of massive pyritic ore, chalcopyrite occurs as veinlets and disseminated grains in the rock in sufficient quantity to be minable. Pyrite is subordinate to chalcopyrite in these areas, and the ore consists predominantly of quartz and chalcopyrite.

Most of the ore bodies of the district are lenticular and flat lying and their greatest dimensions are in a horizontal plane; several are saucer shaped; one is domical; and one is synclinal. Although steeply dipping bodies are not characteristic of the district, they do occur at the Hornet mine at Iron Mountain, and at the Golinsky and Sutro mines.

At the Iron Mountain mine the massive sulfide ore bodies, before postmineral faulting, ranged in size from a maximum of about 4,500 feet in length, several hundred feet in width, and more than 100 feet in thickness to small bodies having maximum dimensions of only a few feet. Most of the individual ore bodies that have been mined were several hundred feet wide and 20 to 50 feet thick. Many ore bodies are discrete lenses in a broad mineralized zone, but at some mines an originally continuous ore body has been offset by postmineral faulting into separate blocks of ore; these are mined separately and are named as individual ore bodies.

HISTORY OF THE MINING DISTRICT

Interest in the mineral possibilities of the Shasta area was first aroused by the discovery of gold placers in Clear Creek at a point about 8 miles southwest of the present city of Redding. Major P. B. Reading discovered gold in Shasta County in March 1848 following his visit to the newly discovered gold locality at Sutter's Mill in El Dorado County. Gold was mined in Shasta County as early as 1849 by emigrants traveling along the Lassen trail. Reading Springs, named Shasta in 1850, was the terminal point of one of the western routes to California that skirted Mount Lassen on the north and crossed the Sacramento River at a ford near Redding. The discovery of rich placer gold led to an influx of prospectors and miners soon after the discovery. The earliest settlement at that time was the town of Shasta, which is 3 miles west of Redding.

As the interest was primarily in the rich deposits of placer gold, little attention was paid by miners to deposits of base metals that were found while prospecting for gold. Transportation difficulties and the lack of treatment plants in the early days of the district ruled out the mining of any minerals other than precious metals. Copper ore in Shasta County was known at an early date, however, as Aubury (1902, p. 24) reports that Langley's State Register for 1859 says, in part, regarding the copper resources of the state:

> The ore from the vicinity of the Pitt and McCloud Rivers, Shasta County, is said to excell in richness the celebrated Arizona mines, and to contain in addition a considerable quantity of gold.

Copper mining along the foothill copper belt of California was begun in 1860 and Aubury (1902, p. 25) reports:

> The copper excitement thus started quickly spread, and in a few months it filled the State. The period of 1862–63 was marked by a speculative mania, the organization of hundreds of copper mining companies, and the wildcat exploitation of slight surface prospects.

In spite of the information on the presence of copper in Shasta County in Langley's early report, the copper mining boom of the 1860's did not result in the exploitation of any copper mines in the West Shasta district, although some copper ore was mined in the East Shasta district as early as 1865. Miners searching for gold made sporadic attempts to mine some of the outcrops of the base-metal ore bodies that were enriched in gold, but without much success, and Aubury (1902, p. 31) reports:

> In 1893, but two years before the beginning of the career of the Mountain Copper Company, copper was not even mentioned in a review of the mineral resources of Shasta County in a local paper.

However, the enormous gossan at the Iron Mountain mine attracted attention and in the early part of the sixties the mine was acquired by William Magee and held for its possible future value as an iron ore.

The discovery of silver ore in the gossan at Iron Mountain in 1879 created a great deal of local excitement. Aubury reports (1902, p. 36):

> This discovery was soon noised abroad and a characteristic stampede to the region ensued. The popular effect is well shown in a news letter to the Mining and Scientific Press from a Whiskeytown correspondent in June 1880. He writes in part: "At this particular time, in this part of Shasta County, the silver boom is up high, and such expressions as 'the most extensive and the richest silver ledge the world has ever seen' are frequent. Some five or six miles from the ancient town of Shasta was known to exist what was called Iron Mountain. Nothing was expected of it and no one prospected there. A curious expert came from the city and has been secretly looking at its formations, assays have been made of his finds, and now the whole country is wild and claims are staked off for miles. A new silver belt has been discovered, the assays of which go away up into the hundreds."

A plant for extracting silver from the gossan at Iron Mountain was built in 1879 by James Sallee, William Magee, and Charles Camden, and some intermittent mining was done in the silver-rich parts of the gossan from 1879 to 1895. The size and importance of the base-metal ore deposits were first realized in 1895, when prospecting was begun on the base-metal ores that had been found during the mining for silver. The Mountain Copper Co., Ltd. began mining copper ore in 1897; this operation revived prospecting for copper, and within the next decade most of the major mines in the West Shasta district had been discovered.

Transportation of ore and supplies has been a considerable expense to the mines in the West Shasta district because of the steep topography. The ores at Iron Mountain were first hauled to the main line of the Southern Pacific Railway by a branch-line railway that was completed in 1896. The branch line was abandoned in 1922 and since that time the ore has been taken from the mine to the railway by an aerial tram.

Tucker reports (1926, p. 153):

> The ore from the Keystone and Stowell mines was transported by aerial tram to bunkers at the Balaklala mine, then by aerial tram to Coram, from which point it was hauled by train over the Southern Pacific tracks to the smelter at Kennett. Ore from the Sutro mine was trammed by mules to bunkers at the Mammoth mine. The ore from the Mammoth mine was drawn from mine chutes on the 500-foot level, into narrow-gauge railroad cars, and taken over a 2-mile electric railroad equipped with two 25-ton electric locomotives, six 25-ton steel gondola cars, and nine 10-ton flat cars to ore bins with a capacity of 1000 tons. From these bins the ore was taken over an incline gravity railroad in skips to another set of bins. The gravity railroad has a length of 4,000 feet and a drop in this distance of 1,700 feet. The skips have a capacity of 20 tons of ore and travel at a speed of 2,000 feet a minute. From these lower bins the ore was taken to the smelter over a steam railroad operated with three 40-ton locomotives and 22 standard steel railroad cars. The capacity of the transportation system is about 1,500 tons of ore per day.

The first smelter in Shasta County was erected in 1875 in the East Shasta district. In the West Shasta district, the smelter built by the Mountain Copper Co., Ltd. began operations at the town of Keswick on the Sacramento River in 1896, and at the peak of its operations in 1904 was treating 1,000 tons of ore per day. The sulfur content of the Iron Mountain ore was first reduced by heap roasting in piles of ore that were scattered along the railway and fired by cordwood. As much as 350,000 tons of ore was burning at one time in heaps, before charging into the blast furnace.

Sulfur was not recovered from the ores of the Iron Mountain mine until about 1906. Inasmuch as the market for sulfuric acid was in the San Francisco Bay area, the Keswick smelter was dismantled in 1906 and moved to Martinez on the bay.

A smelter for treating ores from the Balaklala mine was built in 1906 at Coram, which is about 1 mile below Shasta Dam, but the smelter was closed in 1911 and has been dismantled. A smelter for the ores mined by the United States Smelting Refining and Mining Co. was built 1½ miles from Keswick in 1907. After 1919 it was operated for only a short period during 1924, and was dismantled in 1925.

Copper production was at its height in the district from 1898 to 1919, and except for a spurt in production during 1924, the production has declined since 1919. Operations of some of the mines in the West Shasta district ceased when there was still ore in working faces, and known ore reserves. The reason for this was that the smelters along the Sacramento River that treated their ores were closed by smoke-damage litigation, and in most cases, ore reserves were not sufficient to justify the erection of a smelter at a new location.

Zinc was recovered in the West Shasta district only at the Mammoth and the Iron Mountain mines. At the Mammoth mine, small bodies of high-grade zinc ore were mined separately from the copper ore and sent to a zinc refinery for treatment. The zinc in the massive sulfide ore that was mined for copper was not recovered. At Iron Mountain, although all the zinc contained in ores that were direct-smelted was lost, ore from the Mattie and the Richmond Extension ore bodies was treated in a flotation plant that was erected in 1942, and zinc and copper concentrates were recovered.

PRODUCTION

The Shasta copper-zinc district is preeminent in copper production in California, although the production in this district since the end of World War I has been small. The copper from the East and West Shasta districts accounts for 54 percent of the copper produced from California to 1946, and the West Shasta district accounts for the major part. Zinc production has been small, although if zinc had been recovered from all the ore that was direct-smelted for copper, this district would rank among the major zinc-producing districts of the State. The value of the ore as it was mined and treated in the early days of the district gives an erroneous impression of the present or future value of ore of this type in the Shasta district, as neither the zinc nor the sulfur was recovered from most of the ore. Also, no attempt has been made to utilize iron in the residues that result from the removal of base metals and sulfur from the ore. Residues from similar ores in other countries are used as an iron ore.

Table 5 gives the production of the West Shasta district except for the gold and silver that was produced from 1879 to 1897 from the gossan at Iron Mountain; no record of this gold and silver was found by the writers.

The data in table 5 are from many sources, not all of which are in complete agreement, although checking the figures from one source against those from another has shown that there is essential agreement in the figures for most of the mines. Most of the production data were obtained from private reports of R. T. Walker, the United States Smelting Refining and Mining Co., the Coronado Copper & Zinc Co., the Mountain Copper Co., Ltd., and from W. A. Kerr. More detailed information on the source of the data is included in the description of each mine under "Description of deposits."

No production is known from the Spread Eagle, Sugarloaf, Balaklala Angle Station, Great Verde, Crystal, and King Copper prospects.

The exploitation of lode-gold mines and of the limestone resources of the district was stimulated by the operation of the smelters. The need for siliceous flux in the smelters aided lode-gold mining, as quartz veins with a small amount of gold could be mined at that time for the silica content as well as the gold. Quarries were opened in the limestone of the Kennett to provide flux for the smelters. The Holt and Gregg quarry, 4 miles from Kennett on the road to the Golinsky mine, supplied the Kennett and Keswick smelters, and lime was also burned for other commercial uses. The Kennett limestone quarry was located half a mile northwest of the town of Kennett. An analysis of the lime in this quarry showed carbonate of lime 95.2, silica, 4.4, magnesia 0.5 percent, and a trace of carbon (Irelan, 1888, p. 572).

Logan (1947, p. 234) reports:

> Statistics beginning with 1896, credit [Shasta] County with an output of 247,778 barrels of lime and 711,064 tons of limestone. The larger part of both came from the Kennett limestone tributary to Kennett, although quarries in the Hosselkus limestone were operated near Furnaceville and on the west side of Brock Mountain.

TABLE 5.—*Production and grade of ore of the West Shasta copper-zinc district*

[Published with permission of the owners. For source of production data refer to "Description of deposits"]

Mines	Ore		Grade						
	Kind	Production (short tons)	Gold (ounces per ton)	Silver (ounces per ton)	Copper (percent)	Zinc (percent)	Iron (percent)	Sulfur (percent)	Insoluble (percent)
Iron Mountain:									
Richmond ore body (treated by flotation)	Copper-zinc	380,000	0.02	1.00	2.00	3.50	40.5	47.0	6.5
Old Mine ore body	Copper	1,608,000	.04	1.00	7.50	[1] 2.0–5.0(?)			
No. 8 (disseminated ore)	do	820,000	.001	.04	3.50				
Hornet and Richmond-Complex ore bodies	Pyrite for sulfur	3,600,000	.02	.80	.40		43.0	50.0	3.5
Do	Gossan	2,638,000	.073	8.21					
Mammoth	Copper	3,311,145	.038	2.24	3.99	[1] 4.20	34.3	[1] 40.4	13.3
Do	Zinc	84,000	.078	5.79	2.40	21.10			
Balaklala	Copper	1,200,000	.028	1.00	2.80	1.30(?)	31.5	[1] 35.2	23.0
Shasta King	do	[2] 83,889	.034	1.01	2.92	[1 5] 7.61			
Keystone	do	121,802	.06	2.70	6.00	[1] 8.00	30.0	[1] 35.0	15.0
Stowell	do	39.538	.03	1.09	3.05				
Sutro	do	35,307	.08	6.40	7.44				
Early Bird	do	40,116	.032	1.80	4.28	[1 3] 4.72			
Golinsky	do	[4] 6,267	.134	4.61	3.57	[1] 8.90	23.6	[1] 30.7	[6] 27.5

[1] Not recovered.
[2] Assays are for 68,889 tons mined by United States Smelting, Refining, and Mining Co.
[3] Assay from 11 stope samples.
[4] Assays are for 3,078 tons mined by United States Smelting, Refining, and Mining Co.
[5] Assay from 10 stope samples.
[6] SiO_2.

CHARACTER AND DISTRIBUTION

Most of the ore in the West Shasta district is the massive sulfide type. Minor varietal differences occur and some disseminated copper ore has been found, but the relatively small amounts of ore of other types only emphasize the predominance of the massive sulfide ore. The varietal types of ore, found either adjoining the characteristic massive sulfide ore, or as separate bodies, are: banded massive sulfide, ores that contain a high percentage of zinc, chalcopyrite ore that normally contains chalcopyrite and pyrite in equal amounts in a siliceous matrix, and bodies of magnetite. To these might be added some areas of heavily pyritized rock that could, under some conditions, be mined for their sulfur content.

With few exceptions, during mining the distinction between ore and waste in massive sulfide bodies was determined by the copper content. The ore was for the most part treated by direct smelting, and zinc, iron, and sulfur were lost. Massive pyrite that contained only small amounts of copper was considered waste at all the mines except Iron Mountain. Small bodies of high-grade zinc ore were mined separately at the Mammoth mine, and zinc was recovered by flotation from the Richmond Extension ore body at Iron Mountain, but most of the zinc mined in this district was lost, although the content of zinc generally exceeds copper. An exact ratio between copper and zinc in the ores of the district cannot be obtained because of the lack of assays on zinc at many of the mines.

All the reliable data on zinc content of mined ore are listed in table 6.

TABLE 6.—*Production and grade of copper and zinc ore from five mines in West Shasta copper-zinc district*

[Sulfur was recovered only by the Mountain Copper Co., Ltd.]

Mine	Production (short tons)	Grade	
		Copper (percent)	Zinc (percent)
Balaklala	1,200,000	2.8	1.3
Golinsky	3,078	3.57	8.9
Iron Mountain (Richmond ore body)	380,000	2.0	3.5
Keystone	121,802	6.0	8.0
Mammoth	3,395,145	3.95	4.62

The massive sulfide ore is well named; in many ore bodies sulfide minerals, principally pyrite, make up 90 to 95 percent of the mass, and the ore has a brassy metallic appearance. In a few specimens grains of sulfide are too small to be distinguished megascopically; in other specimens clusters of coarse pyrite cubes as much as 5 millimeters across are set in a matrix of fine-grained pyrite that has an average grain size of 0.5 millimeter or less. Differences are so slight that specimens from one mine cannot be distinguished from those from another mine; ores in the same mine vary as much in appearance as ores from different mines in the district.

The massive sulfide ore is generally structureless. Minor color differences due to a greater or smaller content of chalcopyrite or sphalerite are seen locally. Parts of ore bodies that contain more than the normal amount of interstitial gangue minerals form irregular areas that have indefinite boundaries. Some ore tends to be well jointed and breaks into blocks; nearby parts of the same ore body are very strong, and the ore breaks with an almost conchoidal fracture. In large bodies of massive pyrite, such as those at Iron Mountain, the walls and back in open stopes stand well if a screen of ore a few feet thick is left between the stope and the wall rock. Walls of waste rock will slough into the open stope if the ore is mined to the wall.

At a few locations the massive pyrite is veined by either chalcopyrite, sphalerite, quartz, or calcite. The veins are small, rarely more than 1 inch thick, and are not common.

Massive sulfide that contains small quantities of copper and zinc occurs at many of the mines; it is valuable at present only for sulfur, which has been recovered from ore from the Iron Mountain mine. The massive sulfide that contains low-grade copper and zinc occurs as separate ore bodies and also in parts of ore bodies that contain higher grade copper and zinc. Large bodies of massive sulfide in the Iron Mountain, Mammoth, and Keystone mines contain low-grade copper and zinc. In some mines the upper or the lower part of a flat-lying ore body contained too little copper or zinc to mine for these metals and was left in place. In other mines, or at a different locality in the same mine, irregular areas within one ore body may be lower in copper and zinc than other areas. The range in grade is not great, and the appearance and character of the ore is the same. The contact is gradational between massive sulfide that was considered "ore" and massive sulfide that was considered "waste," and in these places the exploitation of a stope depended on the copper and zinc assays. All massive sulfide contains some copper and zinc—the lowest-grade mass, the Hornet ore body at Iron Mountain, contains 0.5 to 1.0 percent of copper. Content of copper and zinc as shown in assays is the only distinguishing feature between high-grade and low-grade copper and zinc ore bodies.

Indistinctly banded to well-banded massive sulfide ore occurs at a few places in some mines. The banding is caused by dark-gray parallel streaks of sphalerite, or more rarely by streaks of chalcopyrite. The streaks range from a few millimeters to 1 centimeter in width and have hazy, indefinite boundaries; they generally parallel the foliation in the wall rocks, according to observers who have seen this type of ore in underground exposures in the district (Walker, R. T., Bagley, T. P., and Hunt, R. N., oral communication; Hershey, O. H., private report for the Mountain Copper Co., Ltd.). Examples of banding due to alternate layers of coarse and fine pyrite, similar to that described by Brownell and Kinkel (1935, p. 277), were seen at Iron Mountain, but they are not common. Ptygmatic folding in banded ore, such as that described by Gavelin in the ores of the Malänas district (Gavelin, 1939, p. 141) has not been observed here.

Small areas of high-grade zinc ore occur at the Iron Mountain mine; the Richmond Extension ore body contained sufficient zinc in veinlets and in grains interstitial to pyrite to justify making a zinc concentrate by flotation. Massive sulfide that contains a high percentage of zinc occurs mainly in the Mammoth mine; it is coarser grained than most massive sulfide, and is found as small isolated ore bodies for the most part, rather than as zinc-rich parts of the main ore bodies as at the Iron Mountain mine. High-grade zinc ore bodies were mined at the Mammoth mine. Although they have the megascopic appearance of a massive sulfide ore, the percent of insoluble material was 35.5 before and 16.8 after sorting.[4]

Ore that consists of chalcopyrite and pyrite as veinlets and disseminations in schistose rock occurs in quantity at the Iron Mountain mine, and is described in detail in the description of that mine. The ore consists of chalcopyrite stringers and veinlets, largely parallel to the foliation, and of disseminated grains of chalcopyrite in siliceous foliated rhyolite. The pyrite content generally equals chalcopyrite and occurs as discrete grains. The body of disseminated ore is separated from the nearby massive sulfide ore bodies by barren rock; there is no gradation between the two types at the Iron Mountain mine.

The only other known occurrence of disseminated copper ore in the district is at the Balaklala mine. At this mine, the disseminated copper ore lies below the massive sulfide body and is in contact with it. Veins and stringers of chalcopyrite in siliceous rhyolite parallel the contact of the massive sulfide body. The tonnage of disseminated or vein chalcopyrite ore at the Balaklala mine is small.

Deposits of disseminated pyrite, although not now commercial, under some conditions could be a source of sulfur. These deposits differ from the massive sulfide ore in that they contain much gangue and unreplaced rock, do not have sharp boundaries, and the pyrite is coarser grained. The pyrite thus far exposed in the Sugarloaf prospect east of Iron Mountain is of this type. Similar deposits occur in the district; some are separate bodies, but others form bands adjacent to or near bodies of massive sulfide. Deposits of disseminated pyrite contain 50 to 70 percent of pyrite, commonly as separate euhedral crystals (cubes) in a matrix that is composed almost entirely of secondary quartz. Such bodies rarely have sharp walls, and they fade out into weakly pyritized rocks and may contain much unreplaced rock material. Little copper and no zinc accompanies this type of ore, and the few assays that are available indicate that they contain no gold or silver. No gradation between the normal massive sulfide ore and the pyrite-quartz rock has been seen in any of the mines, except at the ends of some ore bodies.

STRUCTURAL FEATURES

FORM OF THE ORE BODIES

In the major mines in the district the ore zone dips gently or is flat lying; within the zone, most of the individual ore bodies are flat lying. The length and width are commonly 2–10 or more times greater than the thickness. Most of the ore bodies occur as pods, lenses, pancake-, kidney- or cigar-shaped forms, although a few small ones have irregular shapes. Many are bent or warped. The Iron Mountain ore body, which has a lenticular vertical cross section in the central part, is also synclinal; the Balaklala and Shasta King ore bodies are lenticular in vertical cross section but are saucer- or basin-shaped forms. The ore zone at the Mammoth mine occurs along the crest of a broad elongate arch, and is thickest along the crest. The small Mattie ore body at Iron Mountain has a flat-lying cigar-shaped form. The Early Bird ore body also is horizontal, but has a lenticular and in part synclinal vertical cross section. The only lenticular ore body that has a steep plunge is in the Golinsky mine.

Where primary rock structures can be determined, the flat-lying or gently dipping ore zones generally follow primary structures in the enclosing rocks, such as contacts between flows, but the location of some of the smaller, steeply dipping ore bodies is probably controlled by a steep foliation.

Metallization, which consists chiefly of pyritization, is widespread in the district. Bodies of partly pyritized rock cover an area of several square miles; most of the rock contains some pyrite. Massive sulfide ore bodies occur as sharply bounded discrete bodies in the broad zone of pyritization or ore zone. There is no correlation in many instances between the amount of scattered pyrite in the broad zone of pyritization and the location

[4] Data furnished by the United States Smelting Refining and Mining Co.

of bodies of massive sulfide ore, and there is normally no gradation between pyritized rock and massive sulfide.

The distribution of the ore bodies, pyritized rock and hydrothermal alteration is shown on plate 4. No pattern in the lateral distribution of mines is shown on the map, but this lack of pattern is somewhat deceptive, as described under "Summary of base-metal ore controls." The vertical location of the ore zone is controlled by the stratigraphy in the Balaklala rhyolite. Ore bodies have been discovered, with a few exceptions, only where the favorable zone has been exposed along the walls of canyons and has not been removed by erosion.

TYPES OF CONTACTS BETWEEN ORE AND WALL ROCKS

The contact is sharp at most places between massive sulfide bodies that normally contain less than 10 percent of gangue minerals and wall rock that normally contains less than 10 percent of pyrite, or no pyrite at all. There are no frozen contacts between massive sulfide and wall rock. The contact between massive sulfide and wall rock is generally marked by a gouge of white clay that ranges from a thin film to a zone several feet in thickness. The rock near most contacts is hydrothermally altered and foliated; there is normally a gradation between gouge and foliated wall rock, and at some places the foliated wall rock grades within a few feet to nonfoliated wall rock. In this type, the massive sulfide ends abruptly against gouge; the sulfide has a smooth wall, but shows slickensides or other evidence of movement at only a few places. No change in the appearance of the massive sulfide near contacts is apparent. Relict quartz phenocrysts from the rhyolite can be found in the foliated rock and at a few places they have been found in the white clay gouge.

Although many ore bodies have blunt, rounded ends and a sharp contact with barren wall rock a few have gradational contacts between the massive sulfide and wall rock. At the Shasta King mine, for example, massive sulfide grades to pyritized rock along the ends of the ore body. The sulfides at these places replaced foliated rocks preferentially along the foliation. At Iron Mountain a few contacts were seen in which the foliated wall rock contained streaks of disseminated pyrite cubes parallel to the foliation and to the massive sulfide. In this example, and as mentioned above, there is evidence that the foliation was presulfide in age.

At a few localities foliated rock adjoining the massive sulfide may contain irregular bodies of pyrite a few inches to a few feet long, which have hazy, gradational boundaries. A main gouge is always present at the outer edge of this pyritized zone; such contacts indicate incomplete replacement along a gouge.

RELATIONSHIP OF ORE BODIES TO STRUCTURES IN THE HOST ROCK

RELATIONSHIP TO STRATIGRAPHY

The minable ore bodies that have thus far been found are at the same general stratigraphic zone in the Balaklala rhyolite throughout the district, and even where no ore bodies occur, the rocks at this zone are more or less pyritized. The favorable stratigraphic zone is the group of flows and pyroclastic rocks that underlie the base of the coarse-phenocryst rhyolite in the central part of the district, or below the tuff beds that are the stratigraphic equivalent of the coarse-phenocryst rhyolite in the outer parts of the district. Ore occurs through a stratigraphic thickness of 600 feet at Iron Mountain, and the ore zone may locally have a greater thickness. In terms of the three units of the Balaklala rhyolite, as described in detail under "Balaklala rhyolite," the ore zone is along the upper part of the middle unit. Details of the types of rock associated with the ore bodies are included in the descriptions of the individual mines.

The upper part of the middle unit of the Balaklala rhyolite is called the favorable zone, or the ore zone, as it contains all the known ore bodies in the mineral belt. It is composed of a group of discontinuous flows, and of lenticular beds of coarse and fine pyroclastic material; more of the bedded, water-deposited tuff and volcanic conglomerate is present in this zone than in any other zone in the Balaklala. The heterogeneous nature of this material makes the detailed stratigraphy at each mine unique, yet it is possible to locate the ore zone with a fair degree of certainty because this heterogeneous group is capped by a recognizable unit throughout the district. Thus at the Iron Mountain mine, although the coarse-phenocryst rhyolite flow was either eroded or did not extend as far south as Iron Mountain, the lithology of the tuff beds above the ore zone is characteristic of the tuff beds that are the stratigraphic equivalent of the transition zone at the base of the upper unit of the Balaklala, and the rocks in which the ore occurs are lithologically characteristic of the middle unit of the Balaklala. Similarly, the tuff beds that overlie the Stowell and Keystone mines and the Spread Eagle prospect are characteristic of the material that occurs either as the stratigraphic extension of the coarse-phenocryst rhyolite or of the material that forms a transition zone immediately below this rhyolite of the upper unit.

In the Balaklala, Mammoth, and Sutro mines, the ore bodies lie in the middle unit of the Balaklala rhyolite immediately below the coarse-phenocryst rhyolite, but the Shasta King ore body is lower in the middle unit, probably several hundred feet below the base of the upper unit, which is eroded from above the Shasta King. The ore body at the Golinsky mine is also probably several hundred feet below the base of the upper unit, although the relationship at this mine is not as clear as at the other mines.

The thickness and types of flows, and the amount and type of pyroclastic material varies widely in the middle unit, and this unit cannot be used alone as a horizon marker. It is necessary to use a combination of the lithologic types characteristic of the upper unit, and the widespread nonporphyritic flows of the lower unit, as well as medium-phenocryst flows and tuff beds of the middle unit, to determine the stratigraphic position of the ore zone at any one locality.

Pyritization in the ore zone is more widespread than the position and the extent of the known ore bodies on plate 4 would indicate. Many drill holes in the central part of the district have cut through the coarse-phenocryst rhyolite and into the upper part of the middle unit. Most of the holes drilled between known ore bodies encountered pyritized rock at the stratigraphic position of the ore zone; they show that the pyritized zone is continuous although massive sulfide ore is in discrete bodies.

At some places the character of the rock that was replaced by ore can be determined; the favored host rock for ore bodies appears to be a porphyritic rhyolitic flow that contains 2- to 3-millimeter quartz phenocrysts. In most ore bodies this rock was unsheared at the time of replacement, and commonly the ore-bearing flow is overlain by thinly bedded tuff or fine pyroclastic material. This is particularly well shown at the Shasta King mine, in the glory hole at the Balaklala mine, and at the Mammoth mine.

RELATIONSHIP TO FOLDS AND FOLIATION

Individual ore bodies, as well as broad pyritized zones that contain several ore bodies, tend to be concentrated on or near the axes of folds. Although synclinal and basin-shaped ore bodies predominate, ore bodies were formed both in anticlines and in synclines, but some occur on the flanks of folds. At a few mines the primary layering of the host rock is not sufficiently apparent to allow the mapping of detailed structures, and the relationship between primary structures and the location of ore is not known.

Bedding is well developed in some types of volcanic material in the West Shasta district, but is lacking in others. Where bedded pyroclastic rocks are present, the large and small folds and warps in the rocks can be recognized, but in parts of the area only the broad structures such as can be interpreted on the basis of large units are decipherable. At the mines where bedded material is present or flow contacts can be mapped, it was possible to relate the location of ore to the location of folds.

Examples of ore bodies in synclines or basin-shaped warps are the Richmond ore body of the Iron Mountain mine and the Balaklala, Shasta King, and probably the Early Bird mines; in these mines the plunge of the folds is gentle. The general ore zone at the Mammoth mine is on a broad arch or dome-shaped structure, and some individual ore bodies within this zone are on minor arch structures. At many places in the district pyritization favors the crests of small folds, or is concentrated under flat rolls in the dip. The Sutro, Keystone, and Stowell mines and the Old Mine ore body of the Iron Mountain mine are probably on the flanks of broad folds, although the folds at these mines could not be mapped in detail except at Iron Mountain.

The rhyolitic rocks in the mineral belt are only locally strongly foliated probably because the folds are broad, but the two types of foliation that develop in flexural-slip folding, both bedding-plane foliation and fracture cleavage, have had a considerable effect on ground preparation before ore deposition.

Flexural-slip folding, in which movement along bedding planes is concentrated in the incompetent layers, is found locally in the West Shasta district. The platy minerals that are formed in these layers are oriented parallel to bedding (Knopf and Ingerson, 1938, p. 159–161; Swanson, 1941, p. 1256). This bedding-plane foliation is concentrated along layers of bedded pyroclastic material, and appears to be as strongly developed on the axes as on the flanks of folds, suggesting some slippage parallel to fold axes. This type of foliation is most common where bedded material between flows is a few inches to a few feet thick; where the bedded material is thicker, foliation is commonly limited to a zone at the top or bottom of the bed.

Poorly developed to well-developed fracture cleavage in competent layers occurs in conjunction with bedding-plane foliation. Fracture cleavage as here used follows the usage of Leith (1923, p. 148–150) and Balk (1936, p. 706); the cleavage ranges from slightly more than subparallel jointing to foliated rocks with reorientation of minerals along cleavage planes. The common type of fracture cleavage in the district has the appearance of closely spaced sheeting in which the rock is divided into lenticles of unaltered rock separated by seams of oriented sericite. The cleavage planes are spaced a few millimeters to 4 centimeters apart. In a few areas the planes of sheeting on which sericite has formed are so closely spaced that the rock has the appearance of a sericite schist, although in thin section the distinction between sericite seams and rock that has not been recrystallized is sharp.

Fracture cleavage is found at many places in the district, but is not widespread in the sense that it is continuous. The characteristic feature of this type of foliation is the range in intensity of sheeting in short distances and the lack of continuity either along or across the strike. The lack of continuity may be more apparent than real, as the bodies of sheeted rock have a plunging linear element. Sheeting may occur in one flow along a fold axis but not in the overlying or underlying flows because of a difference in competence. In plunging folds the body of sheeted rock would appear as an irregular area in a horizontal plane. The fracture cleavage is steep, and although it tends to have a regional orientation and is related to folding, it is not axial plane cleavage.

During the folding steep fracture cleavage formed in parts of the lower and middle units of the Balaklala rhyolite, particularly along axes of folds, but cleavage is rare in the upper unit except in the thin, outer edges of the coarse-phenocryst rhyolite dome. Locally it is well developed in the flows underlying the upper unit, particularly in flows capped by fairly continuous beds of pyroclastic material. The coarse-phenocryst rhyolite is the most competent rock in the Balaklala, but the nonporphyritic rhyolite of the lower unit and the medium-phenocryst rhyolite of the middle unit are also competent except where they contain layers of bedded pyroclastic material. Bedding-plane movement accompanied by foliation was concentrated in the layers of bedded pyroclastic material in the upper part of the middle unit where these beds are most common (fig. 38).

The intersection of steep fracture cleavage and flat foliation controlled by bedding is one of the principal factors that localized the ore on folds, because a zone of rock with steep fractures was formed along the crest of a fold or dome under a relatively impervious cover formed by beds with flat foliation.

RELATIONSHIP TO FAULTS

Premineral and postmineral faults are present in the district. Most of the ore bodies are localized near major faults or shear zones that trend about N. 70° E., but are not necessarily adjacent to them. In a few places it is evident that these premineral faults were feeder channels for mineral-bearing solutions.

At Iron Mountain, the Camden-Sugarloaf fault is premineral, and may have been a channel for the mineral-bearing solutions that formed the main ore bodies as well as the areas of pyritized and altered rock that are present along the fault southwest and northeast of the main ore bodies. Similarly, the ore bodies at the Balaklala mine are localized near the Balaklala fault, and at the Mammoth mine they are localized near the California fault.

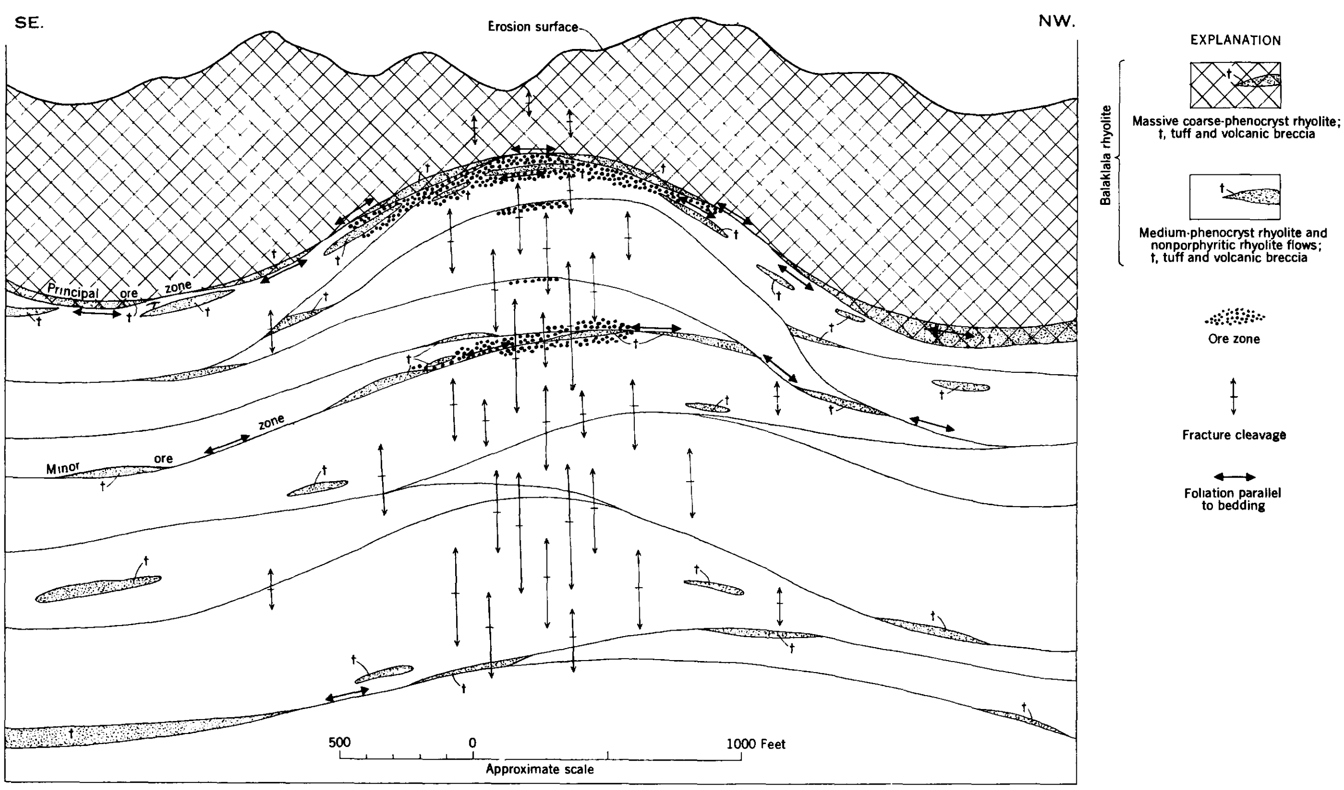

FIGURE 38.—Diagrammatic drawing showing the relationship between bedding-plane foliation, fracture cleavage, and pyritized zones at the Mammoth mine.

Premineral and postmineral faults can be distinguished, even though in most examples postmineral movement has occurred along a fault that was originally premineral in age. Evidence that a fault is premineral in age is shown by pyritized and hydrothermally altered rock along the walls, or centers of metallization at several points along the fault. Ore bodies that form along a fault and terminate abruptly against it with no evidence of being cut off, also indicate that the fault guided solution travel. Contacts of ore bodies that show banding parallel to shear zones, consisting of alternating bands of pyritized and barren rock, are interpreted as incomplete replacements of sheared rock along the fault.

Postmineral faults, many of which contain several feet of gouge, offset ore bodies from a few inches to as much as 300 feet. Few ore bodies in the district still retain their original form; most are divided by postmineral faults into separate blocks of ore that were mined individually. Thus at Iron Mountain, the Old Mine, Brick Flat, Richmond, and the Hornet ore bodies, each of which is offset several hundred feet from the other by faults, were originally parts of one continuous body of ore. The Balaklala, Shasta King, and Early Bird ore bodies are cut by faults that offset the original ore bodies into separate blocks. Many postmineral faults in massive sulfide, with a displacement from a few inches to a few tens of feet, are seen in underground workings. Along these faults the face of the sulfide body has a striated brassy mirrorlike slickensided surface; on other postmineral faults a gouge is present. This gouge consists of crushed sulfide minerals and small rounded slickensided fragments of ore in a claylike matrix that is colored dark gray by the crushed sulfides.

MINERALOGIC DESCRIPTION

GENERAL FEATURES

The mineralogy of the hypogene sulfide ore bodies in the West Shasta district is not complex. The ore bodies are massive sulfide that contains mainly pyrite, chalcopyrite, and sphalerite, and minor amounts of magnetite, galena, tetrahedrite, and pyrrhotite. Gangue minerals are quartz, sericite, chlorite, calcite, barite, and zoisite. The sulfide ores contain small amounts of gold and silver, but no silver minerals or gold were recognized. Sulfide minerals constitute from a minimum of about 65 to a maximum of about 98 percent of the massive sulfide ore. Assays of samples from 1,400 feet of drill core of massive sulfide ore from the Hornet ore body at the Iron Mountain mine averaged 2.68 percent silica, and the average of 3,600,000 tons mined from this ore body was only 3.5 percent insoluble material. The zinc ore mined at the Mammoth mine averaged 35.5 percent insoluble material, the highest of any of the massive sulfide ores, but this included barren rock as the ore after sorting contained only 16.8 percent insoluble.

Two small bodies of iron oxides that contain magnetite, hematite, and limonite are in the Iron Mountain area. These are replacement bodies of iron oxides that are separated from the massive pyritic ore by porphyritic rhyolite, and their genetic relationship to the massive sulfide deposits is not known. Specimens of magnetite found on the dump at the Spread Eagle mine, indicate that bodies of magnetite are present in the mine. Magnetite also occurs as tiny grains in a few specimens of massive sulfide ore, but it is not common.

Scheelite is not known in the massive sulfide deposits. However some of the biotite-quartz diorite in Clear Creek in the Igo quadrangle has thin coatings of scheelite along fractures.

The mineralogic study of the West Shasta copper-zinc district is incomplete. The Iron Mountain mine was the only operating mine in the district during the period of investigation from November 1945 to June 1952, although many of the workings in ore at the Shasta King and Early Bird mines were accessible. Specimens of sphalerite from the Yolo zinc ore body of the Mammoth mine were studied from the Lindgren collection, which was kindly loaned to the writers by Prof. Robert R. Shrock of Massachusetts Institute of Technology. Other specimens of zinc-rich ore from the Mammoth mine were studied from the Grotefend collection, which is on display at the California Division of Mines office in Redding, Calif. Some specimens of zinc-rich ore collected along the tramline near the Stowell mine were studied, together with specimens of pyritic ore from mine dumps. The supergene-enriched ore was not observed underground, and only a few specimens were available for examination from the Old Mine ore body at the Iron Mountain mine.

The list of minerals in the West Shasta copper-zinc district is tabulated below.

Minerals in West Shasta copper-zinc district

Hypogene minerals:

Ore minerals:

Chalcopyrite	$CuFeS_2$
Galena	PbS
Gold	Au
Greenockite (?)	CdS
Hematite	Fe_2O_3
Ilmenite	$FeTiO_3$
Magnetite	Fe_3O_4
Pyrite	FeS_2
Pyrrhotite	$Fe_{1-x}S$
Scheelite	$CaWO_4$
Silver	Ag
Sphalerite	ZnS

Minerals in West Shasta copper-zinc district—Con.

Hypogene minerals—Continued	
Ore minerals—Continued	
Talc	$H_2Mg_4(SiO_3)_5 \cdot H_2O$ (?)
Tennantite	$5CU_2S \cdot 2(CuFe)S \cdot 2As_2S_3$
Tetrahedrite	$5Cu_2S \cdot (CuFe)S \cdot 2Sb_2S_3$
Gangue minerals:	
Barite	$BaSO_4$
Calcite	$CaCO_3$
Chlorite	$(Mg,Fe)_5(Al,Fe^{+++})_2Si_3O_{10}(OH)_8$
Quartz	SiO_2
Sericite and hydromica	$KAl_3Si_3O_{10}(OH)_2$
Zoisite	$Ca_2Al_3(SiO_4)_3(OH)$
Supergene minerals:	
Oxide zone:	
Antlerite	$Cu_3(OH)_4SO_4$
Azurite	$Cu_3(CO_3)_2(OH)_2$
Chalcanthite	$CuSo_4 \cdot 5H_2O$
Copper	Cu
Cuprite	CuO
Copiapite (?)	Basic ferric sulfate.
Goslarite	$ZnSO_4 \cdot 7H_2O$
Limonite	Hydrous iron oxides.
Maghemite	Fe_2O_3
Malachite	$Cu(OH)_2 \cdot CuCO_3$
Melanterite	$FeSo_4 \cdot 7H_2O$
Silver	Ag
Smithsonite (?)	$ZnCO_3$
Sulfur	S
Wad	Hydrous manganese oxide.
Sulfide zone:	
Chalcocite	Cu_2S
Covellite	CuS

HYPOGENE MINERALS

ORE MINERALS

Chalcopyrite, $CuFeS_2$.—Chalcopyrite is present in all the massive sulfide deposits in the district, in the disseminated copper deposits of the No. 8 mine and the adjoining Confidence-Complex ore bodies at Iron Mountain, and in the disseminated ore at the Balaklala mine. It also occurs in minor quantities in the gold-quartz veins in the district. In many of the low-grade pyritic deposits, such as the Hornet ore body which averaged only 0.40 percent copper, chalcopyrite cannot be detected except under a microscope. It commonly occurs, along with a little sphalerite and gangue minerals, as a thin network 0.2 to 0.3 millimeter thick surrounding grains of pyrite and through fractures in pyrite. In massive sulfide ore, the chalcopyrite has little corrosive effect upon pyrite, and mainly replaced a thin selvage of gangue minerals between pyrite grains, or forms poorly defined streaks, irregular patches, or veinlets a few inches thick in massive pyrite.

All the zinc ore contains some chalcopyrite, either as minute globules of chalcopyrite, which are unevenly distributed through the sphalerite and which may have formed by unmixing during cooling (Buerger, 1934, p. 525–530), or as streaks or irregular masses in massive sphalerite. In places the chalcopyrite occurs as irregular, worm-shaped inclusions that form a pseudoeutectic texture.

Some chalcopyrite was observed in the gold-quartz veins in the Igo area, in the Igo quadrangle; in the Old Diggings district, in the Shasta Dam quadrangle; and at the Uncle Sam mine, in the central part of the district. Quartz-chalcopyrite veins occur in the No. 8 mine and the Confidence-Complex workings at the Iron Mountain mine.

Galena, PbS.—Galena was observed only in specimens from the zinc-rich Yolo ore body at the Mammoth mine, and in the zinc-rich parts of pyritic ore bodies, at the Iron Mountain, Stowell, Sutro, Golinsky, and Shasta King mines; it is always associated with sphalerite. Galena occurs as irregular grains 0.1 to 0.2 millimeter in diameter in gangue minerals in sphalerite, along contacts between gangue and sphalerite, and as inclusions in sphalerite. Figure 39 is a polished section of zinc-rich ore from the Mammoth mine, which contains galena (*g*) that appears to be corroding sphalerite (*s*).

Gold, Au.—Gold occurs in small amounts in massive sulfide deposits in the West Shasta copper-zinc district; it also occurs in gold-quartz veins in the Uncle Sam and Clipper mines in the central part of the base-metal district and in many quartz veins of the district, although

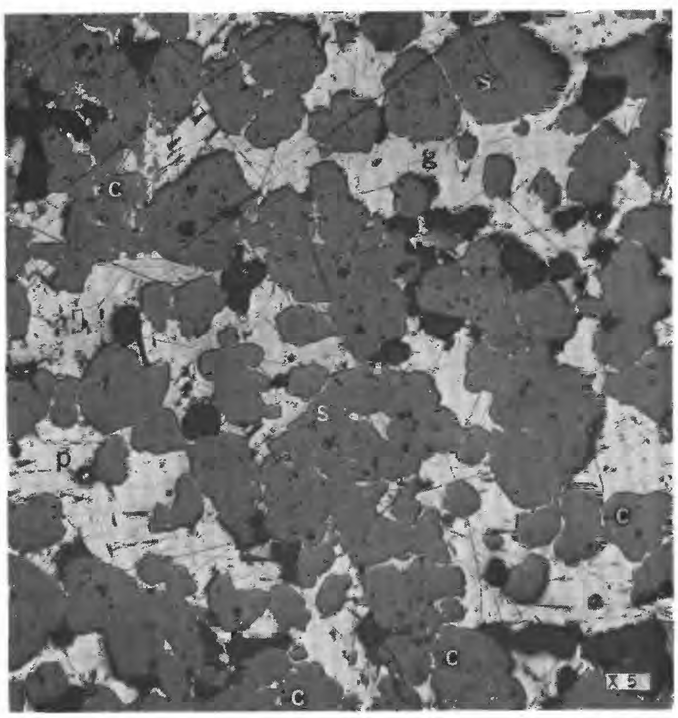

FIGURE 39.—Photomicrograph of ore from the Mammoth mine. Galena (*g*) appears to corrode sphalerite (*s*). A few grains of pyrite (*p*), quartz (dark gray), and chalcopyrite (*c*). × 50.

it is rare in veins in the biotite-quartz diorite. Placer gold has been mined from most of the stream gravels and from a few areas in the gravel of the Red Bluff formation.

Gold in recoverable amounts occurs in most of the massive sulfide deposits and in gossan derived from these deposits, but no gold was found in the polished sections studied.

Greenockite(?), *CdS*.—No cadmium minerals were definitely recognized in the massive sulfide ore, although in one specimen of zinc-rich ore from the Mammoth mine, thin films of a yellow, transparent mineral between a limonite coating and massive sulfide may be greenockite. Cadmium was recovered during 1917–18 from the ores of the Mammoth mine. Spectrographic analyses of massive sulfide ore from the Mammoth, Early Bird, Balaklala, and Iron Mountain mines were made by the Geological Survey. A specimen of zinc-rich ore from the Mammoth mine contained 0.1 to 1 percent cadmium, and a specimen of massive pyritic ore from the Balaklala mine contained 0.01 to 0.1 percent cadmium. The other samples all had less than the threshold value of 0.01 percent of cadmium.

Hamilton (1922, p. 241), reporting on the Mammoth mine, says:

> Cadmium there occurs [at the Mammoth mine] associated with zinc sulfide, sphalerite, probably as the sulfide greenockite.

Eakle (1923, p. 47) states:

> Cadmium as greenockite occurs in the copper-zinc ores of [Shasta] County and the Mammoth Copper Company recovers it in their electrolytic zinc plant.

Murdock and Webb (1948, p. 164) report:

> Several thousand pounds of cadmium were produced [1917–18] at the Mammoth Copper Company plant, presumably from cadmium in the sphalerite. This may in part be as greenockite associated with the sphalerite * * *

Hematite, Fe_2O_3.—Specular hematite, which is strongly anisotropic from colorless to brownish-purple, makes up about 10 percent of the two small magnetite bodies at Iron Mountain. This mineral has a radial, platy structure, the plates of which are 0.5 to 1 millimeter long. No hematite was observed in primary massive sulfide ore.

Ilmenite, $FeTiO_3$.—Ilmenite was not observed in any of the massive sulfide deposits, but it occurs in small quantities in the two magnetite bodies at Iron Mountain as tiny irregular plates intergrown with magnetite. A sample of the magnetite contained 0.25 percent titanium oxide.

Magnetite, Fe_3O_4.—Magnetite occurs at Iron Mountain as two lenses several hundred feet long that lie along the side of, but not in contact with, the massive sulfide ore. Other occurrences are at the Spread Eagle mine, where magnetite was found on a dump, and as minute black grains, that megascopically, are visible disseminated through some of the high-grade chalcopyrite-pyrite ore from the Clark ore body of the Mammoth mine. Under the microscope magnetite is seen to be disseminated through gangue that is interstitial to pyrite.

Pyrite, FeS_2.—Pyrite is the predominant mineral in the massive sulfide deposits. The pyrite ranges in content from about 96 percent in the Hornet ore body of the Iron Mountain mine to less than 10 percent in the zinc ore bodies of the Mammoth mine. It is disseminated through much of the Balaklala rhyolite in amounts about as much as 15 percent by weight, particularly where the rhyolite is hydrothermally altered. Pyrite is also widely disseminated in all the pre-Cretaceous rocks, although it is rarely present in the Shasta Bally batholith.

Megascopically, massive pyritic ore looks very fine grained, but it contains a few disseminated 1- to 2-millimeter euhedral pyrite cubes or, less commonly, pyritohedrons, and some irregular clumps of more coarsely crystalline pyrite averaging as much as 2 millimeters in grain size.

Under the microscope the pyrite is seen to be in euhedral, subhedral, and fractured grains about 0.1 to 0.4 millimeter in diameter separated by a thin network, from 0.05 to 0.5 millimeter thick, of chalcopyrite, some sphalerite, and quartz and rock gangue (figs. 40

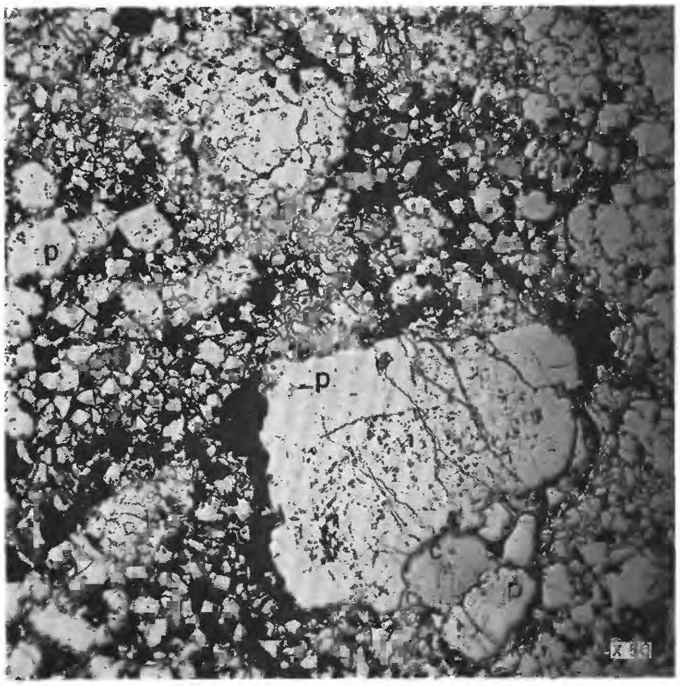

FIGURE 40.—Photomicrograph of ore from the Golinsky mine. Pyrite (*p*) with minor chalcopyrite (*c*) and quartz and rock gangue (dark gray), × 50.

and 41). In places pyrite forms a granular aggregate without the thin network of gangue and chalcopyrite.

The pyrite that is disseminated throughout much of the Balaklala rhyolite is much coarser grained than that in the massive sulfide ore. Individual grains are generally 1 to 2 millimeters in diameter but a few cubes are as much as 1 centimeter. This more coarsely crystalline pyrite commonly occurs as euhedral cubes or less commonly as pyritohedrons.

Pyrrhotite, $Fe_{1-x}S$.—Pyrrhotite was observed in only one polished section of a high-grade chalcopyrite ore from the Iron Mountain mine. This polished section contains massive pyrrhotite surrounding pyrite.

Scheelite, $CaWO_4$.—Thin films of scheelite are present in biotite-quartz diorite along Clear Creek in the Igo quadrangle. No scheelite was observed in any of the massive sulfide deposits.

Silver, *Ag*.—No silver minerals have been seen in any massive sulfide deposits, although silver has been recovered from the ore. The Old Mine ore body at Iron Mountain averaged 1.0 ounce of silver per ton. Copper ore produced from the Mammoth mine averaged 2.24 ounces of silver, and the zinc ore averaged 5.79 ounces of silver per ton. Tetrahedrite is present in the zinc-rich ore and may account for a higher silver content.

High-grade silver veins were mined during the late 19th century in the South Fork mining district 2½ miles northwest of Igo in the Igo quadrangle (Tucker, 1926, p. 201–210). Tucker reports that these veins contained native silver, tetrahedrite, argentiferous galena, sphalerite, pyrite, and small amounts of gold in quartz veins in granodiorite. The principal mine was the Silver Falls mine.

Sphalerite, *ZnS*.—Most of the sphalerite in the massive sulfide deposits is a fine-grained, reddish-black variety that probably contains considerable iron. Few cleavage faces as much as 1 millimeter in diameter are visible. Sphalerite occurs with chalcopyrite in lenses, veinlets, and irregular gray masses in pyritic ore bodies, in minor quantities with chalcopyrite and gangue in the network around grains of pyrite, and in isolated zinc-rich ore bodies as in the Yolo and 313 ore bodies at the Mammoth mine. Sphalerite in the Yolo zinc ore body is lighter colored and much coarser grained than that found in other zinc ore bodies in the district. The sphalerite in this ore body has a resinous color in specimens from the Lindgren collection and has cleavage faces that range from 2 to 10 millimeters in diameter.

Sphalerite actively corroded pyrite when it was introduced; and where much sphalerite is present, the pyrite content is very low. In high-grade sphalerite ore chalcopyrite always occurs as streaks, lenses, or irregular areas, as in figure 42, and as minute blebs or tiny, irregular-shaped inclusions erratically distributed through sphalerite.

The zinc-rich ore contains more insoluble material than the massive pyrite. Polished sections of zinc-rich ore contain 25 to 35 percent corroded relicts of quartz and rock gangue. These relicts are 0.5 to 1 milli-

FIGURE 41.—Photomicrograph of ore from the Early Bird mine. Pyrite (*p*), quartz and rock gangue (dark gray), and minor chalcopyrite (*c*), × 50.

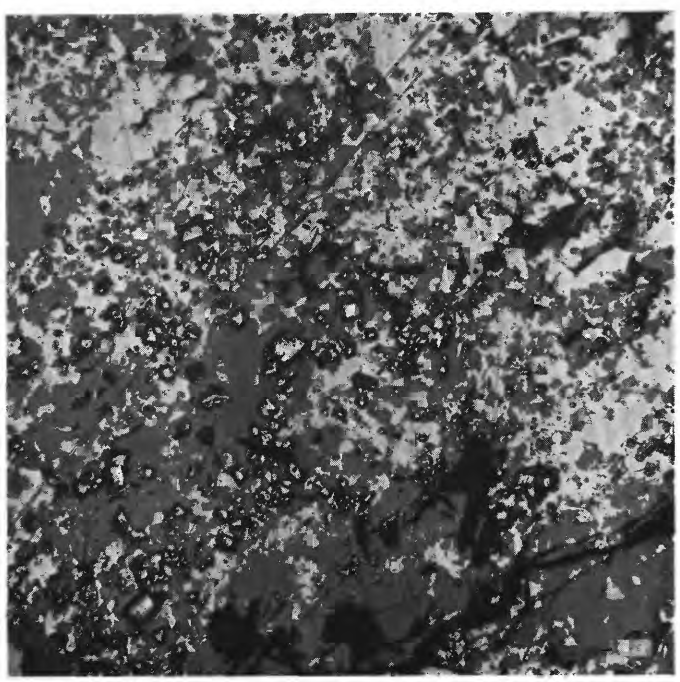

FIGURE 42.—High-grade zinc-copper ore from the Shasta King mine. White, chalcopyrite; gray, sphalerite, × 50.

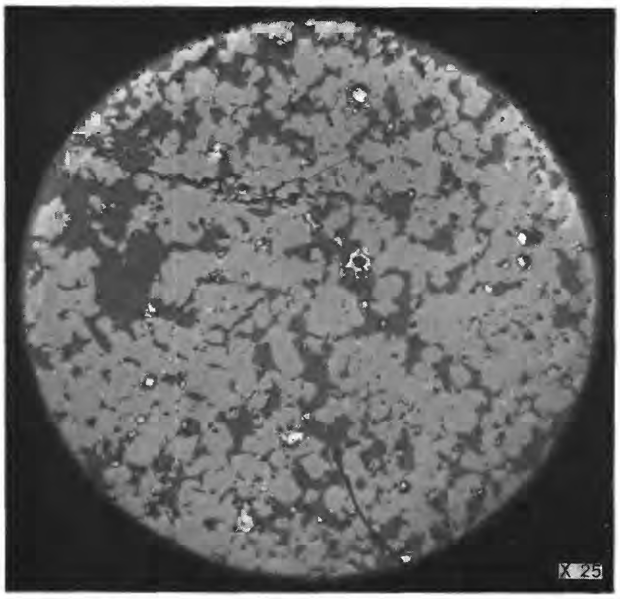

FIGURE 43.—Photomicrograph of polished section of zinc ore from Mammoth mine. Corroded quartz gangue (dark) in sphalerite (light) forming an island-and-sea texture. Ordinary light, × 25.

meter in diameter and are distributed through the sphalerite forming an island-and-sea texture (fig. 43).

Talc, $H_2Mg_4(SiO_3)_5 \cdot H_2O$ (?).—Talc is not associated with base-metal ore, but it occurs in a pyritized shear zone in Copley greenstone at the Ganim mine, which is situated 2½ miles northwest of Whiskytown in the Whiskytown quadrangle. It also occurs in hornblende bodies near the margin of the biotite-quartz diorite in the Igo quadrangle, and in the plug of coarse-phenocryst rhyolite near the Uncle Sam mine.

Two grades of talc are found in the Ganim mine; one is nearly pure talc or steatite, whereas the other is a mixture of talc and carbonate, not usable as steatite (Page and Wright).[5] The talc occurs in a zone of intensely sheared pyritized and altered greenstone 100 feet wide and 1,200 feet long.

The talc near the Uncle Sam mine is in intensely sheared and silicified rhyolite. It occurs as small pockets and lenses less than 6 inches long in quartz-talc schist in the margin of the plug of coarse-phenocryst rhyolite, and in nonporphyritic rhyolite near this plug.

Tennantite, $5Cu_2S \cdot 2(CuFe)S \cdot 2As_2S_3$ *and Tetrahedrite*, $5Cu_2S \cdot 2(CuFe)S \cdot 2Sb_2S_3$.—Tetrahedrite or tennantite (or both) is present in small quantities in the zinc-rich ore in the district. They are in anhedral grains up to 0.7 millimeter in diameter in sphalerite, chalcopyrite, and galena and have mutual borders with sphalerite and chalcopyrite.

[5] Page, B. M., and Wright, L. A., 1943, Unpublished geologic maps and sections of the Ganim talc mine, Shasta County, Calif.: In open file of U. S. Geol. Survey.

The mineral was determined by etch tests as being in the isomorphous series tetrahedrite-tennantite. Seager in an unpublished report, 1934, identified tetrahedrite from the Mammoth mine and tennantite from the Golinsky mine.

GANGUE MINERALS

Barite, $BaSO_4$.—Barite is rarely present in the West Shasta district (Graton, 1910, p. 102), although it is one of the common gangue minerals in the East Shasta copper-zinc district. Tiny plates of barite were recognized in a thin section and a polished section of zinc ore from the Mammoth mine. Spectrographic analyses of ore from the Mammoth, Early Bird, and Balaklala mines showed 0.01 percent barium.

Seager noted small amounts of barite in the sulfide ore from the Golinsky mine and in the gossan from the Mammoth mine. Graton (1910, p. 102) reports a small amount of barite in the Mammoth sulfide ore, the Mammoth gossan, and in the sulfide ore from the Golinsky mine.

Calcite, $CaCO_3$.—Calcite is a minor gangue mineral in the massive sulfide ore. It is present in veinlets, irregular patches, and in vugs in the massive sulfide.

Chlorite $(Mg,Fe)_5(Al,Fe)_2Si_3O_{10}(OH)_8$.—Chlorite is present at some places in a thin network between sulfide grains. Flakes of this mineral are interlayered with sericite and are commonly oriented perpendicular to the grain boundaries of the sulfide minerals. Chlorite is a much less characteristic gangue mineral than sericite.

Quartz, SiO_2.—Quartz is the predominant gangue mineral in the massive sulfide ore bodies of the West Shasta district. It occurs as a network surrounding grains of sulfides; in thin veinlets, which may contain small amounts of calcite, cutting the sulfides; and as isolated grains, some of which resemble unreplaced quartz phenocrysts of the Balaklala rhyolite.

The quartz in the network surrounding sulfide grains in massive sulfide ore commonly has a characteristic feathery habit (fig. 44). The orientation of the quartz is commonly perpendicular to sulfide grain boundaries, but may have any orientation. Quartz surrounding euhedral pyrite cubes disseminated through hydrothermally altered rhyolite is commonly oriented perpendicular to the pyrite faces.

Some quartz grains in the massive sulfide ore appear to be relict quartz phenocrysts. They are isolated, poorly preserved dipyramids of quartz about 2 millimeters in diameter irregularly and sparsely distributed through the ore. Graton (1910, p. 100) also noted these relict quartz phenocrysts in the ore.

Sericite-hydromica, $KAl_3Si_3O_{10}(OH)_2$.—Sericite and hydromica occur with quartz, chlorite, and chalcopyrite

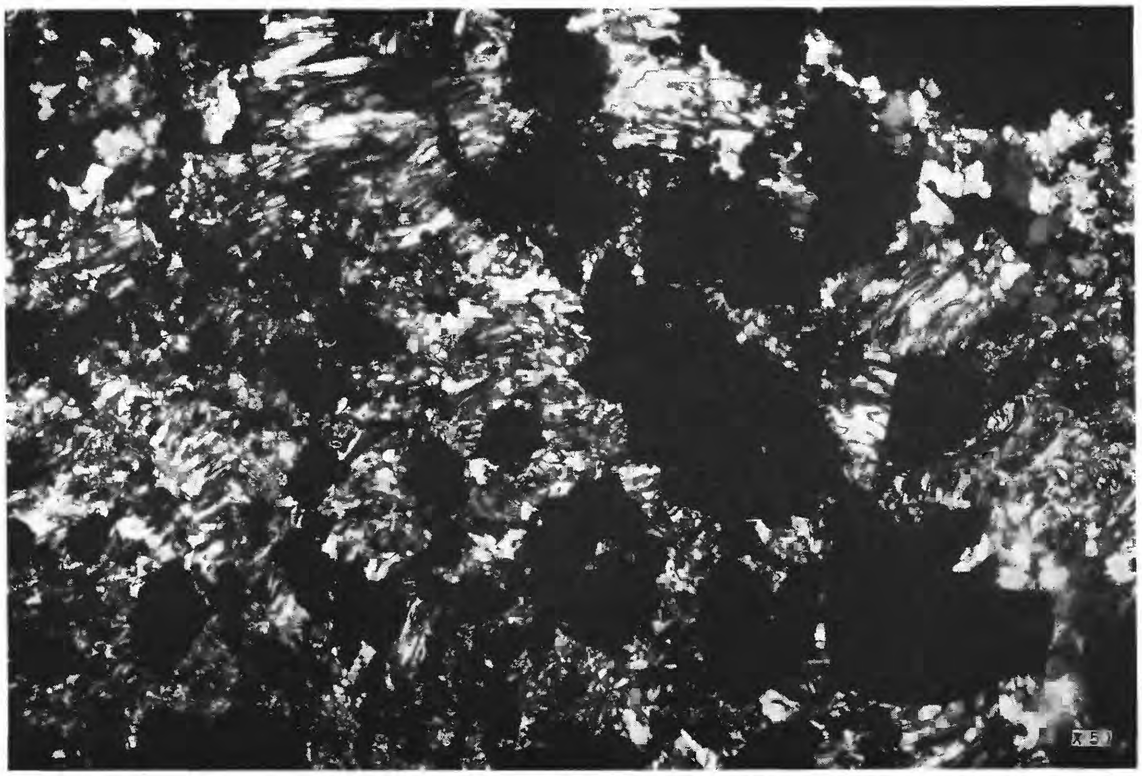

FIGURE 44.—Photomicrograph of thin section of massive sulfide ore from the Balaklala mine showing the feathery character of quartz (light) surrounding pyrite grains (dark). Crossed nicols, × 50.

in a network surrounding sulfide grains. They may be in part recrystallized rock gangue minerals. Some of the sericite and hydromica is oriented perpendicular to pyrite faces and is interlayered with feathery quartz and chlorite in flakes generally 0.2 to 0.3 millimeter long. The mica flakes are commonly gently curved, apparently by stress caused by the crystallization of pyrite.

Zoisite, $Ca_2Al_3(SiO_4)_3(OH)$.—Zoisite is a very minor gangue mineral observed in thin sections of ore from the Mammoth mine. It is in anhedral grains associated with sericite and chlorite gangue. The zoisite is probably a relict rock gangue mineral.

SUPERGENE MINERALS

OXIDE ZONE

Supergene enrichment of the massive sulfide ore has been of little economic importance in the West Shasta district except at the Old Mine ore body at Iron Mountain and possibly at the Mammoth and Golinsky mines. As the Old Mine ore body was mined out during the early part of this century, little is known of the secondary minerals that were present in the enriched zone. Only a few specimens of supergene ore were available for study by the writers. Graton (1910, p. 100–107), who worked in the district during its early history, noted azurite, chalcanthite, native copper, cuprite, goslarite, limonite, magnetite, malachite, melanterite, and smithsonite(?). Diller (1906, p. 12) noted native silver associated with native copper in fractures in the gossan. The writers also noted antlerite coating fractures in supergene enriched ore from the Old Mine ore body. Basic iron sulfate is present in the gossan quarry at the Iron Mountain mine. It forms a yellow, ocherous, earthy, loose powder coating much of the gossan. This mineral may be copiapite.

Seager in his unpublished report states that small crystals of native sulfur in the Iron Mountain gossan were observed by L. Raymond, formerly of the Mountain Copper Co., Ltd.

Black earthy manganese oxide (wad) is admixed with iron oxides in the gossan. It forms stains along fractures in the gossan and coats some of the siliceous cells of the gossan.

Maghemite, described on page 119, occurs in the gossan at the Iron Mountain mine.

Gossans of individual ore deposits are discussed under "Description of deposits." The most prominent gossans are at the Iron Mountain, Shasta King, Balaklala, Balaklala Angle Station, and the Stowell mines.

SULFIDE ZONE

Chalcocite, Cu_2S.—Small amounts of sooty chalcocite were noted coating fractures in many specimens of massive sulfide ore found on the mine dumps. Chalcocite was abundant in the Old Mine ore body at Iron Mountain, and a small amount is present in the pyritic ore below the gossan in adit 8 of the Shasta King mine. Seager reports sooty chalcocite in specimens of ore from the Mammoth, Sutro, and Golinsky mines. He states that some of the Golinsky ore consisted almost entirely of black crumbly chalcocite enclosing a few residual grains of pyrite, sphalerite, and tennantite.

Supergene enrichment took place by the selective replacement of chalcopyrite and sphalerite by chalcocite. This mineral occurs as minute veinlets in fractures in chalcopyrite and sphalerite, but is absent where the fractures cut pyrite.

Covellite, CuS.—Covellite is intergrown with chalcocite in veinlets through chalcopyrite and sphalerite, and along grain boundaries. Covellite commonly forms a thin film on the outside of thin supergene veinlets, whereas chalcocite is in the center of the veinlets.

Secondary silver minerals.—The sandy disintegrated sulfides immediately below the gossan in the Old Mine ore body at Iron Mountain contain abundant secondary silver minerals. This zone was mined as a silver ore in the early days of the district; it is described under the "Iron Mountain mine."

PARAGENESIS

Six stages of mineralization are recognized in the formation of the massive sulfide deposits. The first five stages are hypogene; the sixth stage is supergene. The first stage is earlier than the massive sulfide ore bodies and consisted of silicification, deposition of some disseminated pyrite, and the formation of sericite, hydromica, and chlorite in the wall rocks. The next five stages produced the massive sulfide ore. The stages are as follows:

(1) Early barren stage.
(2) Pyrite stage, containing small amounts of pyrrhotite and magnetite.
(3) Chalcopyrite-quartz stage, containing small amounts of sphalerite.
(4) Sphalerite-chalcopyrite stage, containing small amounts of tetrahedrite, tennantite, and galena.
(5) Quartz-calcite stage.
(6) Oxidation and enrichment stage.

Early barren stage.—The early barren stage can be divided into three substages—(1) a widespread introduction of quartz-pyrite, (2) an apparently later mineralization of sericite-hydromica that is more closely associated spatially with ore, and (3) a chloritic alteration.

The quartz-pyrite substage affected a large part of the lower and middle units of the Balaklala rhyolite, and, in places, the lower tuffaceous zone at the base of the upper unit. Euhedral pyrite cubes and pyritohedrons, commonly ranging from 1 to 2 millimeters in diameter, are disseminated through the Balaklala in amounts generally less than 10 percent by weight. In addition milky quartz that contains pyrite formed stringers 3–4 inches thick. The weathering of the disseminated pyrite and stringers of quartz-pyrite forms the widespread iron-staining in the rhyolite. This substage is so widespread that it cannot be used as a guide to ore, except as it marks a broad zone in which hydrothermal solutions were present.

Intense sericite-hydromica alteration of the Balaklala rhyolite is localized near ore bodies, as observed at the Iron Mountain, Balaklala, and Mammoth mines, but it is not necessarily adjacent to ore nor does it assure the presence of ore. It does not form alteration halos around ore bodies in the manner of those described by Sales and Meyer (1948) at Butte. The solutions that deposited the massive sulfide ores apparently followed more restricted channels than the solutions that formed sericite-hydromica in the rocks, and the earlier and later solutions did not always follow the same channels. Where sericite and hydromica are formed, the rhyolite is altered to a soft white crumbly mass that contains small amounts of disseminated pyrite. This alteration zone has no systematic relationship to massive sulfide ore; it may be adjacent to the ore or may be several hundred feet away in any direction. Hard unaltered rhyolite separates the sericite-hydromica zone at some places from massive sulfide ore. At the surface the zone of sericite-hydromica alteration locally has a lavender color owing to the weathering of pyrite to hematite, and the selective adhesion of the hematite to the sericite. The sericite-hydromica zone is probably later than the disseminated pyrite minerals as the sericite-hydromica has a more local distribution and closer association to massive sulfide ore bodies. Also sericite-hydromica is localized in some places along the boundaries of pyrite cubes.

Chloritic alteration that probably preceded sulfide mineralization occurs with the disseminated chalcopyrite ore at the No. 8 mine at Iron Mountain, in the U. S. Geological Survey diamond drill hole at Iron Mountain, and at the west end of adit 6 at the Shasta King mine. This chloritic alteration of the rhyolite is not widespread, but at some places seems to be associated with disseminated chalcopyrite. Possibly it formed a rock particularly favorable for the later deposition of chalcopyrite.

Pyrite stage.—Pyrite, was the first mineral deposited in the formation of the massive sulfide deposits, and the bulk of the pyrite was deposited before the mineralization of the chalcopyrite-quartz stage. Large bodies of extremely fine grained pyrite, which contain a few euhedral 1- to 5-millimeter pyrite cubes and pyritohedrons, were formed by the replacement of Balaklala rhyolite. A thin network of quartz and sericite-hydromica, which probably are minerals formed from relict rock gangue, remains between individual pyrite grains. This network of quartz and hydromica averages about 25 percent of the mass by volume but the network was largely replaced by minerals of the chalcopyrite-quartz stage.

Small amounts of magnetite and pyrrhotite were probably deposited during the pyrite stage, although their relationship is not definitely known. Magnetite occurs as minute anhedral grains disseminated through the quartz-sericite-hydromica network. Pyrrhotite was observed only in one specimen from Iron Mountain; it occurred with pyrite and was veined by sphalerite.

The large pyritic bodies probably formed after the minerals of the barren stage, as veinlets of massive sulfide ore cut and corrode relicts of spongy-textured pyrite-quartz rock that formed during stage (1) in the gossan quarry at Iron Mountain. The euhedral pyrite cubes and pyritohedrons that are disseminated through the massive sulfide ore are similar in size and appearance to the disseminated pyrite grains and may be unreplaced coarse-grained pyrite of the barren stage. Much silica but no copper and zinc ore minerals accompanied the early barren stage; little silica, much pyrite, and possibly minor amounts of ore minerals accompanied the pyrite stage, which is much more localized than the early barren stage of mineralization.

Chalcopyrite-quartz stage.—The pyritic bodies and enclosing rocks were fractured before the minerals of the chalcopyrite-quartz stage began to deposit. Quartz crystallized before the other minerals in this stage, and filled fractures in pyrite and in the gangue network between pyrite grains. Chalcopyrite and minor amounts of sphalerite were deposited after quartz, as these minerals contain corroded relicts of quartz, but spatially they are closely associated with it. Chalcopyrite greatly predominates over sphalerite in this stage and minor amounts of pyrite accompanied the quartz and chalcopyrite. The minerals of this stage mainly replace the selvage network of rock gangue around pyrite. There was little corrosion of the pyrite, so a network pattern surrounding pyrite grains was largely retained, but with quartz, chalcopyrite, and sphalerite replacing the network of gangue minerals. The disseminated chalcopyrite and chalcopyrite-quartz veins at the No. 8 mine that underlie the massive sulfide ore bodies at Iron Mountain were probably formed in this stage.

Sphalerite-chalcopyrite stage.—The minerals deposited during this stage listed in decreasing order of abundance are sphalerite, chalcopyrite, galena, tetrahedrite, and tennantite, but there is no clear break at some places between the chalcopyrite-quartz stage (3) and the sphalerite-chalcopyrite stage (4). Sphalerite was deposited in small amounts in the chalcopyrite-quartz stage, but it is more abundant than chalcopyrite in the sphalerite-chalcopyrite stage. The two stages are evident at some localities where one or the other greatly predominated.

Sphalerite forms streaks, lenses, and veinlets cutting the massive pyrite ore, and forms isolated pods of high-grade zinc ore, as in the Yolo zinc lens at the Mammoth mine, that replace rhyolite. Sphalerite had an extremely corrosive effect upon pyrite in this stage, and the streaks and lenses of sphalerite in massive pyritic ore contain only a few percent of corroded relict pyrite. Sphalerite and chalcopyrite were deposited contemporaneously, as they have mutual contacts, and each mineral contains corroded relicts of the other. The tiny blebs of chalcopyrite in sphalerite may have formed by unmixing.

Either tetrahedrite or tennantite is found as small inclusions in sphalerite. It has mutual contacts with sphalerite and is veined by galena. Probably the tetrahedrite or tennantite were deposited near the end of the sphalerite-chalcopyrite stage but before deposition of galena.

Galena was the last mineral to deposit in this stage. It veins sphalerite and is present mostly along boundaries between sphalerite and gangue.

Quartz-calcite stage.—A small amount of quartz and calcite was deposited after the sulfide mineralization. Both occur as small, sharp-walled veinlets cutting massive sulfide ore and as small irregular masses in sulfide ore.

Oxidation and enrichment stage—Supergene enrichment was of little importance in the West Shasta district except in the Old Mine ore body at Iron Mountain and the gossan ore body at the Mammoth mine. The enrichment at these mines is described under "Iron Mountain mine" and "Mammoth mine." The enriched ore was mined out many years ago, so a mineralogic study could not be made.

RELATIONSHIP OF HYDROTHERMAL ALTERATION TO MASSIVE SULFIDE ORE

Hydrothermal alteration processes that are believed to be genetically related to ore deposits are silicification, sericitization, and chloritization. However, at many places no close spatial relationship was recognized between hydrothermally altered rock and massive sulfide deposits. Evidence indicates that the most probable sites for the deposition of sulfide minerals are in areas of hydrothermal alteration where the rocks of the district were open to solution travel, but the evidence does

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not indicate that sulfide minerals necessarily are present. The larger areas of hydrothermal alteration are shown on plate 4.

Intense silicification generally is spatially related to sulfide mineralization, but is too widespread to be a direct guide to massive sulfide ore. Although largely limited to the main mineral belt, silicification is not necessarily more intense in areas where massive sulfide ore was deposited. At Iron Mountain, relict nodules of rhyolite in gossan are silicified, and bodies of secondary quartz that contain as much as 50 percent euhedral pyrite cubes and pyritohedrons are present as relicts in gossan derived from massive sulfide ore. These pyrite-quartz relicts are believed to represent more siliceous areas, probably formed during the early barren stage of mineralization, that were replaced by early pyrite. Wall rock that contains pyrite and silica of the early stages occurs along the side of the ore bodies at Iron Mountain; at some places it is in contact with massive sulfide ore, but at others the silicified rock is separated from ore bodies by rhyolite that contains no secondary silica or pyrite.

The porphyritic rhyolite immediately adjacent to most of the ore bodies of the district is only slightly silicified, and relict nodules of nonporphyritic and porphyritic rhyolite in massive sulfide ore, such as those in the Shasta King and Balaklala mines, are not silicified. However, quartz-pyrite bands are common near or between many of the ore bodies of the district, and the transition zone of tuff at the base of the coarse-phenocryst rhyolite at the Mammoth mine is strongly silicified in the mine area, but is not silicified away from it.

Bands of quartz and silicified rhyolite that contain from a few percent to 50 percent pyrite are abundant in the lower and middle units of the Balaklala rhyolite. These bands are more abundant near or between ore bodies, and they may be useful in delimiting general areas favorable for ore deposits.

Localities where the rocks have been altered to sericite-hydromica commonly occur along the main ore zone, but sericitic rocks are formed by regional metamorphism as well as hydrothermal metamorphism, and it is not always possible to distinguish between the two processes. In the mineral belt sericitic rocks are commonly associated spatially with silicified rocks, although the rocks are either sericitized or silicified but not both. These zones of soft, crumbly, intensely sericitized rhyolite are mainly concentrated in the general area of known ore deposits. All gradations exist between slightly altered rhyolite and a rock containing relict quartz phenocrysts in a secondary matrix of quartz, sericite, hydromica, hematite, and pyrite, or limonite pseudomorphous after pyrite. All the sericitic rock contains minor amounts of disseminated pyrite in the form of euhedral 1- to 2-millimeter cubes, and seams and flakes of hydromica intimately mixed with sericite. The color of the soft, altered rock is predominantly white but ranges from white to light pink and shades of lavender, red, and purple. Sericitic alteration is particularly common around the Iron Mountain and Balaklala mines, but some areas of strongly sericitized rhyolite occur away from known ore bodies. Much of the porphyritic rhyolite about 1 mile south of Iron Mountain on the north and east sides of South Fork Mountain is altered to a crumbly pink to lavender rock, by hydrothermal solutions. Although the area is one of strong hydrothermal alteration, no massive sulfide deposits have been found.

Owing to hydrothermal alteration and weathering the color of the altered rhyolite ranges from pink to lavender. This color is found only in porphyritic rhyolite that has been hydrothermally altered to a rock containing sericite, hydromica, and pyrite. The pink is due to selective adhesion of minute hematite specks, derived from oxidation of pyrite, on sericite and hydromica. Lavender-pink altered rocks occur on South Fork Mountain, Iron Mountain, and at the Balaklala and Mammoth mines. In some places the whole rock is uniformly lavender-pink. Under the microscope the color is seen to be due to minute flecks of hematite disseminated through the groundmass, and to thin coatings of hematite along cracks that cut across schistosity and cut through sericite veinlets. At other localities only quartz and feldspar phenocrysts are colored. In these, the lavender-pink color is due to thin coatings of hematite surrounding phenocrysts, and in cracks through phenocrysts. Pyrite casts and dark-red hematite are disseminated through the groundmass. The color of the quartz and feldspar phenocrysts is caused by migration of ferric iron due to weathering of pyrite.

The lavender-pink alteration is commonly found along the main ore zone because it forms only where the rocks contain sericite, hydromica, and pyrite, and these altered areas are commonly associated with ore.

The mineralogy of the soft pink, lavender, and purple rhyolite is the same as that of the soft white altered rhyolite except for the amount of hematite present. Quartz is the only relict primary mineral. Albite is altered to sericite and hydromica. The sericite has the following optical properties:

$$n_\alpha = 1.565 \pm 0.003$$
$$n_\gamma = 1.598 \pm 0.003$$

birefringence=0.033

Hydromica is present in nearly equal amounts with sericite and is intergrown with the sericite. It has the following optical properties:

$$n_\alpha=1.560\pm0.003$$
$$n_\beta=1.569\pm0.003$$
$$n_\gamma=1.580\pm0.003$$

birefringence=0.020
elongation=positive

At some places it is difficult to determine how much of the sericitelike alteration of the surface rock is due to hydrothermal solutions and how much is due to acid solutions formed from weathering of the sulfide minerals. Rock that lies downhill from oxidizing ore bodies is partly altered to claylike products and may have been altered mainly by acid surface waters. Where such rock is most intensely altered, it can be crumbled apart in the hand; the disintegration of the rock is largely due to alteration of feldspar to clay minerals by reaction with acid solutions derived from weathering of sulfide minerals, and in the field such argillic alteration could not be distinguished from sericitic alteration at some places.

Chloritic alteration is due largely to regional metamorphism, but some chlorite appears to be of hydrothermal origin. All the Balaklala rhyolite was replaced by chlorite to some extent during regional metamorphism, but at Iron Mountain a more intense chloritic alteration seems to be genetically related to the deposits of disseminated chalcopyrite. Seager in his unpublished report noted chloritic alteration that is related to chalcopyrite mineralization at Iron Mountain in the disseminated chalcopyrite deposits of the No. 8 mine ore body. Rhyolite in core from the Geological Survey diamond-drill hole at Iron Mountain is intensely altered at places to a secondary chlorite-quartz rock and chalcopyrite is more common in the chloritized rhyolite than in the siliceous rhyolite. Although the chlorite may have been introduced by the solutions that deposited the chalcopyrite, it seems more probable that the chlorite was earlier and formed a favorable host rock for chalcopyrite deposition.

GENESIS OF THE HYPOGENE ORES

The base-metal ore bodies and the widespread disseminated pyrite in the West Shasta district were formed by the replacement of parts of certain favorable zones in the Balaklala rhyolite. The evidence for a replacement origin is indicated by relict rock structures in the mineralized areas, and by the detailed control of mineral localization along preexisting rock structures such as foliation. Evidence for this mode of origin for the disseminated pyrite is convincing, and a replacement origin was recognized by all the geologists who have worked in the district. There are many features of the massive sulfide bodies for which the writers can offer no adequate explanation. The most puzzling feature of the ore deposits is the presence of sharply bounded bodies of massive sulfide that contain gold, silver, copper, and zinc in a general zone of disseminated pyrite that contains little or none of the ore minerals. The absence of relict rock or other evidence of replacement in the massive sulfide ore except for a few poorly-preserved quartz crystals that appear to be relict quartz phenocrysts cannot be adequately explained. The lack of evidence for a mode of origin other than replacement for these massive-sulfide bodies does not warrant the assumption that their formation was by replacement.

Evidence favoring a replacement origin for pyritized rock, as distinguished from the massive-sulfide deposits, is abundant; the effectiveness of pyritization ranges from the formation of a few scattered pyrite cubes in otherwise unaltered rock to bands of quartz-pyrite rock. The disseminated pyrite is euhedral, and the cubes and pyritohedrons are rarely in contact with each other; even in the quartz-pyrite rock the pyrite does not tend to grow in clumps around a center of crystallization or to form bodies of massive pyrite.

The quartz and heavily disseminated pyrite that form the quartz-pyrite rock commonly occur in poorly defined bands and lenticular bodies. These are oriented parallel to preexisting rock structures such as foliation, bedding, or flow contacts, but the outlines of the pyritized areas are not sharp; they fade out into barren rock. Pyritized rock is concentrated in the flat-lying ore zone, and is almost without exception sharply bounded at the top by the base of the coarse-phenocryst rhyolite, which forms the "cap rock." The lower boundary and the edges of the pyritized zone are hazy.

Bodies of massive sulfide, on the other hand, are sharply bounded by a thin band of gouge at almost all localities; the sharply bounded lenses of sulfide are generally in contact with rocks that contain little or no pyrite as described under "Types of contacts between ore and wall rock." Rock containing sparse, scattered pyrite grades to heavily pyritized rock, but gradations from heavily pyritized rock to a massive sulfide ore, except at the ends of a few ore bodies, were seen at only one locality, in the Shasta King mine.

Interpretation is equivocal that most structures in the massive sulfide ore are actually formed by replacement, such as banding in the ore, which is interpreted as replacement along planes of foliation. The massive sulfide ore contains a few poorly preserved quartz dipyramids that are probably relict quartz phenocrysts from the rhyolite and some anhedral quartz grains that may be corroded phenocrysts. Gavelin also notes relict quartz phenocrysts in the massive sulfide ore at Bjurliden (Gavelin, 1939, p. 99–102). Thin sections of massive sulfide ore show that at some places the pyrite grains are surrounded by thin films of quartz, sericite, and chlorite, which represent recrystallized host rock material. At the few localities where there is a gradational contact from massive sulfide to a foliated sericitic wall rock at the ends of some ore bodies, a complete gradation from pyritized rock to massive sulfide can be traced, and the observer has no doubt that the sulfides have replaced the rock at these points.

A further difference between massive sulfide ore and zones of disseminated pyrite is that most of the bodies of massive sulfide contain copper and zinc minerals and appreciable amounts of gold and silver, but the areas of pyritized rock contain only small amounts of copper minerals, no visible zinc minerals, and no gold and silver, as far as is known. In the description of the ores of the Malanäs district (ores that are remarkably similar to ores of the Shasta district) Gavelin (1939, p. 146) emphasizes the distinction in mode of origin between massive sulfide and disseminated pyrite. The evidence for this distinction is not only the sharp contact at most places between the two types, but also the difference in the distribution of the metallic elements in the two types (Gavelin, 1939, p. 181). The same distinction was made by Brownell and Kinkel (1935, p. 274) between the massive sulfide and the disseminated ores at Flin Flon, Manitoba. The sharp contacts between massive sulfide ore and schist and pyritized schist were pointed out by Emmons (1910, p. 55) at the Milan mine in New Hampshire.

The reason for the sharp distinction between massive sulfide ore bodies and zones of pyritized rock in this district is not known. No direct evidence has been found here that would suggest that the massive sulfide ore was emplaced as a concentrated plastic mass as was postulated by Ödman for the ores at Boliden (Ödman, 1941, p. 159), by Spurr (1933, p. 110–122) for the ore of the Mandy mine, and in a somewhat modified form, as advocated by Gavelin for some of the ores of the Malanäs district (Gavelin, 1939, p. 127–130). No apophyses of ore that resemble intrusive apophyses such as occur at Boliden and at the Mandy mine are found in the ores of the Shasta district, nor was ptygmatic folding seen here in the banded ores, such as was found in the ores of the Malanäs district and the Mandy mine. Ptygmatic folding in banded sulfide ore has been interpreted by Gavelin and by Spurr at Boliden and at the Mandy mine as indicating plastic flow during the intrusion of a sulfide magma, although Bastin (1950, p. 74) and Emmons (1910, p. 58) interpret similar structures as evidence that the banded ore has been crumpled after its formation by movement during regional metamorphism.

The massive sulfide ore in the West Shasta district is locally crushed and fractured to a moderate extent, but the fractures are healed by minerals of a later stage; there is no evidence that the ores were affected by dynamometamorphism.

Ödman and Gavelin recognize replacement phenomena at the ends of some massive sulfide ore bodies and these phenomena are seen in the West Shasta district also at only such places. The perplexing question of whether the evidence of replacement observed along the edges of an ore body (or body of rock) illustrates the mode of formation of the entire body, or whether in a body formed by other means, they are phenomena that are restricted to the edges, cannot be settled for the ores of the West Shasta district. The writers favor a replacement origin for the massive sulfide ore as well as for the widespread pyrite for the following reasons: (1) Replacement phenomena are evident in the zones of disseminated pyrite; (2) replacement is evident at the ends of some massive sulfide bodies; (3) paragenesis suggests that the minerals were deposited in sequence. Minerals of succeeding stages followed channels different from those followed by earlier minerals, and deposited in different areas at some places, thus ruling out a paragenetic magmatic sequence of crystallization in the strict sense; and (4) internal structures are not present that would indicate an intrusive origin.

The most probable explanation for the presence of sharply bounded bodies of massive sulfide in a zone of disseminated pyrite appears to be that some massive sulfide bodies were formed at a slightly different time and under the influence of different controls than the pyritized zones, and that some massive sulfide bodies formed in places where, for a time, the solutions were confined to restricted channels and effected complete replacement. In this district both of these controls were probably effective in some ore bodies, whereas in other ore bodies one control predominated.

Pyritized rock is so widespread in the district that it cannot be related to localized individual feeder channels. The solutions that deposited widespread disseminated pyrite along the mineral belt also carried quartz; hydrothermal quartz is less widespread than pyrite. Quartz deposited with pyrite principally where pyritization was more intense; weakly pyritized rock generally is not silicified. It seems probable that solutions that rose along a few main channels in the deeper levels spread into the rocks along zones of fracture cleavage in the rhyolite. The rhyolite rocks were probably the first that the solutions contacted that were sufficiently brittle to allow fracture cleavage to form and be preserved, and solutions are thought to have traveled along these fractured zones away from main feeders for considerable distances and deposited widespread pyrite. Solutions in the zones of fracture cleavage were widespread, and were not confined to small areas except locally. The solutions were channeled by the gouge of premineral faults, and the increased flow along these paths, together with the erratic distribution of zones of fracture cleavage, account for the differences in the intensity of pyritization.

Some areas of heavily pyritized rock probably were formed where solutions, rising along zones of steep fracture cleavage, came in contact with flat bedding-plane foliation, or impervious beds. These structures deflected the flow of solutions into more confined channels resulting in more complete replacement of the rocks. Minor increases in solution pressure at points where the flow was disrupted or confined to a smaller channel may also have aided in effecting a more thorough penetration of parts of the rocks by the ore-bearing solutions. The pyritized rock in the ore zone along the base of the upper unit of the Balaklala rhyolite was probably formed in this manner, but the formation of bodies of massive sulfide ore probably required that the channels be further confined by bands of fault gouge, or that the solutions reached limited areas in which ground preparation was particularly favorable. Such favorable areas are along the axes of some folds, where these were cut by gouge-bearing faults. These faults were thus not necessarily feeder channels, even though sulfide bodies occur in proximity to them.

Strong and widespread pyritization is not only confined to the rhyolite, but it also is confined at many places to flows of one type and at one horizon. This suggests that although structural control and ground preparation is of great importance, minor chemical differences may possibly have helped to determine which of several flows in the same area were replaced.

Studies of polished sections of the ores show that although there are overlaps in the sequence of mineral deposition, there were definite periods in which the deposition of one mineral predominated. Pyrite that formed during the period of widespread pyritization is associated with only small amounts of chalcopyrite and no sphalerite, gold, or silver as far as is known. In polished sections of ore the chalcopyrite and sphalerite not only formed after the main part of the pyrite, but a period of minor movement occurred before the introduction of these minerals, as they are present in fractures in the pyrite and in shattered pyrite.

Copper- and zinc-bearing solutions, which probably also carried the gold and silver and other minor metals and some pyrite, had a much more limited distribution than the earlier solutions that deposited principally pyrite and quartz. The later solutions followed more restricted channels, although these were confined to the central part of the ore zone and coincide largely with areas of most intense pyritization. However, some of the channels that were filled during the early quartz and pyrite stage were not reopened, others were opened only to a limited extent, and some new channels were formed. Thus while the copper and zinc minerals follow the pyrite zone fairly closely, they are not coextensive with it.

There was also a break at some places between the deposition of chalcopyrite and sphalerite. Chalcopyrite is earlier than sphalerite in most of the ore, and there was a change in the character of the solutions between the main periods of deposition of chalcopyrite and of sphalerite. The copper-bearing solutions like those that formed the disseminated ore in the No. 8 mine at Iron Mountain, deposited pyrite as well as chalcopyrite. The copper ore in the No. 8 mine contained no sphalerite, and very little gold and silver. It should be pointed out here that although the chalcopyrite-quartz stage as seen in polished sections is not accompanied by the deposition of any pyrite, the ore in the No. 8 mine contains about equal amounts of chalcopyrite and pyrite. This is either an instance where the chalcopyrite-bearing solutions followed the same channel as the earlier pyrite-depositing solutions and one was superimposed on the other, or it is a locality where chalcopyrite and pyrite were deposited from the same solution. The latter hypothesis is the most probable explanation, as the chalcopyrite and pyrite are coextensive throughout the No. 8 mine. The zinc-bearing solutions deposited no pyrite and had a very corrosive effect on preexisting pyrite. Also some high-grade zinc ore bodies at the Mammoth mine were formed along a different channel than that along which the major pyritic bodies formed, indicating a time interval between the two types of deposits.

The source of the solutions that deposited the base metals and performed the hydrothermal alteration is not known, but some discussion of their possible sources may be justified. It is possible that the solutions were derived from either the albite granite or from the biotite-quartz diorite, but it is also possible that they were solutions that were mobilized as a result of orogeny and were not necessarily of direct igneous origin.

The albite granite is locally pyritized and altered, and solution action is commonly localized along zones of foliation or along fractures. Although this pluton is not extensively mineralized, nevertheless it is mineralized at many places throughout the exposed area, most commonly along structures that apparently formed during regional metamorphism. There is no indication that mineralization or alteration is more common in the borders or in the wall rocks. The only suggestion that mineralization is related to this pluton is in the geographic location of the mineral belt at the north end of the elongate pluton and the presence of minor pyritization that is concentrated in the wall rocks at the south end. Pyrite is the most abundant sulfide mineral in the pluton, but gold-quartz veins are common within the intrusive mass, and some quartz veins contain pyrite, chalcopyrite, gold, and small amounts of galena. This mineralization apparently is not related genetically or spatially to the massive sulfide-type mineralization, and as described under "Age of mineralization," there are reasons for believing that the main period of mineralization is younger than the albite granite.

Mineralized or hydrothermally altered rocks in the biotite-quartz diorite pluton are extremely rare except in a limited area near the border of the intrusion, at the Silver Falls mine about 2½ miles northwest of the town of Igo in the Igo quadrangle. In this area silver-rich quartz-pyrite-chalcopyrite-galena veins formed along northerly trending faults in the biotite-quartz diorite. The wall rock of these veins is biotite-quartz diorite that is altered to soft silvery quartz-mica schist. Pyrite occurs only in a few thin seams in the main mass of the pluton. Mineralization is equally rare in the zone of contact metamorphosed rocks along the walls of the intrusion, or in the surrounding greenstone.

The only apparent relationship between the biotite-quartz diorite pluton and mineralization is one of broad geographic correlation. This pluton is mineralogically similar to some of the granitic intrusive rock of the Sierra Nevada, it has the same northwesterly trend, and is apparently of the same age as the granitic rocks of the Sierra Nevada. On a statewide scale there is a correlation between base-metal deposits, many of which have a high content of pyrite or pyrrhotite, along the west side of the granitic rocks of the Sierra Nevada and those of Shasta, Trinity, and Siskiyou Counties in northern California. Such a correlation does not necessarily imply a genetic relationship between granitic rocks and base-metal mineralization; it might imply only a correlation between structural conditions that favored the introduction of granitic rocks and mineral-bearing solutions.

The correlation between plutonic rocks and mineralization could also result from the catalytic effect of orogeny. Although no direct evidence of such an origin is likely to be found, the type of base-metal deposits and the accompanying hydrothermal alterations that occur in the Shasta district may be explained in this way without the introduction of material from an igneous source. If pore solutions in deeply buried rocks, or water expelled from hydrous minerals during metamorphism, were heated and mobilized by igneous intrusion and orogeny, it is probable that they could collect and transport elements for deposition at higher horizons. Thus the leaching of rocks which have a high-silica content such as the Abrams mica-schist that occurs west of this district and is thought to underlie the Copley (Hinds, 1933, p. 81), the alteration of the andesites of the Copley to greenstone, the albitization of some of the Copley rocks, and the formation of granitic gneiss and amphibolite from the Copley rocks could account for the migratory solutions of all the elements (with the possible exception of potassium) necessary to form the zones of epidotized, chloritized, silicified, and sericitized rocks in the overlying formations. Similarly, much of the iron and relatively small amounts of copper and zinc minerals and minor metals that form the ore deposits could have been derived by leaching of the rocks that underlie these deposits, possibly with additions from magmatic emanations.

Leaching of the wall rocks of the solution channels is indicated by the difference between the ores in the East and West Shasta districts. The ores of the two districts are similar except that in the East Shasta district the ores contain large amounts of gypsum and anhydrite, whereas those of the West Shasta district contain neither of these minerals. The ores of the East Shasta district are probably underlain at depth by the McCloud limestone, but no limestone is known to occur under the ores of the West Shasta district. It seems probable that the presence of gypsum and anhydrite in one district and not in the other is best accounted for by the presence or absence of limestone along the solution channels.

OXIDATION AND ENRICHMENT

Oxidation and enrichment are relatively unimportant in the West Shasta district. The flat-lying ore bodies crop out in areas of steep topography in most of the district, and except at Iron Mountain oxidized rock extends only a short distance below the surface. At Iron Mountain the Old Mine ore body crops out and dips about parallel to the hill slope; under these conditions much of the ore in the Old Mine ore body was converted to gossan, and an enriched zone was present under the gossan.

Oxidation.—Prominent gossans have formed where bodies of massive sulfide ore crop out, and broad areas of iron-stained rock occur where there is widespread pyritized rock. In these areas, the pyrite has been decomposed at the surface, and the rock is stained by transported limonite. Cavities after pyrite are generally free of limonite at the surface, but limonite-filled cavities and relict pyrite are present a few inches below the surface. Hard brown limonite pseudomorphs after pyrite cubes are common at some localities and may be picked up on the surface where soft rocks are deeply weathered. The decomposition of pyrite rarely extends more than a few feet below the surface in zones of disseminated pyrite, although limonite staining may go deeper. Where sulfide bodies form dip slopes, as in parts of the Iron Mountain ore body, most of the sulfide has been altered to gossan; but where the flat-lying massive sulfide bodies crop out in areas of steep topography, as at the Shasta King, Stowell, and Balaklala mines, the gossan capping rarely extends into the hill for more than 50 feet.

The gossan at the Iron Mountain mine is the most extensive in the district, and is well exposed in the quarry. This gossan is described under "Iron Mountain mine," but some general features that are pertinent to the oxidation of the ores of the district are described here.

The ore bodies at each end of the ore zone at Iron Mountain have been altered to gossan, where the zone is exposed at the surface; the central part lies under a ridge and is not oxidized. The horizontal Hornet ore body is oxidized at the exposed end, and at the top of the ore body where the rock cover is less than 150 feet thick. No oxidation occurred along the top of the Richmond ore body, which lies 350–450 feet below the surface, or along the top of the Brick Flat ore body, which is in some places as little as 100 feet below the surface. The Old Mine massive sulfide ore, on the other hand, was almost completely oxidized: it cropped out at the surface, and inasmuch as the dip of the ore body is parallel to the hill slope, the entire length of the ore body was exposed to oxidation. Oxidation extends to a considerable depth where a steeply dipping ore body crops out at the surface; along parts of the Camden ore body at Iron Mountain oxidized minerals extend locally to a depth of 500 feet. Where ore bodies do not crop out, as little as 50–100 feet of rock is sufficient cover to prevent oxidation of the sulfides, even though the ore body is cut by steep faults that reach the surface.

Gossan derived from massive sulfide ore is resistant to weathering and commonly forms bold outcrops. Even where the outcrop of a massive sulfide body is covered by wash, float boulders of gossan can usually be found in the debris below the covered outcrop.

The appearance of the gossan derived from massive sulfide ore is distinctive, and differs from gossan derived from quartz-pyrite rock or from heavily pyritized rock. This distinction is of economic importance because, with only minor exceptions, deposits of massive sulfide contain gold, silver, and copper and zinc minerals, whereas heavily pyritized rock contains none of these minerals. It can be assumed that gossan derived from massive sulfide ore, and which contains gold as a relict mineral, indicates the presence of copper and zinc minerals in the sulfide zone beneath the gossan. As far as the writers could determine, there are no exceptions to this generalization in the West Shasta district.

Gossan derived from massive sulfide is composed largely of hard dark-reddish-brown limonite that has an irregular, cellular texture, and contains rather widely spaced, irregularly distributed ribs of silica (fig. 45). This silica is secondary, as most of the massive sulfide ore contains too little quartz or relict-rock material between the grains of pyrite to form a quartz skeleton upon removal of the pyrite by oxidation, and the silica septa that are in the gossan have no counterpart in the primary ore. The cellular gossan containing silica septa makes up the bulk of the gossan, but dense limonite occurs both as irregular areas in the gossan and as rims surrounding nodules of relict sulfide.

Relict nodules of massive sulfide ore in gossan have sharp boundaries (fig. 46), and are commonly rimmed by a band of dense limonite 1 to 4 inches thick between the cellular limonite and the sulfides (fig. 46). It is evident that silica as well as iron was redistributed around some sulfide nodules in the oxidized zone. Bodies of relict sulfide ranging in size from small nodules to masses 10 to 20 feet long locally show a characteristic zoning and alteration of the sulfides. This alteration consists of a zone of silica at the contact between limonite and sulfides and the formation of a zone of dense limonite between the sulfide nodule and the cellular limonite. The zone of silica consists of a friable white silica sponge a few millimeters to 3 or 4 centimeters thick and contains very thin, closely spaced septa; the contact is gradational against the sulfides, but sharp against the dense limonite. Neither the sulfides nor the silica sponge are iron stained. The dense limonite forms a hard reddish-brown band around the sulfide mass, and has a sharp contact on its inner edge against the silica sponge, but a gradational contact to porous limonite at its outer edge. The band of dense limonite maintains a width of 1 to as much as 4 inches as it encroaches on the sulfides, regardless of the size of the relict sulfide mass.

Some relict-sulfide nodules contain more silica than is commonly present in massive sulfide ore. Possibly these are relicts of the early silica stage, although similar areas of high-silica content were not seen in the ore in underground workings. It seems equally possible

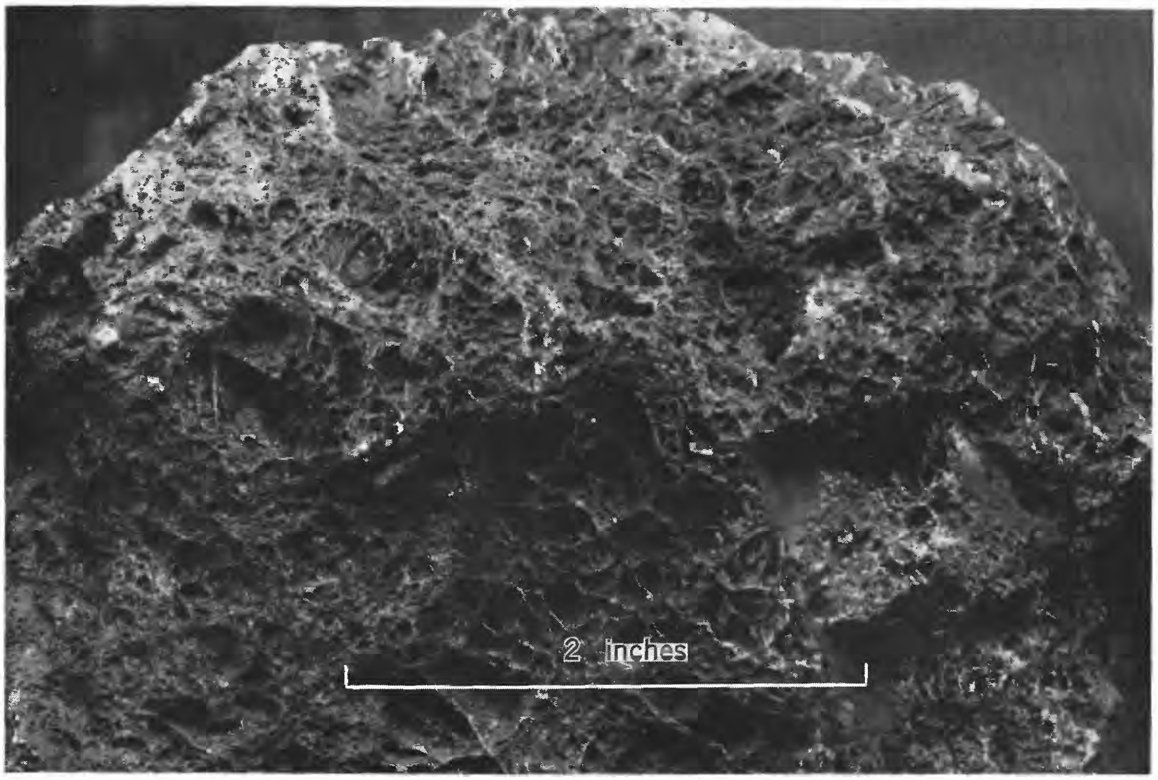

FIGURE 45.—Gossan derived from massive sulfide ore. Limonite with irregularly distributed ribs of secondary silica. The size or shape of the cells is not inherited from the primary ore.

that the silica that was mobilized during oxidation replaced some of the sulfides in local areas. Also, small nodules of soft, friable silica sponge surrounded by a rim of dense limonite are found in the gossan; these are pyrite free and resemble the friable silica sponge that formed between the sulfide and the dense limonite, which suggest that the nodules represent the last stages of pyrite removal when all the iron from the nodule had moved outward into the rim of dense limonite.

FIGURE 46.—Relict nodule of massive sulfide in gossan. Dense limonite separates sulfides from cellular limonite.

The common succession of silica sponge, dense limonite, and cellular limonite that contains ribs of secondary silica, from the center of a relict-sulfide nodule outward, indicates the sequence of oxidation. Silica was transported from the center outward, the strongest precipitation being at the edge of the sulfides, although it may have first replaced part of the sulfides. Iron was transported outward past the band of iron-free silica sponge and deposited as dense limonite. As the dense limonite-silica sponge zone encroached on the sulfides, the cellular limonite encroached on the band of dense limonite, so that the latter maintains about a uniform thickness.

No correlation was possible between types of massive sulfide gossan and the copper-zinc content of the primary ore as no detailed information is available on the grade of the massive sulfide below most gossans.

Collapse breccia was observed in the gossan at Iron Mountain, and may have been present at other mines. The decrease in volume was large where gossan formed from massive sulfide ore, as much iron was removed and large caves lined with botryoidal and stalactitic limonite have been found. Angular rock fragments 50 feet or more from the walls of the gossan in this breccia at Iron Mountain indicate a considerable collapse of the walls and roof. It is probable that some of the gossans observed at the surface in the district represent only a small part of the width of the massive sulfide ore that was present before oxidation and consequent collapse of the walls. The schistose character of much of the wall rock would allow a considerable dilation toward the gossan without an obvious appearance of collapse in the outcrop.

Enrichment.—Gold occurs in gossan as a residual enrichment, and silver and copper have been enriched in the upper part of the sulfide ore below the gossan. Zinc that was present in the primary ore has been removed. The weight of the gossan at the Iron Mountain mine is reported by the mine staff as 165 pounds per cubic foot. Jackson and Knaebel (1932, p. 125) report 100 to 125 pounds per cubic foot. The massive sulfide ore weighs 265 pounds per cubic foot. Gold in Iron Mountain gossan averaged 0.073 ounce per ton but gold in the massive sulfide ore averaged 0.04 ounce per ton, which appears to be a normal residual enrichment of about 2:1, due to loss of weight in the formation of gossan.

Silver is enriched along the contact between massive sulfide and gossan. No records remain of the silver-mining operations that were carried on in the late 1800's at the Iron Mountain mine, which is the only massive sulfide deposit at which silver ore was mined, but some assays are reported by the Mountain Copper Co., Ltd. (C. W. McClung, 1947, oral communication) to have been as high as several hundred ounces of silver per ton. The zone of silver enrichment above the Old Mine ore body at Iron Mountain was fairly extensive owing to its favorable topographic setting, but at other mines this zone is only a few inches thick. Few exposures of the enriched zone remain at any of the mines, but where it has been seen, it is a zone of sandy, disintegrated sulfides, locally stained gray by chalcocite. The sandy sulfides contain some clay and silica sand or silica sponge. The enriched zone is similar in many respects to the enriched zone over the massive sulfide ore at Flin Flon, Manitoba (Brownell and Kinkel, 1935, p. 269–270) except that it is less extensive at all the mines except the Iron Mountain mine.

Data on the enrichment in copper are available only for the Iron Mountain, Golinsky, and Mammoth mines, and supergene enrichment was of economic importance only at these mines; assay data are given in the descriptions of these mines. The topography over parts of these ore bodies favored the oxidation of the ore, resulting in appreciable enrichment in copper in the sulfide body under the gossan. The secondary copper mineral was probably largely sooty chalcocite; no copper oxides, carbonates, or silicates are present. At most of the mines the occurrence of flat-lying ore bodies in areas of steep topography did not favor oxidation, and little or no enrichment is present. Also, as noted above, ore bodies under 50–100 feet of rock cover were not affected by oxidation.

AGE OF MINERALIZATION

The base-metal mineralization, as distinct from the gold mineralization, in the Shasta district is probably Late Jurassic or Early Cretaceous in age, but no direct evidence was found in the West Shasta district to accurately date the mineralization. All the massive sulfide deposits in the West Shasta district occur in the Balaklala rhyolite of Middle Devonian age, but the Bragdon formation of Mississippian age and the Mule Mountain stock of albite granite of Late Jurassic or Early Cretaceous age are also slightly mineralized. The Upper Cretaceous rocks in the West Shasta district and the Lower Cretaceous (Paskenta and Horsetown of Anderson (1902)) rocks west of the district are not mineralized.

Base-metal ores, similar to those in the West Shasta district, occur in Balaklala rhyolite at the Greenhorn mine, 9 miles west of Iron Mountain, and in Bully Hill rhyolite and the Pit formation of Triassic age in the East Shasta district. The mineralization in the East and West Shasta districts is so similar in its distinctive mineralogy, hydrothermal alteration, and geologic setting that it would be a remarkable coincidence if it were not all the same age.

The age of base-metal mineralization can be dated fairly closely by relating it to the period of orogeny. The youngest formation near the head of the Sacramento Valley that was affected by the orogeny is the Potem formation of Jurassic age (Diller, 1906, p. 5). The oldest formation that is of postorogenic age is the upper part of the Horsetown formation, which nonconformably overlies the Shasta Bally batholith 18 miles southwest of Redding, and which is Early Cretaceous in age (Anderson, 1933, p. 1259; Hinds, 1933, p. 113). The location of many base-metal ore deposits of the West Shasta district was controlled by structures that were formed during the main period of orogeny. The ores at the Mammoth mine are in a structural arch, and those at Iron Mountain, Balaklala, and Shasta King mines are in structural troughs. The ore deposits at the Iron Mountain, Balaklala, and Mammoth mines are also spatially related to major east-west faults. The folding, and probably also the faulting, occurred during the Late Jurassic or Early Cretaceous orogeny, although the only definite dating of the faults is post-Bragdon (Mississippian) in age in the West Shasta district and post-Triassic in age in the East Shasta district. Flat bedding-plane foliation and steep fracture cleavage, formed during orogeny, are major ore controls. It therefore seems certain that the period of base-metal mineralization was later than the main period of orogeny during which these controlling structures were formed.

No intrusive bodies are present to which the ore deposits can be related with certainty, either spatially or genetically. Syntectonic stocks of albite granite of Late Jurassic or Early Cretaceous age that have the closest spatial relationship to the base-metal deposits are the Mule Mountain stock of albite granite, at the south end of the West Shasta mineral belt, and a stock of albite granite near the bridge over the Pit River on U. S. Highway 99N, but this stock is 5 miles west of the Bully Hill mine and 13 miles west of the Afterthought mine in the eastern part of the East Shasta district. Another small stock of albite granite is shown by Diller (1906) to be 6 miles southeast of the Afterthought mine. The Mule Mountain stock is cut by gash veins that have a filling of gold-quartz and small amounts of pyrite, chalcopyrite, and galena; foliated zones in the albite granite contain a little pyrite but the stock does not contain any base-metal ore. The Shasta Bally batholith of biotite-quartz diorite, which is also of Late Jurassic or Early Cretaceous age, is in the southwestern part of the Igo quadrangle, but no similar intrusive is present in the East Shasta district; ore deposits in the East Shasta district are almost 30 miles from the Shasta Bally batholith. The dikes of dacite porphyry and quartz latite porphyry that are related to the Shasta Bally batholith are not mineralized and do not show any evidence of hydrothermal alteration except where dikes of Birdseye porphyry are adjacent to later gold-quartz veins.

Some of the faults that cut the massive sulfide deposits show premineral and postmineral movement. The writers interpret the premineral movement on these faults as having taken place during the later part of the period of orogenic stress, and the postmineral movement as the minor readjustments during the relaxation of the stress. The mineral deposits are controlled by structural features formed during the orogeny, but the Cretaceous rocks, and the dikes related to the Shasta Bally batholith of postorogenic age are unaffected by this mineralization. Thus all the evidence indicates a Late Jurassic or Early Cretaceous age for the base-metal deposits of the Shasta copper-zinc district.

SUMMARY OF BASE-METAL ORE CONTROLS

Three main controls of the massive sulfide ores can be recognized in the copper-zinc district; these are the stratigraphic control in the Balaklala rhyolite, the structural control by folds and foliation, and the feeder-fissures along which the solutions ascended.

All the base-metal ore bodies in the West Shasta district are in the Balaklala rhyolite, and are limited to one stratigraphic zone in the rhyolite. This zone comprises the flows and pyroclastic material of the upper part of the middle unit in the Balaklala. Scattered pyritized rock and local areas of fairly strongly pyritized rock are found in the lower unit of the Balaklala, but no ore bodies are known in this unit. Likewise, the upper unit, of coarse-phenocryst rhyolite, is not known to contain ore bodies with the possible exception of one small ore body near the base of the unit at the Mammoth mine. All the known ore is in the middle unit of the Balaklala, and where the stratigraphy can be mapped in detail, the ore is in the upper part of the middle unit.

Ore bodies are localized along broad folds and warps, although they show little preference between anticlines and synclines. The mineral belt as a whole follows the trend of a series of broad folds that constitute a gently arched anticlinorium that has a low culmination in the central part of the mineral belt. There is good correlation between the axis of the anticlinorium and the mineral belt, and between the culmination and the central part of the mineral district. The southwest end of the anticlinorium is cut by intrusions, but at the northeast end there is little mineralization where the folds plunge beneath the overlying sediments. Individual ore bodies are localized along minor folds, basins, and small arch or dome structures.

The intersection of steep fracture cleavage and gently dipping bedding plane foliation has provided a shattered area capped by impervious material that has localized some ore bodies. The amount of foliation is dependent on the competence of the rocks and on their stratigraphic position in folds. Where much bedded pyroclastic material is present at the base of the upper unit, bedding-plane foliation has formed, but steeply foliated rocks in the lower and middle units of the Balaklala rhyolite are generally capped by unfoliated rock of the upper unit.

Some faults of premineral age have acted as channelways for ore-bearing solutions. They generally cut across the strike of folds and foliation at a considerable angle, and are influential in localizing ore bodies in certain parts of a fold. A conjunction of the three types of ore controls was probably a necessary prerequisite to the formation of a major ore body.

Two other types of ore controls for base-metal deposits can be mentioned, although their importance cannot be evaluated. One of these is the presence of a very thick cover of shale that overlay the Balaklala rhyolite at the time of ore deposition. This cover of shale was relatively impervious to solutions, as fractures in the brittle rocks below tend to die out on entering the shale. An impervious "cap rock" many thousands of feet thick over the anticlinorium would hardly fail to have a ponding or channeling effect on rising solutions.

Another possible control of ore localization is the tendency for solutions that rose along steep feeder channels to leave these channels and follow fractured zones in gently plunging folds. The importance of this type of control for ores in this district has been advocated by Walker (written communication), and the continuity of the pyritized rock (but not massive sulfide ore) at the ore zone, as indicated by surface exposures and by drill holes, lends weight to this concept. Known feeder fissures are widely spaced, although undoubtedly some have not been recognized, but pyritized rock along the ore zone occurs at considerable distances from possible feeder channels. In addition, some ore bodies have formed where no feeder fissure is recognizable. It thus seems probable that the ore-bearing solutions did travel laterally for considerable distances from the steep feeder channels in some instances, probably guided by relatively more pervious channels formed along the intersection of fracture cleavage and bedding-plane foliation in the crests and troughs of folds. Such solutions probably account for the widespread pyrite formed during the early pyrite-quartz stage.

EXPLORATION POSSIBILITIES OF THE WEST SHASTA DISTRICT

Many favorable areas where the ore zone has not been eroded remain to be explored in the West Shasta district. The areas that are worthy of exploration contain the middle unit of the Balaklala rhyolite within the limits of the main mineral belt; those that can be eliminated are areas in which the middle unit has been eroded or was not deposited, or where it is covered by a considerable thickness of younger sedimentary rocks. These covered areas are eliminated because they are off the trend of the main mineral belt and because favorable structures and mineralized zones cannot be delimited. Base-metal ore bodies may be present under the sedimentary rocks, although the main covered areas lie either on the flanks of the anticlinorium or farther down the plunge from the culmination of the folding; both of these locations are less favorable for prospecting than the higher parts of the folded structure along the trend of the mineral belt.

The areas within which it is probable that new bodies of massive sulfide ore can be found are outlined on plate 4. This map shows faults, areas of hydrothermally altered rocks, pyritized rocks, and the known ore bodies and major prospects. Plate 4 also shows the location of the exploratory drill holes that already have been drilled, with the exception of some holes drilled near known ore.

The principal feature used in determining areas in which new ore bodies may be found within the mineral belt is the stratigraphic sequence, as all known ore in the district thus far has been found in the middle unit of the Balaklala rhyolite. In addition, the largest ore bodies occur in the upper part of the middle unit, particularly where it contains much bedded pyroclastic material. No minable ore bodies have been found in the upper or lower units, although the lower unit is heavily pyritized at some places. No massive sulfide deposits have been found in the underlying Copley greenstone. The Copley is exposed in very few places along the mineral belt, and the possibility of copper deposits in the chloritic rocks of the Copley along feeder channels should not be ignored. However, the Copley is deeply buried along most of the mineral belt and evidence of feeder channels in certain areas is indicated, but not proved. In addition, the broad expanse of Copley that is exposed in the mapped area, which includes areas within the mineral belt, contains scarcely any copper minerals. No massive pyritic bodies are known in the Copley except the small bodies explored at the Akers prospect on Squaw Creek, and a few very small bodies in the southeast corner of the Igo quadrangle.[6] No ore bodies or mineralized rocks have been found in the upper unit of the Balaklala, but areas in which this unit is present are considered favorable for exploration, because the middle unit can be assumed to underlie the upper.

No lateral controls for ore can be used with certainty to eliminate areas, where the middle unit of the Balaklala is known to be present. Although most ore occurs along the crests or troughs of gentle folds or warps, some ore occurs on the flanks of folds, and these structures, while probably less favorable, cannot be eliminated from exploration.

The most favorable areas in which to search for new ore bodies are between known ore bodies along the trend of the mineral belt, which is also the trend of foliation and folded structures. Equally favorable areas, in some instances, are the extensions of folded structures beyond known ore bodies. Other particularly favorable, but less obvious, areas lie along the trend of fissures that appear to be main feeder channels, where these fissures cross folds in the productive zone.

Less favorable areas are where the middle unit of the Balaklala rhyolite is thin, or where the upper part of the middle unit has been eroded. Areas where the rocks are closely folded are less favorable than those in which the dips are gentle.

[6] Reconnaissance in the unmapped area southeast of Mule Mountain suggests that prospecting for base-metal ore bodies in the hydrothermally altered rhyolite in this area is warranted.

Some bodies of rocks that are lithologically similar to those of the middle unit are shown on the geologic maps, but they are not included in the favorable zones because they are intrusive rhyolite or because they are off the trend of the main mineral belt and are not mineralized or hydrothermally altered.

Some exploration holes have been drilled along the trend of the ore zone, particularly between the Keystone and the Stowell mines and at several other localities. Many of these holes penetrated the ore zone and cut pyritized rock, but none penetrated massive sulfide ore, as far as is known. The position of the drill holes is shown on plate 4. Some of the holes were drilled in areas which the writers consider the most favorable geologically, and they reduce the possibility of locating new ore in these areas; they do not eliminate the areas because most of the holes are widely spaced, and because ore bodies may occur on either side of a line of holes that follows the main ore trend.

Plate 4 shows the particularly favorable places within the more promising areas in which to start exploration for new ore bodies. These areas are not listed in order of favorability.

(1) The area between the Balaklala and the Keystone mines. Drilling already done indicates that there is a good possibility of bodies of massive sulfide ore between these two mines.

(2) East and southeast of the Balaklala mine. Considerable drilling has been done in this area but the possibilities are not exhausted.

(3) The area between the Keystone and Stowell mines. A line of drill holes cut the ore zone at a moderate depth, and additional drilling should be done northwest and southeast of this line of holes.

(4) Southwest of the Early Bird mine.

(5) The area immediately north of the Shasta King mine should be prospected for a continuation of the Shasta King ore zone. This could be done by underground exploration from the workings of the Shasta King mine.

(6) The hydrothermally altered rocks north of the anticline which has a core of Copley greenstone (2,000 feet northwest of the Brick Flat and Richmond ore bodies at the Iron Mountain mine) should be explored.

(7) The area under Sugarloaf Mountain between the Busy Bee workings at the northeast end of the Iron Mountain ore bodies and the Sugarloaf mine probably contains massive sulfide bodies. However, the character of the ore in the Busy Bee workings and the Sugarloaf mine suggests that these sulfide bodies contain very little copper or zinc, and the topography makes exploration by drilling difficult.

(8) Southwest of the Mammoth mine. The contact between the upper and middle units of the Balaklala rhyolite may be expected to flatten southwest of the Friday Lowden workings of the Mammoth mine, and the ore bodies in the Mammoth mine are larger where the contact has a gentle dip than where it has a steep dip, as in the Friday Lowden workings.

(9) The area immediately northeast of the Golinsky mine.

(10) Northwest of the Sutro mine.

(11) Southwest of the Balaklala Angle Station gossan.

DESCRIPTION OF DEPOSITS

AKERS PROSPECT

The Akers prospect (pl. 4) is in the canyon of Squaw Creek in secs. 6 and 7, T. 33 N., R. 5 W., at an altitude of 1,100 feet; the portals of the several adits are about 300 feet west of the southeast corner of sec. 6. A pyritized shear zone oriented N. 10° W., 48° E. cuts the Copley greenstone and intrusive rhyolitic dikes at the adits. Some massive pyrite and pyritized rock is on the dumps. Another adit is 700 feet north of Squaw Creek at an altitude of about 1,400 feet. The only information on the ore or on the underground workings is that given by Tucker (1926, p. 144). He reports, in part:

> The ore occurs in small irregular lenticular ore bodies, along irregular fissures, one of which trends north and south, with a dip of 60° east, and the other having a N. 40° W. trend. The ore is chiefly pyrite with more or less chalcopyrite and occasional traces of bornite, and carries $2 per ton in gold and silver. The present work is confined to No. 3 tunnel, where a series of parallel north and south and a N. 40° W. fracture are being drifted on; and some small lenses of ore have been exposed along these fractures, varying in width from 2 inches to 2 feet, and from 10 to 15 feet in length. . . . On the claims located on the south side of Squaw Creek are two tunnels which have lengths of 100 feet. In the lower tunnel, which is driven on a N. 40° W. fracture, a small lens of ore 40 feet in length and about 2 feet in width has been developed. Samples taken from this orebody are reported to carry from 2 percent to 6 percent copper.

The prospect has been inactive since shortly after 1926.

BALAKLALA MINE

The Balaklala mine is located in secs. 11, 12, 13, and 14, T. 33 N., R. 6 W., on the southern slope of the rugged canyon of Squaw Creek at an altitude of 2,250 feet in the central part of the West Shasta copper-zinc district. The mine is owned by W. A. Kerr. It was accessible by car in 1950 along a steep one-way dirt road that started from the site of the old smelter town of Coram on the Sacramento River. Washouts from the winter rains make this road impassable each year, although a few days work with a bulldozer is usually sufficient to make the road passable again. The mine has not operated since 1928; a fire in 1924 destroyed the mine plant, and another fire in 1931 destroyed the tram terminal and ore bunkers. The 16,500-foot aerial tramway on which ore was once hauled from the mine to a smelter on the Sacramento River is down, and the towers have burned. The Balaklala smelter at Coram on the Sacramento River was closed in 1911 because of smoke-damage suits and was dismantled in 1926. Recent salvage operations for scrap metal have removed the vestiges of the mine plant.

W. A. Kerr owns a group of claims that contain several individual ore bodies. These are in the Balaklala mine: the Balaklala Angle Station gossan; the Early Bird mine; and the Vulcan (Great Verde) prospect (pl. 4). The Balaklala mine contains two large ore bodies, the Windy Camp, and its faulted extension, the Weil, and several smaller ore bodies and prospects.

Although the Balaklala ore bodies were among the early discoveries in the district, the date is not known. Information on the history of the mine is mainly from reports of the California Division of Mines, but also from an unpublished report by G. F. Seager and from private mine reports.

The mine was first named the Balaklava mine, and is listed as a "quartz" mine in the report of the California State Mineralogist for 1894 (Anon., 1894, p. 245), but by 1896 the name had been changed to Balaklala, and it has been subsequently known by that name. Adits were driven into the gossan at the outcrop of the Windy Camp, Weil, and Balaklala Angle Station gossans from 1890 to 1900. The Windy Camp ore body cropped out as a large gossan, but its faulted extension, the Weil, cropped out only as fringes of gossan along the northeasterly edge of the ore body. The Weil ore body was located mainly by drilling. Many smaller ones did not crop out and were located by drilling.

The Balaklala mine was operated by the Balaklala Mining Co. of San Francisco, but was leased by them to the Western Exploration Co. in 1902. The latter company explored the mine, and by early 1902, extensive drilling and 3,500 feet of drifting and crosscutting had been done. The property was taken over in 1905 by the First National Copper Co. and operated under the name of the Balaklala Consolidated Copper Co. This company operated the mine from 1906 to 1911, and from 1914(?) to May 1919, when work was suspended. The mine was leased by the Mammoth Copper Co. (United States Smelting Refining and Mining Co.) and was reopened in November 1923. No mining was done until 1924, but during 1924–25 some ore was removed from the Windy Camp and High Grade ore bodies. In March 1926 the Balaklala mine was leased to the Mason Valley Mining Co. This company mined 5,000 tons of ore from pillars in the Windy Camp and North Weil 170 ore bodies from March 1926 to May 1928, when the mine was again closed. It has not operated since 1928. The mine has been inaccessible for many years, but the portal of tunnel 8 from the opencut was reopened by a private company in 1948. Bad air and caved stopes allowed only a limited examination of the underground workings at this level.

Production and grade of ore.—About 1,200,000 tons of ore has been produced from the Balaklala mine, but the exact total is not known. Estimates range from 900,000 to 1,200,000 tons. Table 7 gives the data on production and grade of ore compiled by the writers from many sources. The greatest average daily tonnage produced from the mine was 300 tons per day in 1918.

GEOLOGY OF THE MINE AREA

Stratigraphy.—The rocks in the mine area include the lower part of the upper unit of the Balaklala rhyolite and the upper part of the middle unit of the Balaklala. The upper unit ("cap rock") includes massive coarse-phenocryst rhyolite and the pyroclastic rocks of the transition zone between the upper and middle units of the Balaklala. This massive coarse-phenocryst rhyolite, which overlies the ore zone, is well exposed in the

TABLE 7.—*Production and grade of ore from the Balaklala mine* [1]

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Ore body or periods of mine operation	Production (short tons)	Grade						
		Gold (ounces per ton)	Silver (ounces per ton)	Copper (percent)	Zinc (percent)	Iron (percent)	Sulfur (percent)	Silica (percent)
Weil [1]	720,000	0.025	0.8	2.75	1.3			
Windy Camp	295,000	.02	.8	2.50				
North Weil 170		.03	1.0	3.4	1.2			
High Grade		.10	5.0	12.0	1.2			
9, 83, 84, 153		.025	1.0	2.85	1.6			
Bull	2,000 (approx.)	.048	4.0	10.0	1.2			
North Weil 3		.025	.8	2.75	1.3			
Windy Camp and High Grade (mined by Mammoth Copper Co.).[2]		.026	.96	2.56	2.2	29.0	39.0	24.4
Balaklala mine operation, 1915–19 [3]	351,171 [4]	.031	1.0	3.08				
Weil (assays by Calif. Div. Mines) [5]		.0228	.82	2.734		29.1		29.8
Windy Camp (assays by Calif. Div. Mines) [5]		.025	.85	2.46		29.8		23.5
Balaklala ore bodies (assays by Calif. Div. Mines) [6]		.03	.9	2.7	2.2	31.5	35.2	21.4 {4.3 CaO; 3.4 Al_2O_3}
Pyritic ore remaining in mine (estimated grade) [2]		.02	.86	2.46		34.8	39.4	18.3
Total	1,200,000 (estimate)	.028	1.0	2.89	1.3(?)			

[1] Source of data, Seager, G. F. (unpublished report 1934, quoting S. A. Holman, manager and general superintendent of mine), unless otherwise noted.
[2] Data from United States Smelting Refining and Mining Co.
[3] W. A. Kerr, Balaklala Consolidated Copper Co.
[4] This figure includes ore from some of the ore bodies above.
[5] Calif. State Min. Bur., 1908, Bull. 50, p. 90.
[6] Calif. State Min. Bur., 1926, v. 22, p. 145.

FIGURE 47.—View of opencut at Balaklala mine showing part of the middle unit of the Balaklala rhyolite. Coarse pyroclastic layer (*a*) with rhyolite fragments as much as 12 inches in diameter; white porphyritic rhyolite (*b*) with 2-millimeter quartz phenocrysts; nonporphyritic rhyolite pyroclastic bed (*c*); mineralized rhyolite (*d*) that contains a few small quartz phenocrysts and is strongly altered; massive sulfide ore (*e*). Note the irregular contact to ore.

opencut where the overlying rocks have been dropped to the level of the ore zone by the Balaklala fault, and is also exposed on the road from the opencut to the portal of the Weil tunnel, and on the ridges above the opencut. The coarse-phenocryst rhyolite is a hard light-gray to white rock in unweathered exposures, but weathers to soft, punky light-tan rock. The rock is massive and essentially structureless, except for a local formation of what appears to be flow banding. The phenocrysts observed in hand specimens are predominantly quartz, although locally, as in the area north and west of the portal of the keystone mine, feldspar phenocrysts are prominent. In thin sections, feldspar and quartz phenocrysts are seen to be equal in amount, but weathering makes the feldspar difficult to recognize. The phenocrysts are abundant and range from 3 to 5 millimeters across.

The massive coarse-phenocryst rhyolite in this area is underlain by shaly tuff and coarse and fine pyroclastic material, which is exposed above the ore in the glory hole (fig. 47) and on the ridges south and southwest of it. The different types of pyroclastic material have little horizontal continuity, but the zone as a unit is continuous. The pyroclastic material ranges from a thin black shale containing minor tuff layers, to tan, shaly tuff and thin beds of lapilli tuff that have a maximum thickness of 40 feet, 2,500 feet east of the mine, to coarse lapilli tuff mixed with coarse volcanic breccia that contains rounded fragments 6 inches in diameter. A few of the larger fragments have lighter or darker colored rims and appear to be volcanic bombs. Some fragments of coarse-phenocryst rhyolite occur in the volcanic breccia, and chips of quartz phenocrysts as much as 4 millimeters across are common in layers of crystal tuff. Some of the bedded material appears to be arkosic.

The contact between the massive coarse-phenocryst rhyolite and the underlying rocks of the transition zone east of the opencut is sharp, but to the south, the massive rhyolite thins abruptly and interfingers with tuff. The facies change occurs between the opencut and the summit of the ridge south of the Keystone mine. Shaly and rhyolitic tuff layers are interbedded with coarse-phenocryst porphyry flows on the ridge east of the Balaklala mine. The contact between the upper and the middle units of the Balaklala rhyolite is at the base of the transition zone.

The upper part of the middle unit of the Balaklala in the mine area is composed of porphyritic rhyolite containing 1- to 3-millimeter quartz phenocrysts, volcanic and flow breccia that contains fragments of porphyritic and nonporphyritic rhyolite, and flows of nonporphyritic rhyolite.

Geologic structure of the mine area.—The rocks in the mine area have been gently folded and warped and are cut by many faults. A poorly developed steeply dipping foliation has been superposed on the rocks of the middle unit. Dips as high as 75° have been measured in bedded tuff near the mine, but dips of 10°–30° are more common. The folds could not be mapped in detail because of the limited exposures of bedded material and the lack of sufficient underground data from mine workings and drill holes. However, several small synclines and anticlines were recognized on the basis of bedding in tuff and from the outcrop pattern, but little parallelism between the axes of these small folds or warps is apparent. Dips on bedded rocks at the surface near the Windy Camp–Weil ore body indicate that the flows that contain this ore body have a basin shape in the mine area. Two similar folds occur east of the ore body, but these folds appear to be offset by the faults in the two gulches immediately east of the Balaklala mine and could not be traced.

Many faults are present in the underground workings; the larger ones are shown on the map of underground workings (pl. 5). Most of these faults have a small offset and cannot be recognized at the surface. The Balaklala fault, however, has a vertical displacement of about 220 feet, and can be traced west of the mine with fair accuracy for 1¾ miles. It is a steeply dipping normal fault that is downthrown on the north.

ORE DEPOSITS

General description.—The mine was inaccessible in 1934 when G. F. Seager was doing geologic work in the West Shasta copper-zinc district, but he was able to obtain a great deal of detailed information from S. A. Holman, who was manager and general superintendent during much of the time when the mine was operating.

The ore of the Balaklala mine occurs as large, flat-lying, tabular bodies of massive pyrite that contain copper and zinc minerals and small amounts of gold and silver. Two large ore bodies and many small ones are present in the mine (pls. 5 and 6). The two large ore bodies, the Windy Camp and the Weil, before being separated by the Balaklala fault, were formerly one elongate basin-shaped lens of pyritic ore. The smaller ore bodies are all massive sulfide ore, although the copper and zinc content varies considerably. The Windy Camp-Weil ore body before faulting was 1,400 feet long and slightly less than 500 feet wide. If the gossan in the vicinity of adits 1, 2, 3, and 4 is an extension of this ore zone, as is indicated by cross section *H–I–J–K*, plate 6, the ore lens was 1,800 feet long. In addition, fairly continuous mineralization at the ore zone has been demonstrated by drilling between the Balaklala and the Keystone mines, which lies 1,400 feet southwest of adit 1, and between the northwest end of the stopes on the Weil ore body and the sulfide bodies in the Horse tunnel.

The operators of the Balaklala mine recognized the basinlike shape of the prefaulted ore body, and the presence of an ore zone, as their extensive exploration by drilling appears to have taken this structure into account. The lowest point in the ore body is in the central part of stope 5 (pl. 5). From this point the base of the ore zone rises (correcting for offset on the Balaklala fault) 135 feet to the High Grade stope to the north, 154 feet to stope 16 to the northeast, 100 feet to the 153 ore body to the southeast, and 340 feet to the gossan in adit 1 to the southwest. The rise to the northwest is not as steep as in other directions. Drill holes show that the basin structure flattens to the southwest (pl. 6, *D–E–F–G*).

Character of the ore bodies.—The ore in the Balaklala mine is a hard fine-grained pyritic ore that contains chalcopyrite and sphalerite and small amounts of gold and silver. The ore as mined contained 23 percent SiO_2 (in some of the early reports, no distinction is made between silica and insoluble material), but this was in part due to remnants of unreplaced rock in the ore and to small included bands of waste; the ore had an average weight of 7.25 cubic feet to the ton.

The aggregate thickness of the sequence of rhyolitic rocks in the ore zone as here used is at least 300 feet. The top of this sequence is the base of the upper unit of the Balaklala rhyolite, and occurs throughout the district, but the base of the ore zone is not marked by any stratigraphic break and includes the lowest known ore. In areas between individual ore bodies, the pyritized zone is commonly limited to a thin layer a few feet to a few tens of feet thick immediately below the base of the upper unit. Thus in the general ore-bearing area the mineralized zone is continuous, although individual drill holes may cut through areas that contain no massive sulfide, and locally little or no pyrite. Massive sulfide ore bodies occur as sharply bounded masses in a general ore zone of weakly pyritized to well-pyritized rock.

The outlines of the stopes on the composite map of the Balaklala mine (pl. 5) commonly mark the limits of the massive sulfide ore, and beyond those limits the wall rock is either barren or weakly pyritized. There is a sharp contact along most stope walls between the massive sulfide body and the wall rock. However, a considerable amount of massive sulfide ore that was below a minable grade or was inaccessible from workings remains in place in some areas, as along the southeast edge of the Windy Camp ore body (pl. 6, *H–I–J*), and massive sulfide, observed in some drill holes, has not been mined. Also, at a few places, the end of a stope marks the point at which lenses of ore wedge down to ore that was too thin to be minable. However, according to Holman's description, the edges of most of the sulfide bodies were smooth and sharp, although curved.

One main layer in the ore zone includes the Windy Camp-Weil ore bodies and is fringed with smaller ore bodies, but a few lie above or below the principal ore layer. The North Weil 170, the High Grade stope, the Bull, and stopes 3, 8, 9, 84, 92, and 153 were isolated ore bodies, not connected to the main sulfide mass in the Windy Camp-Weil except by the weakly pyritized and locally barren ore zone. The Bull is an isolated bean-shaped ore body well below the main ore layer. The North Weil 170 lies above the Weil, and stope 92 lies below the main zone of the Windy Camp. The North Weil 3, although it is connected with the Weil, lies in part about 20 feet below the main Weil. The occurrence of ore bodies at several levels in the ore zone is shown on the cross section, but it is probable that some of these are fault segments. The ore bodies east of the Windy Camp are isolated sulfide bodies, however, as the rock between stope 84 and the Windy Camp is either slightly pyritized or barren rhyolite where it is cut by drill holes. Ore was found in two drill holes southeast of stope 5 (pl. 5), which extends the ore in this stope toward stope 9, but an area of barren ground is probably present between stopes 9 and 5.

The thickness of the massive sulfide ore in individual ore bodies ranges from a few feet to 75 feet. The Windy Camp-Weil ore body reaches a maximum thickness of 75 feet in the central part of the lens from which the ore decreases in thickness in all directions. The North Weil 170 is a lens-shaped ore body that lies about 65 feet above the top of the Weil; it pinches out to the north, south, and east, but is reported to be terminated on the west by a fault that dips 45° E. The average thickness is 25 feet, and the maximum is 40 feet. Ore body 153 averages 20 feet in thickness; no. 84, 20 feet; no. 83, 25 feet; and no. 9, 30 feet. Stope 92, lying west of the Windy Camp, averages 25 feet in thickness.

The grade of the Balaklala ore is in general uniform, but some separate ore bodies and a few masses within the larger bodies differ considerably from the general average. Table 7 gives the assay data that are known on different ore bodies. The Weil was of uniform grade throughout, including the North Weil 3, but in the main Windy Camp ore body, richer grade ore occurred in the lower 10 feet in the opencut. This higher grade ore extended 200 feet east of the opencut, and averaged 5 percent copper, 0.036 ounce of gold, and 1.0 ounce of silver. The ore lying southwest and west of the opencut averaged 1.75 percent copper. Seager reports that near the base of the ore body in the opencut, and east under the portal of tunnel 8, flat-lying bands of chalcopyrite from 2 to 12 inches thick cut the massive pyrite and replace porphyritic rhyolite. No reason is known for the localization of high copper content in ore bodies such as the High Grade (12.0 percent copper) and the Bull (10 percent copper). In each of these ore bodies the gold content increases with the copper, but the grade of the zinc remains constant, suggesting that copper and gold are closely associated at the Balaklala mine.

Ore controls.—The broad ore control for the main ore zone at the Balaklala mine is the contact between the overlying upper unit of the Balaklala rhyolite and the middle unit of the Balaklala. All ore occurs at or a short distance below this contact. More detailed ore controls appear to be the basin-shaped fold in the rocks and the thin layers of bedding-plane foliation in the flat-lying bed of pyroclastic rock and shaly tuff that immediately overlies the main ore zone.

Rounded nodules of unreplaced rock 2 to 12 inches in diameter occur in the ore in the opencut. Most of the rock remnants are composed of porphyritic rhyolite containing 2-millimeter phenocrysts, but some are fragments of nonporphyritic rhyolite. The massive sulfide ore appears to have replaced both varieties of rhyolite, but apparently did not replace the overlying pyroclastic material.

Many faults are present in the mine area, but no evidence of feeder channels was seen, except that some ore bodies stop abruptly against faults, and no faulted extensions of such ore bodies have been found.

Exploration possibilities.—Even though a large amount of exploratory drilling has been done in and around the Balaklala mine, some favorable areas remain to be explored. The most favorable zones are between the Windy Camp ore body and the Keystone mine ore bodies, which lie 1,800 feet to the southwest. Some drilling has been done in this area, and pyritized rocks were found in the ore zone. Additional prospecting should be done to the south and east of the present holes.

A large area of the ore zone lies south and east of the small lens of ore at the Horse tunnel. In this area the ore zone immediately under the base of the upper unit is covered by rhyolite of the upper unit. Drilling, based on detailed mapping of the folds and faults, might lead to the discovery of new ore bodies in this area.

The possibility of finding small isolated bodies of ore below the main Windy Camp-Weil ore body is indicated by the present drill holes. Some additional exploration below the main ore zone seems warranted if the lower mine levels are reopened.

BALAKLALA ANGLE STATION GOSSAN

The Balaklala Angle Station gossan crops out in sec. 7, T. 33 N., R. 5 W., about 1 mile east of the Balaklala mine, along the road leading from Coram to the mine. The prospect is at an altitude of 2,500 feet and is north and immediately below the Balaklala Angle Station on the old Balaklala tram line (pl. 4).

This gossan has been explored by many shafts, short adits, and drill holes, but no unoxidized sulfide has been found (fig. 48). None of the gossan has been mined. The prospect is owned by the Balaklala Consolidated Copper Co.

Geology.—The gossan lies immediately below the base of the upper unit of the Balaklala rhyolite, which in this area is composed largely of pyroclastic material. These pyroclastic rocks comprise poorly bedded tuff, much of which contains quartz crystals more than 5 millimeters across, and fine volcanic breccia that contains many lithic fragments of nonporphyritic rhyolite. The gossan is in a hydrothermally altered porphyritic rhyolite that is light greenish gray to pink or lavender, and contains 2- to 3-millimeter quartz phenocrysts. Part of the rock that contains the gossan appears to be pyroclastic material. Rocks of the upper and middle units of the Balaklala in this area are strongly pyritized and hydrothermally altered and are punky white, or light-greenish rocks, or have shades of blue-green, red, or lavender.

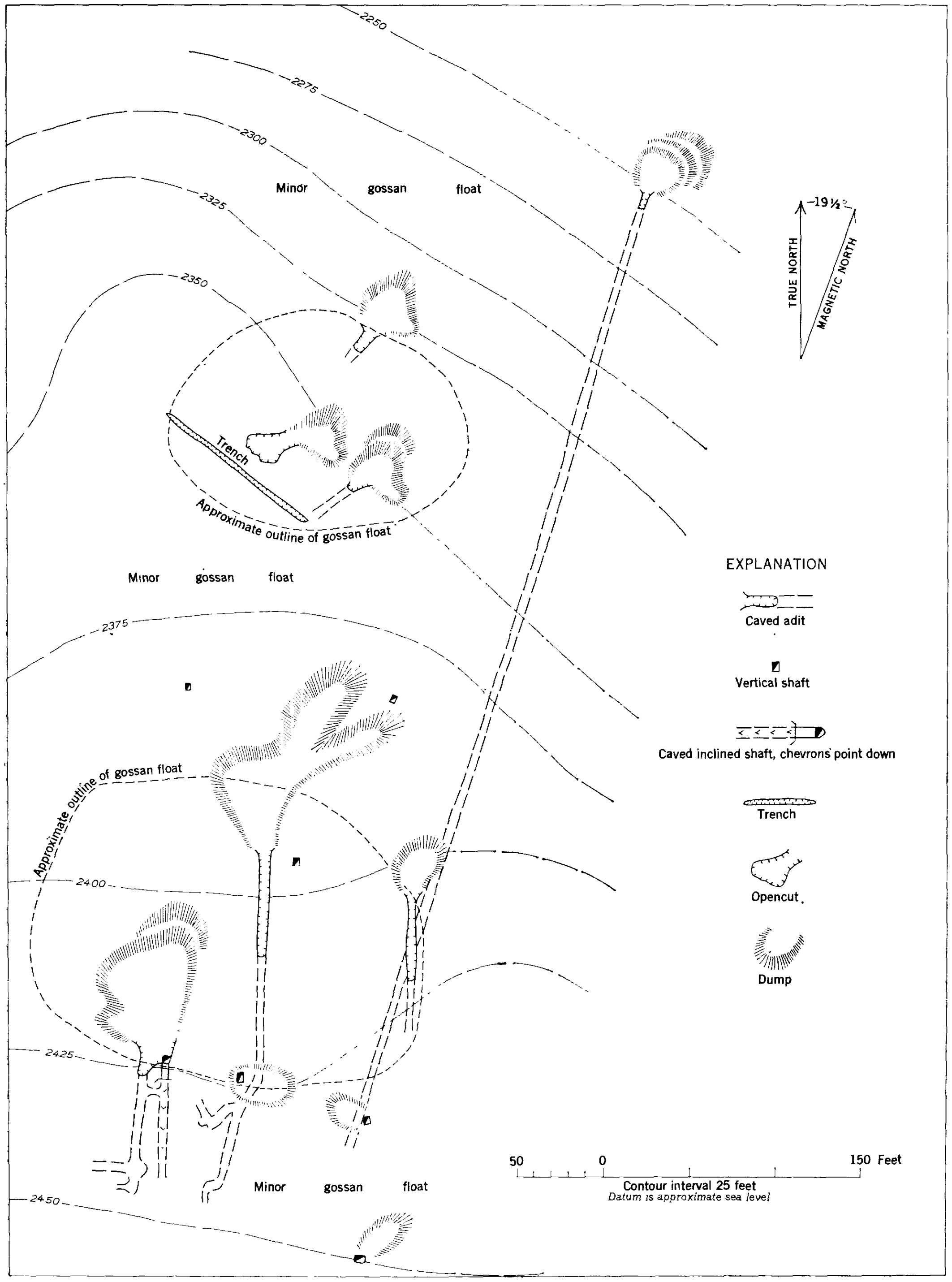
A notable feature of the Balaklala Angle Station gossan is that the gossan and enclosing rocks are not in place, but are parts of a shallow landslide. This slide, in turn, is only one of a series that are present as far as the canyon of Squaw Creek 2,500 feet to the north. The jumbled, broken character of the gossan and enclosing rocks in the slides is shown in the deep washes on the hill slopes below the gossan and in the many shallow shafts and adits in the gossan. At some places in the slide the rocks are broken and disoriented; at others the gossan and enclosing rocks are broken into rather large blocks that have moved only small amounts. One shaft 49 feet deep exposes large blocks of unbroken rock that has tuffaceous layers conformable to the general southwesterly dip, but the gossan and rock surrounding these large blocks is crushed and broken by the movement of the slide.

The large scarp behind the landslide is little eroded, although soil creep and wash have completely covered the sole of the slide. The maximum amount of movement that could have occurred is somewhat less than 200 feet vertically, and the appearance of the topography indicates that the movement was probably not much more than 100 feet.

Ore deposit.—Gossan from massive sulfide is exposed at the surface and in shallow workings, but much of the material that is intermingled with the true gossan is derived from silicified rhyolite that contained from a small percentage to 50 percent pyrite. The gossan and oxidized, pyritized rhyolite occur as boulders as much as 10 feet in diameter scattered over the surface, and none of it is considered to be in place. It is impossible to outline areas of gossan and heavily pyritized rock in more than a general way.

The gossan appears to dip about 20° SW., comparable to the apparent southwesterly dip of the enclosing pyroclastic beds, although no reliable attitudes could be obtained on the latter. The gossan is at least 15 feet thick in one adit and locally may be thicker, but it seems doubtful that a large area of continuous gossan is present. Rather, the gossan is probably in small bodies surrounded by strongly pyritized rock, although the jumbled character of the rocks in the slide may contribute to this assumption.

This deposit is of interest because of the relatively high gold content of some of the gossan, and because of the possibility of finding an unoxidized extension of the deposit to the southwest or west. The gossan has been extensively sampled, but most of the samples were taken from dumps or from gossan exposed in pits and adits. An arithmetic average of three separate sets of samples is shown in table 8.



Compiled from maps furnished by West Shasta Exploration Co. Published with permission of Balaklala Consolidated Copper Co., owner

FIGURE 48.—Map of the workings of the Balaklala Angle Station gossan.

TABLE 8.—*Arithmetic average of three separate sets of samples from the Balaklala Angle Station gossan*

[Data furnished by Coronado Copper and Zinc Co. and the United States Smelting, Refining and Mining Co.]

Sample	Gold (ounces per ton)	Silver (ounces per ton)
Group:		
1	0.118	1.48
2	.124	3.24
3	.178	2.04

Churn-drill samples indicate that an appreciable part of the gossan will average 0.25 ounce of gold per ton (R. T. Walker and W. J. Walker, written communication). All the samples show that the Balaklala Angle Station gossan carries appreciably more gold than gossan in other parts of the district. As discussed on page 97, the writers believe that the evidence indicates that this gossan was derived from a deposit of massive sulfide that contains not only pyrite but also sulfides of copper and zinc.

The area between the Balaklala Angle Station gossan and the Spread Eagle, Keystone, and Balaklala mines is almost entirely covered by the upper unit of the Balaklala rhyolite. Many warps in the base of the upper unit are shown by dips in tuff beds, but reliable attitudes on bedding are too scattered to outline the folds or warps in detail. The gentle southwest dip of the tuff beds at the Balaklala Angle Station gossan must flatten and reverse to a northeast dip to the southwest and west of the gossan, as the base of the upper unit of the Balaklala in these directions is at a higher altitude than it is at the Balaklala Angle Station gossan, although the difference in altitude may be due in part to faulting. Much of the upper unit west and southwest of the Balaklala Angle Station gossan is pyritized and hydrothermally altered to a white to lavender argillic rock. Thus almost the entire area between the Balaklala Angle Station gossan and the Balaklala, Keystone, and Spread Eagle mines is worthy of exploration because the ore zone is concealed in this area, and much of the overlying upper unit of the Balaklala rhyolite contains products of hydrothermal alteration.

CRYSTAL COPPER PROSPECT

The Crystal Copper prospect is on a small, northward-flowing tributary stream in the canyon of the South Fork of Squaw Creek about 2,000 feet north of the Early Bird mine. It is in sec. 11, T. 33 N., R. 6 W., at an altitude of 2,300 feet (pl. 4). Exploration work consists of many short adits driven into mineralized zones in rhyolite. The property is owned by the United States Smelting Refining and Mining Co.

Geology.—The prospect occurs in a flow of porphyritic rhyolite containing 2-millimeter phenocrysts, which is interlayered with minor pyroclastic rocks and flow breccia. The flow of porphyritic rhyolite is surrounded by nonporphyritic rhyolite a few hundred feet from the workings; its stratigraphic position is not known. Foliation is limited to narrow zones, and most of the rocks are massive.

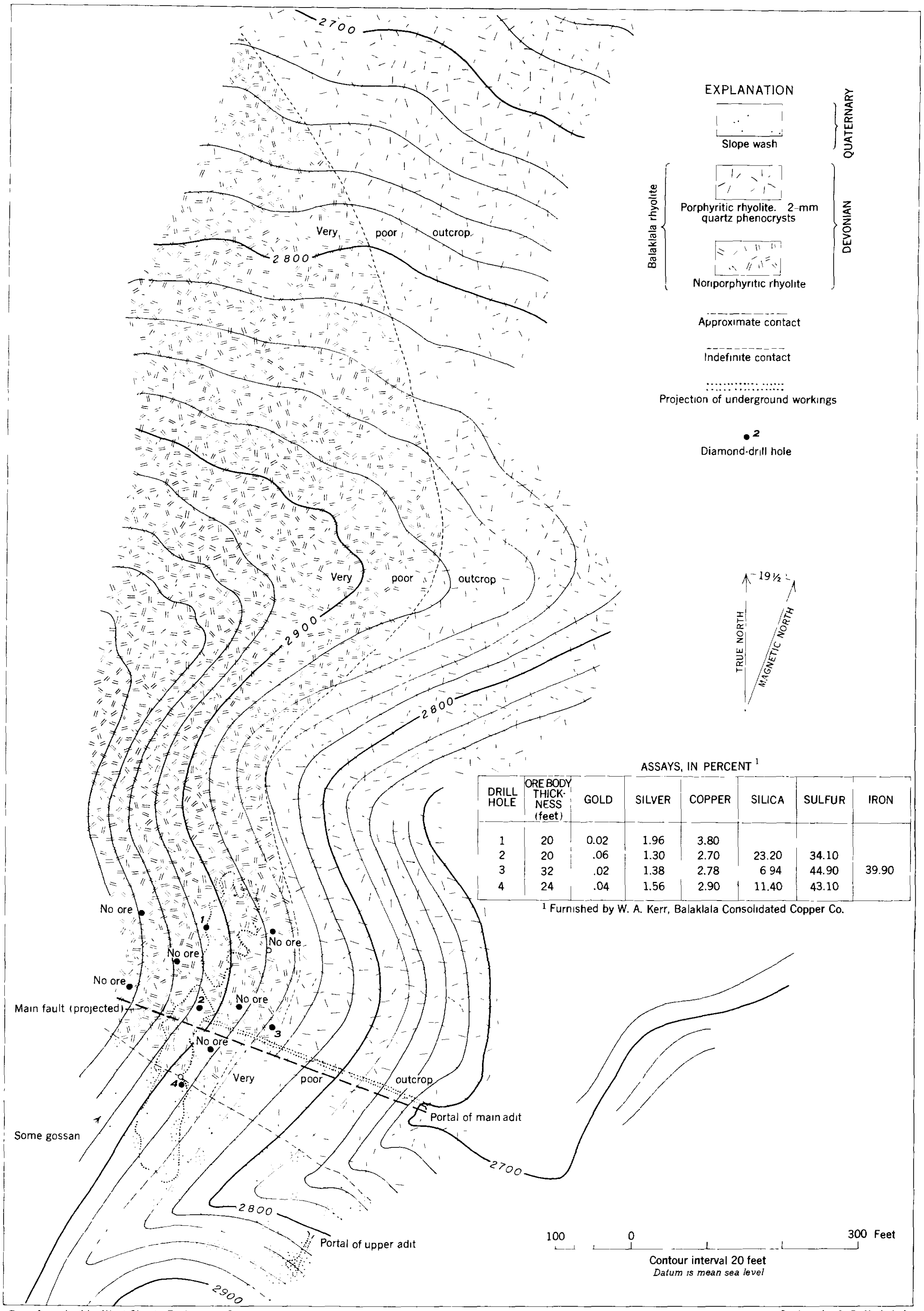
Ore deposits.—Bands of pyritized rhyolite and small lenses of gossan are prominent in the mine area. The general trend of mineralization appears to be N. 20° E. The rock on the dumps is silicified rhyolite that contains as much as 50 percent disseminated pyrite. No massive sulfide ore was seen on the dumps, but one gossan 2 feet thick was apparently derived from massive sulfide. Some of the rock in the pyritized areas has been hydrothermally altered to soft white claylike products. No ore has been mined from this prospect.

EARLY BIRD MINE

The Early Bird mine is owned by the Balaklala Consolidated Copper Co. It is located in sec. 11, T. 33 N., R. 6 W. at an altitude of 2,700 feet on the south side of the canyon of the south fork of Squaw Creek. The mine is on an isolated body of massive sulfide ore about 1 mile west of the Balaklala mine (pl. 4). A level road once extended from the Balaklala mine to the Early Bird mine. Although this road was not passable in 1951, it could be made passable with a small amount of repair. The mine was explored by a main adit 350 feet long, and by 460 feet of drifts under the ore body. The portal of the main adit was caved in 1951, but was opened for a short time in 1948. The underground workings, which were examined by the senior author and were mapped by O. W. Jarrell,[7] were open and accessible except for the first 50 feet near the portal, when the mine was reopened in 1948.

Some work was done on small outcrops of gossan near the upper adit (fig. 49) probably as early as 1908, but the ore body was discovered by drilling in 1918. Mining began in 1922, when the mine was leased to the United States Smelting Refining and Mining Co. This company mined a small tonnage of selected high-grade copper ore during 1922–25. The mine was leased to the Mason Valley Mines Co., 1926–28, during which time the major part of the stoping was done. The production from the Early Bird mine is given in table 9.

[7] Map furnished by the West Shasta Exploration Co.



ASSAYS, IN PERCENT[1]

DRILL HOLE	ORE BODY THICKNESS (feet)	GOLD	SILVER	COPPER	SILICA	SULFUR	IRON
1	20	0.02	1.96	3.80			
2	20	.06	1.30	2.70	23.20	34.10	
3	32	.02	1.38	2.78	6 94	44.90	39.90
4	24	.04	1.56	2.90	11.40	43.10	

[1] Furnished by W. A. Kerr, Balaklala Consolidated Copper Co.

Base furnished by West Shasta Exploration Co.
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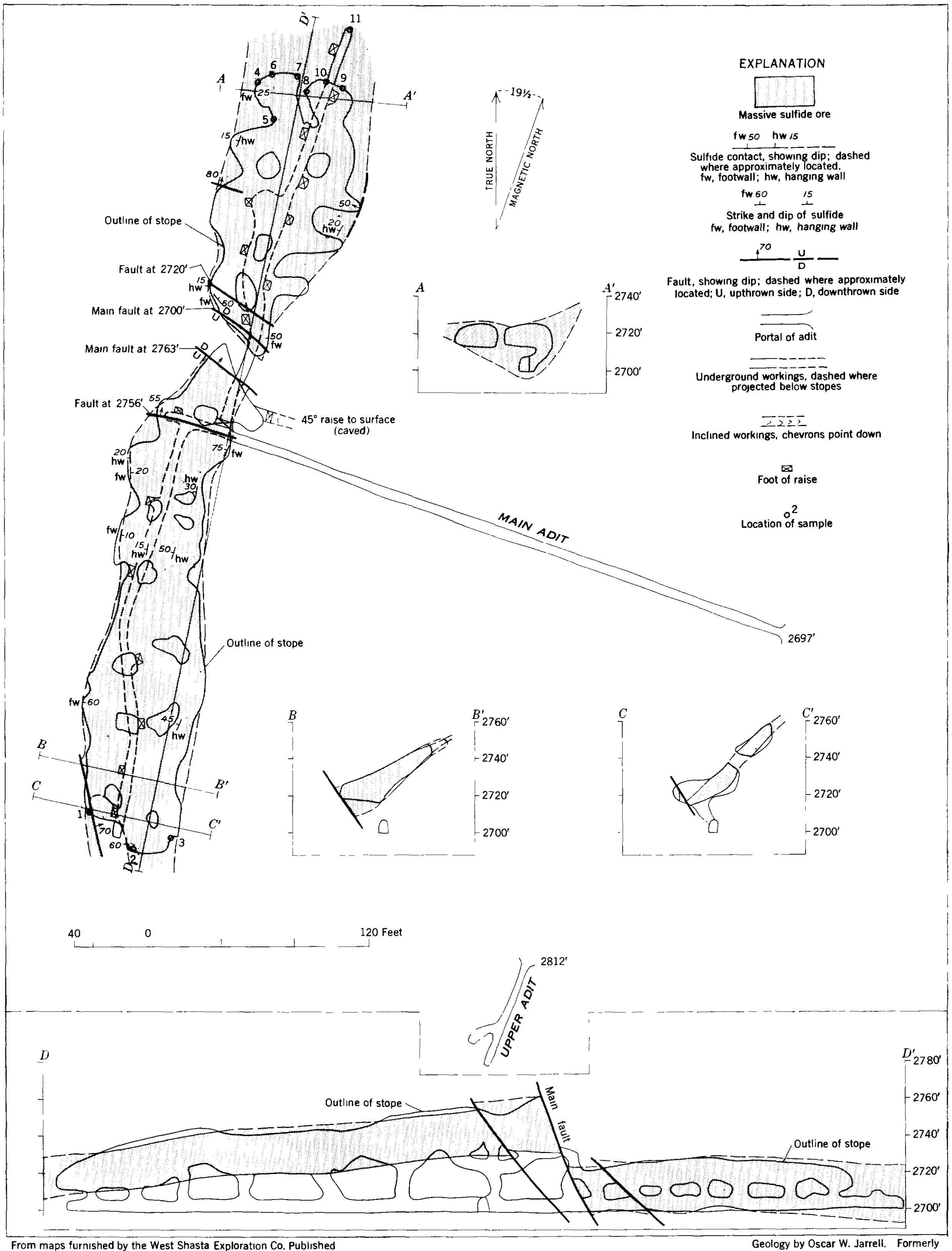
Consolidated Copper Co , owner

Geology by A. R. Kinkel, Jr.

FIGURE 49.—Geologic map of the Early Bird mine area.



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Geology by Oscar W. Jarrell. Formerly Geologist, West Shasta Exploration Co.

FIGURE 50.—Early Bird mine. Composite map of mine level and projection of stopes and ore outline at stope level.

TABLE 9.—*Production and grade of ore from the Early Bird mine*

[Data furnished by R. T. Walker and W. A. Kerr. Zinc content not given in assays]

Operator	Production (short tons)	Grade		
		Gold (ounces per ton)	Silver (ounces per ton)	Copper (percent)
United States Smelting Refining and Mining Co., 1922–25	5, 116	0. 015	0. 41	10. 27
Mason Valley Mines Co	35, 000	. 034	2. 00	3. 40

Geology.—The Early Bird ore body is in light-gray siliceous, porphyritic rhyolite that contains 2-millimeter phenocrysts, typical of the middle unit of the Balaklala rhyolite, but it is overlain by a flow of nonporphyritic rhyolite. The ore body is near the top of the Balaklala rhyolite, however, as shale of the Kennett formation crops out on the ridge about 300 feet higher than the mine. The upper unit of the Balaklala appears to wedge out a short distance east of the Early Bird mine, and the base of the Kennett formation takes the place of the base of the upper unit as a stratigraphic marker. The ore body probably occurs between 200 and 300 feet stratigraphically below shale of the Kennett formation, although the attitude of the beds is not known in detail because of the poor outcrop in the mine area.

Ore body.—The Early Bird ore body is a flat-lying elongate body of massive sulfide that is cut by a cross fault (fig. 50). The fault divides the ore body into two separate blocks; the south block is about 30 feet higher than the north. The stoped part of the ore body is 428 feet long, but the exposed strike-length of ore is 460 feet, and the faces of both the north and south stopes are in ore. The ore averages about 55 feet in width and has a maximum width of 85 feet; it averages about 15–20 feet in thickness, although the thickness is not uniform. The maximum stope height is 20 feet and the maximum thickness of ore is probably not more than 30 feet.

The north end of the ore body has a synclinal shape as shown in section *C–C′*, figure 50; the footwall contacts are exposed in the north stope, and in the north drift below the stope (fig. 51). In the south stope the ore has a uniform west dip at the south end but is cut by a fault on the west side (fig. 50, *A–A′* and *B–B′*). Dips on the footwall of the ore in the central part of the south ore body indicate that it also has a synclinal shape.

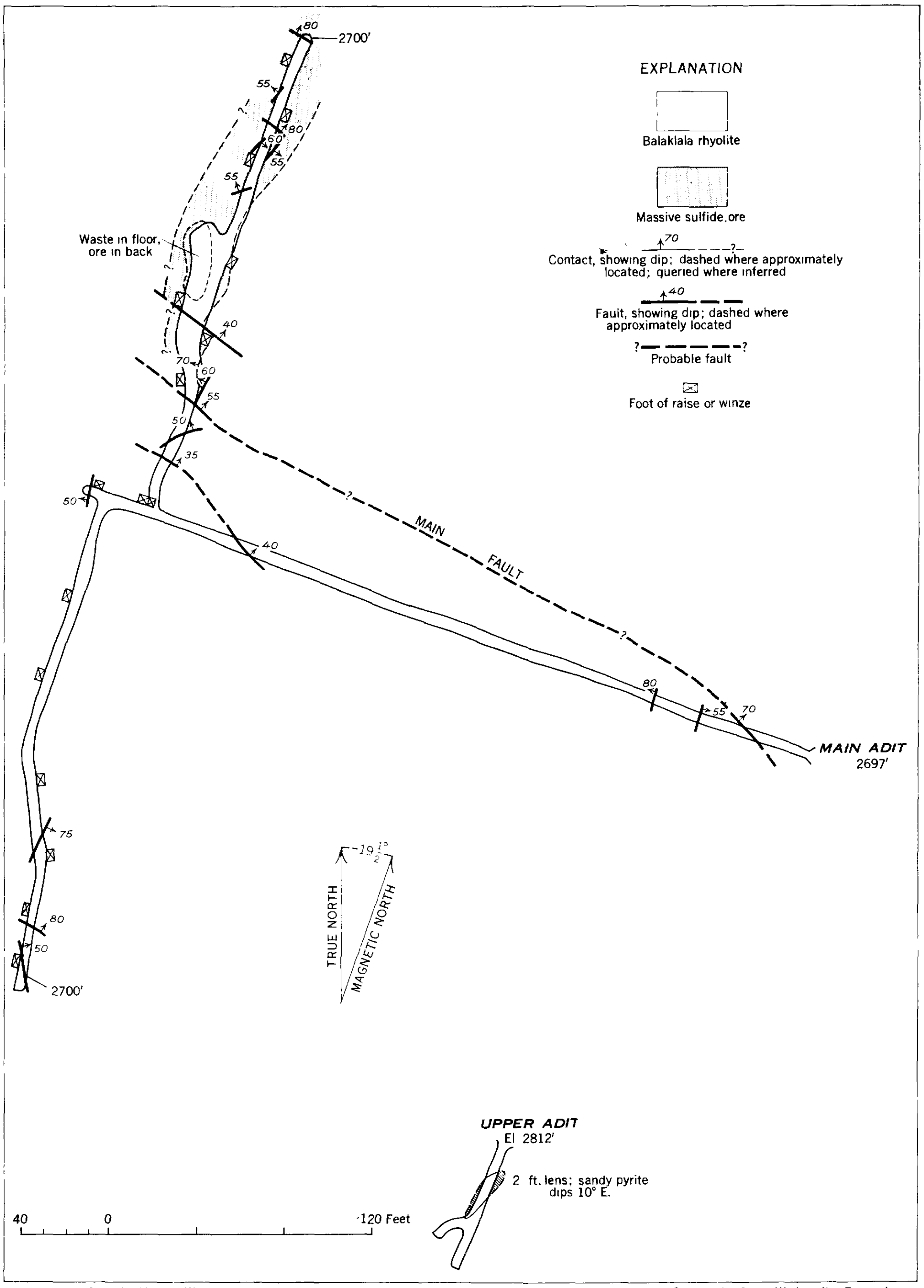
The ore is dense, metallic looking, and structureless and has no visible gangue minerals. It is cut by a few veinlets of chalcopyrite, sphalerite, and quartz. The appearance and grade of the ore is uniform.

The contacts are sharp between massive sulfide and rhyolite containing 2-millimeter quartz phenocrysts that forms the wall rock of the deposit. Foliation is not as well developed at the contact of the rhyolite and the ore as it is near the wall rock of most of the massive sulfide bodies of the area. A thin yellowish sericite or clay gouge ranging from a quarter of an inch to rarely 2 inches in thickness is present along much of the ore contact, but this might be due to minor postmineral movements. No frozen contacts were seen, and the wall rocks contain pyrite at only a few places. Most ore contacts in the Early Bird mine are regular, smooth planes that are sharply curved in places. The continuity of many contacts is broken by cross faults that have a displacement of a few feet, but no persistent direction or offset is apparent in these small faults. Crushed sulfides along minor faults and slickensided surfaces are locally present in the ore, but not along the contact between massive sulfide and wall rock.

In polished section the ore is composed largely of fractured anhedral pyrite grains as much as 1 millimeter in diameter. Chalcopyrite occurs interstitial to and in fractures in pyrite. Sphalerite is interstitial to pyrite and at some places to chalcopyrite. The gangue minerals are quartz and minor amounts of calcite.

The ore is not oxidized at its upper contact in the backs of the stopes, 150 feet below the surface. Drill holes show no enrichment in copper or silver at the top of the ore. The low silver content and absence of silver enrichment in the high-grade copper ore mined by the United States Smelting Refining and Mining Co. (table 9) also show that the high copper content of this ore was not due to enrichment. Other than this small tonnage of high-grade copper ore, the ore samples are very uniform in grade. Ore from four drill holes averaged 0.03 ounce of gold, 1.53 ounces of silver, and 3.01 percent of copper, which checks fairly closely with the record of production if the small tonnage of high-grade copper ore is omitted. No zinc assays are available.

The drill holes shown on figure 49 did not reveal the limits of the ore along its strike. If the ore continues unfaulted horizontally northward, it should crop out about 1,000 to 1,500 feet from the end of the north stope. In this stope the north face has a larger area than most cross sections of the ore body, and ore can be expected to continue for some distance north of this stope face. The fact that its outcrop has not been located at the surface may be due to the absence of good outcrops in the mine area and on the slopes to the north, or its northerly extension may be dropped by faulting. No exploration has been done to the south of the ore body, and although the face of the south stope contains less than the average cross section of ore, a large area of favorable ground remains to be explored south of the Early Bird ore body.



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Geology by Oscar W. Jarrell. Formerly Geologist, West Shasta Exploration Co.

FIGURE 51.—Map of adits of the Early Bird mine.

GOLINSKY MINE

The Golinsky mine is 1 mile northeast of the Mammoth mine on the northeast side of Little Backbone Creek, in sec. 28, T. 33 N., R. 5 W., at an altitude of 1,840 feet. Access to the Golinsky mine was cut off when Shasta reservoir was flooded. In 1950 the mine was owned by the U. S. Government. The mine is accessible only by boat to a road that leads from Shasta Lake to the mine. The road, about 2 miles long from the lake to the mine, was not passable in 1950, but it could be repaired by putting culverts in the washed-out parts. The mine workings are inaccessible, and the mine plant has been dismantled.

The outcrop of gossan was discovered before 1902, as Aubury (1902, p. 90) reports 750 feet of exploratory work had been done by that date. The mine was owned in 1902 by B. Golinsky of Kennett. During 1906–7 the mine was under lease to the American Smelting and Refining Co. This company mined 3,078 tons of sulfide ore and shipped it to the Tacoma smelter for treatment. The mine was idle from 1907 until 1931, when it was leased to Vickery Brothers, who attempted to recover the gold and silver from the gossan, but the process used was not successful. In 1935 the mine was leased to the Backbone Gold Mining Co.; this company built a road from the old town of Kennett to the mine and installed a small smelter at the mine. During their operations, 3,189 tons of ore was mined and treated in the smelter, but the mine was closed again in 1937 and has not been operated since. Table 10 gives data on production.

TABLE 10.—*Ore mined by American Smelting and Refining Co. and Backbone Gold Mining Co.* [1]

Grade of ore mined by American Smelting and Refining Co.

[Tons mined—3,078]

Gold	ounces per ton	0.134
Silver	do	4.61
Copper	percent	3.57
Zinc	do	8.9
Sulfur	do	30.7
Iron	do	23.6
Silica	do	27.5

Production of ore mined by Backbone Gold Mining Co.

[Tons smelted—3,189]

Matte produced	tons	398.5
Gold	ounces	551.54
Silver	do	11,123
Copper	pounds	91,689

[1] Data from U. S. Bureau of Reclamation, unpublished report by E. C. Galbraith and W. I. Gardner.

The gossan at the Golinsky mine was mined in a broad, shallow opencut in an attempt to recover the gold that the gossan contained. The subsequent slumping and the soil that was washed into the cut have concealed the geology near the outcrop of the ore body. Three samples taken at the surface in gossan by the U. S. Bureau of Reclamation contained 0.01 ounce of gold and 2.28 ounces of silver. Samples taken by the United States Smelting Refining and Mining Co. on the gossan exposed in the underground workings assayed much higher than the U. S. Bureau of Reclamation assays at the surface. The weighted average of 13 samples taken on gossan underground showed that the gossan contained 0.36 ounce of gold, 8.10 ounces of silver, 1.20 percent copper, trace of zinc, 24.0 percent iron, 6.4 percent sulfur, and 47.4 percent silica.[8]

Geology.—The massive sulfide ore body at the Golinsky mine lies just above a gently dipping contact between siliceous white foliated, nonporphyritic rhyolite below, and massive tan soft, porphyritic rhyolite above. The nonporphyritic rhyolite below the ore body is a heterogeneous mixture of flow-banded rhyolite and rhyolitic pyroclastic material. Several types of pyroclastic rocks are present, but none of it shows bedding. Coarse and fine volcanic breccias composed of rounded knobs of rhyolite in a matrix of rhyolite, probably a flow breccia, are mixed with stubby layers of nonbrecciated rhyolite. One bed of volcanic breccia in Little Backbone Creek below the mine is composed of a felsitic, green chloritic matrix that contains 3-inch oval bombs of white rhyolite containing closely packed chlorite and quartz amygdules. Much of the rhyolite below the mine has a steeply dipping foliation, and sericite or clay minerals are formed along closely spaced planes. Locally the rock is pyritized.

The nonporphyritic rhyolite and volcanic breccia grades upward into a tuffaceous bed that is commonly present along the lower contact of the overlying porphyritic rhyolite in this area. The tuffaceous bed is not found at the mine and may be thin or absent, but is well exposed several thousand feet north and south of the mine. This bed is a well-bedded, water-deposited crystal tuff, that contains material ranging from grit to fine volcanic conglomerate. It dips gently to the west, somewhat flatter than the hill slope.

The tuff bed is overlain by medium-phenocryst porphyritic rhyolite about 100–200 feet thick, but part of the hillside appears to be almost a dip slope, and the thickness is difficult to estimate due to the absence of bedding. Most of the porphyritic rhyolite contains 2- to 3-millimeter quartz phenocrysts, and is interlayered with pyroclastic rocks and thin layers of rhyolite or rhyolitic tuff, which is the typical lithology of the middle unit of the Balaklala rhyolite. However, some bodies of coarse-phenocryst rhyolite characteristic of the rock in the upper unit of the Balaklala are present

[8] Data furnished by United States Smelting Refining and Mining Co.

and these probably are intrusive masses, but some may be local bodies of extrusive coarse-phenocryst rhyolite that here antedate the main period of extrusion of this rock and occur as small flows in the middle unit.

The base of the upper unit of the Balaklala rhyolite is marked by a layer of locally well-bedded crystal tuff that lies beneath shale of the Kennett formation on the ridge above the mine, but it is not certain whether the coarse-phenocryst rhyolite that lies a short distance above the Golinsky ore body is an extrusive mass and therefore part of the upper unit of the Balaklala rhyolite or whether it is an intrusion into the middle unit.

Ore body.—The Golinsky ore body is a relatively small steeply dipping lens of massive sulfide. It is explored by two adits and by several intermediate levels (pl. 7). The massive sulfide has sharp boundaries along the sides of the lens, and in the northeast end of the ore body on the 55-foot level the ore is basin shaped or keel shaped.[9] At the west end of the ore on the 55-foot level, the lens of massive sulfide pinches out along a clay seam in a shear zone, as it does at the east end of the level of tunnel 1. At the east end of the 70-foot level and along the southeast edge of the ore on the 55-foot level, the contact is gradational to pyritized rock, or to rock containing scattered bodies of massive sulfide. Likewise, the ore pinches out in the winze below the 55-foot level at the west end of the ore body (pl. 7, *C–C′*).

The ore is shown on the underground map as localized in a shear zone that strikes N. 60° E. and dips 55°–60° SE.; the shear zone was not seen on the surface.

The following information on the appearance of the ore is taken from a report furnished by the United States Smelting Refining and Mining Co.:

At about the centre of the workings, the fractured zone widens and makes a shoot of ore that has maximum dimensions of about 30 feet wide, 65 feet long, and 60 feet high as stoped. This zone narrows to the east and west of the main lens. The hanging wall is generally pretty well defined, with porphyry and some bunches of sulphides south of it. The ore is not quite so sharply limited on the north, and sometimes ore makes out irregularly into the footwall. A coarse, firmly cemented breccia-zone often occurs on the footwall. Cross-cutting has exposed no ore to the north or south of the main fracture-zone. Above the sulphide orebody, is oxidized material and gossan, extending to the surface.

Generally speaking, the main orebody shows in plan as a lens-shaped mass, thick in the middle, thin on the edges, occupying a widened bend in the fractures. In vertical cross-section, the main orebody would also show as a lens, grading into gossan above and thus losing its upper outline, and grading into waste and tailing out to a thin edge shortly below the 55-ft. level.

The sulphide ore is a mixture of compact sulphides, with bands of light and dark clay, soft sulphides, and quartz, mixed with some bands of porphyry. The banding is parallel to the walls, and some bands of clean sulphide are as much as 6 ft. wide. An old cross-section through the main orebody, made in 1903, shows a large horse of waste occurring as an inverted wedge in the centre of the main sulphide body.

Seager in his unpublished report noted that much sooty chalcocite was present locally and that all the ore appears to be somewhat enriched in copper.

The stope outlines are not sufficiently known to be shown on plate 7. The main stope probably extended from the 70-foot level to the level of tunnel 1, and a small stope extends 45 feet up along the raise at the east end of this level. However, old maps from different sources are not in agreement on the extent of stoping.

No exploratory work was done along the strike of the ore lens except on the 84-foot level. The ore lens at each end pinches down to a narrow band of sulfide or pinches out completely, and there was little evidence to encourage exploration. The steep topography above the mine and the lack of pyritization or hydrothermal alteration discouraged surface exploration. In spite of these features, the present study of the geologic setting of the deposit suggests that exploration to the northeast along the trend of the shear zone is warranted because mineralized rocks are exposed northwest of the mine, and because the main ore zone is concealed along the extension of the shear zone.

The steep ore lens as far as known appears to be localized along a steeply dipping shear zone. However, as the rocks above the mine contain local pyroclastic beds that dip gently southwest parallel to the overlying shale contact, flat structures similar to those known to have localized ore at other mines are present. The southwesterly dip of these structures make surface drilling to the ore zone feasible, despite the steep topography above the mine. Any flat-lying ore zone would lie within reach of drilling even near the crest of the ridge. A broad warp or arch structure that might localize mineralization occurs northeast of the mine. The ridge east of the mine that is capped by limestone of the Kennett formation marks the crest of a broad warp trending north or northwest. The southwest flank of the warp is eroded, but it apparently had a more gentle dip than the northeast flank; this warped structure is of the type that is favorable for ore deposition elsewhere in the district.

Pyritized rock is present northwest and northeast of the Golinsky mine. Gossan occurs at the base of the upper unit of the Balaklala rhyolite northwest of the Golinsky mine across the North Fork of Little Backbone Creek, at what is considered to be the ore zone throughout the district. Also, a minor amount of

[9] The information on the underground geology of the mine is derived largely from a private report furnished by the United States Smelting Refining and Mining Co., but it is in part from a report furnished by the U. S. Bureau of Reclamation, from an unpublished report by G. F. Seager, and from other sources.

gossan occurs in the lower unit of the Balaklala northeast of the Golinsky mine near Backbone Creek, but the favorable middle unit is eroded in this area. The Golinsky ore body is probably in the lower part of the middle unit of the Balaklala, and the rock stratigraphically above the mine and below the horizon of the upper unit, marked by the crystal tuff beds along the base of the shale of the Kennett formation, is favorable for prospecting.

The Mammoth and the Golinsky mines are both associated with steeply dipping faults or shear zones. Although these zones are not definitely known to be feeder channels, the California fracture at the Mammoth mine is about in line with the shear zone at the Golinsky mine, and there are pyritized rocks in the lower unit of the Balaklala along the trend of the shear zone between the two mines, and extending northeastward from the Golinsky mine. It can be assumed that there is a trend of mineralization from the Mammoth mine through the Golinsky mine and on to the northeast. The favorable zone is eroded in Little Backbone and Backbone Creeks, but it is present under the crest of the warps, which trend northwestward, northward, and northeastward from the Golinsky mine.

GREAT VERDE PROSPECT

The Great Verde prospect, formerly known as the Vulcan group of claims, is in sec. 11, T. 33 N., R. 6 W., on the north side of Squaw Creek immediately west of the Shasta King mine. The former Vulcan camp which was on the ridge about 1,500 feet east of the upper workings, is completely dilapidated, and all trails to the prospect have been washed out and overgrown. Considerable exploratory work was done, beginning about 1900, but no ore was mined from the property. Short adits were driven in a prominent gossan outcrop near the top of the ridge at an altitude of about 2,600 feet, and a 1,500-foot exploration adit was driven north from Squaw Creek at an altitude of about 2,000 feet to prospect under the gossan. A small amount of drilling was done near the upper workings, but the writers could find no record of these holes. The claims are owned by W. A. Kerr.

Geology.—The gossan at the upper workings, at an altitude of 2,600 feet, lies immediately below the base of the upper unit of the Balaklala rhyolite. The upper unit ("cap rock") is here composed of crytal tuff that contains quartz phenocrysts more than 5 millimeters across; it is underlain by buff-colored shaly tuff. Only the lowest part of the upper unit is exposed where it caps the ridge; the remainder has been removed by erosion. The gossan occurs in porphyritic rhyolite of the middle unit of the Balaklala, which is thin at the Great Verde, as it is at the Early Bird mine to the south, and is composed largely of pyroclastic material. Poor outcrop, strong hydrothermal alteration, and mixtures of porphyritic and nonporphyritic rhyolite to the north of the Great Verde make it impossible to outline the lower boundaries of the middle unit accurately in this area. Also, a varietal type of rhyolite below the Great Verde, shown on the quadrangle map (pl. 1) as belonging in the lower unit of the Balaklala rhyolite, locally contains 1-millimeter quartz phenocrysts, as at the Shasta King mine. No sharp division can be made between the middle and lower units in the vicinity of the Shasta King, Great Verde, and Early Bird mines, but at all these localities, the middle unit appears to be relatively thin.

Ore deposit.—A small amount of gossan derived from massive sulfide crops out at the upper workings and is exposed in several short adits. Some relict massive pyrite in gossan was seen on the dumps, but no information was available on the grade of the gossan or sulfides. Much of the material in the outcrop is gossan derived from heavily pyritized rock rather than gossan from a massive sulfide body.

A poorly exposed zone of strongly pyritized rhyolite, some of which contains 1-millimeter quartz phenocrysts, occurs west and southwest of the upper workings of the Great Verde and about 200 feet vertically below the upper adit. Float boulders of gossan from massive sulfide were found on this hill slope, but they may have come either from the ore zone at the upper workings or from the lower mineralized zone. There is no indication that sulfide was found in the 1,500-foot adit that was driven north from the level of Squaw Creek; the adit probably is too low in the stratigraphic sequence and is below the ore zone.

Some of the operators in the district believed that the Great Verde was a faulted segment of the Shasta King ore body as noted by Seager in his unpublished report. A strong fault occurs at the west edge of the Shasta King ore body, and the rocks west of this fault have moved up relative to the east side. However, the two deposits are believed to be discrete ore bodies because they are not in the same stratigraphic position in the middle unit of the Balaklala rhyolite, and because the major trend of the Shasta King is northeast-southwest rather than east-west (Kinkel and Hall, 1951, p. 7).

The possibility of locating ore bodies at the Great Verde depends upon the stratigraphic correlation between the rocks of the Shasta King mine and those at the Great Verde. The ore body at the Shasta King mine cannot be much less than 350 feet below the contact between the upper and middle units of the Balaklala, that is, it is on an horizon that is stratigraphically low in the

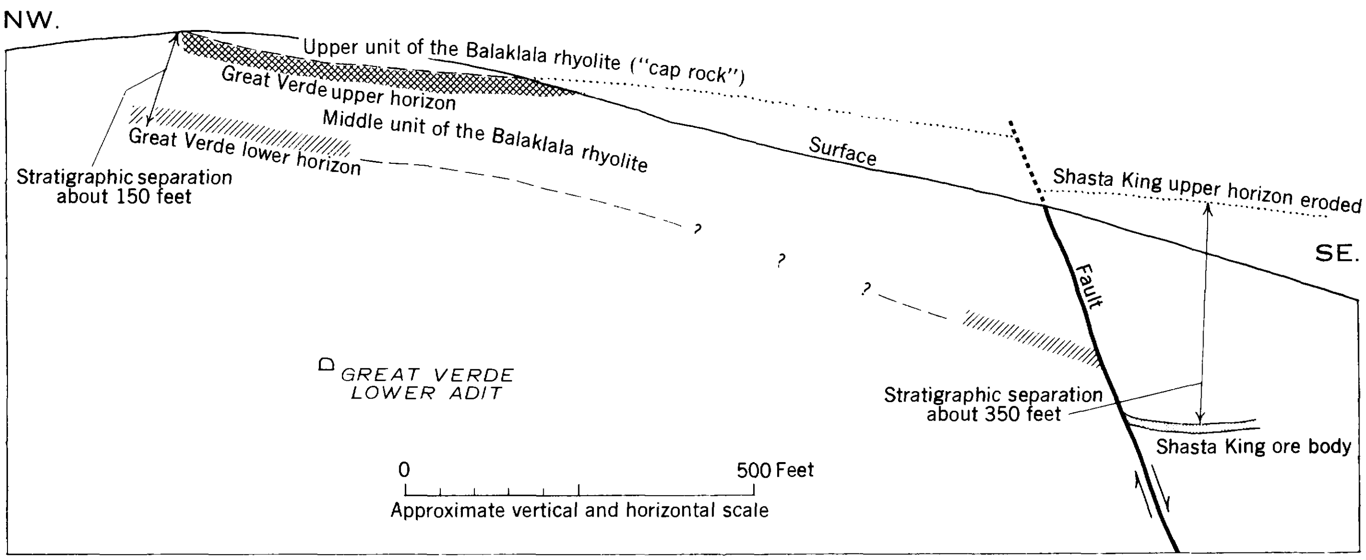


FIGURE 52.—Diagrammatic section showing the relationship between the Shasta King and the Great Verde ore zones.

middle unit. The upper ore zone at the base of the upper unit is eroded at the Shasta King mine, but hydrothermal alteration and mineralization occur at this stratigraphic horizon north of the Shasta King. Thus the lower, or Shasta King ore zone, could be correlated with the lower ore zone west and southwest of the Great Verde if, as seems probable, the middle unit of the Balaklala thins to the west (fig. 52). The Great Verde gossan at the upper workings could not be a faulted extension of the Shasta King ore body because it lies at a different stratigraphic horizon in the Balaklala. The upper gossan at the Great Verde probably can be correlated with the mineralization under the base of the upper unit north of the Shasta King mine.

On the above hypothesis, a part of the area north of the Great Verde might warrant prospecting for the Shasta King ore zone, although it is probable that lower zones do not have as much continuity as the main ore zone.

IRON MOUNTAIN MINE

The Iron Mountain mine, which is owned and operated by the Mountain Copper Co., Ltd., is the southernmost mine in the West Shasta copper-zinc district. The mine is 17 miles by road northwest of Redding and lies at an altitude of 2,600 feet; a hard-surfaced county road connects the Iron Mountain mine with U. S. Highway 299W. The Southern Pacific Railroad passes through Redding, and the ore from the mine is carried to a spur of the railroad by an aerial tramway.

The information on the Iron Mountain mine is summarized from a more detailed report by Kinkel and Albers (1951).

HISTORY AND PRODUCTION

The first claims on the large gossan outcrops on Iron Mountain were staked in the early 1860's and held for the future value of the gossan as iron ore. Silver ore was discovered in the gossan in 1879 and some exploratory work and mining were done in the silver-rich parts. At that time little interest was shown in the disseminated chalcopyrite and the massive sulfide ores that were found in the search for precious metals. It was not until 1895, when a thorough prospecting of Iron Mountain disclosed large bodies of copper-bearing sulfides, that the mineral possibilities of the region now known as the West Shasta copper-zinc district were recognized.

Silver ores were mined intermittently in the gossan at Iron Mountain from 1879 to 1897, when the present owners, the Mountain Copper Co., Ltd. (formerly Mountain Mines, Ltd.) began mining the massive sulfide ores for their copper content. This company has operated the mine continuously since 1897. The exploration of separate bodies of sulfide ore led to the naming of individual ore bodies as different mines, although they were mined as part of one operation by the Mountain Copper Co., Ltd. Thus the Old Mine ore body, the No. 8 mine, and the Hornet mine are separate and were worked at different times, but all are part of the Iron Mountain mine (pl. 8).

Copper has been produced by the Mountain Copper Co., Ltd., from direct-smelting ore, from sulfide ore treated in a flotation plant, and from the leaching of pyritic ore that was mined for its sulfur content. The Iron Mountain mine produced 197,951,738 pounds of copper by the end of 1919 from direct-smelting ore, but

TABLE 11.—*Production and grade of ore from the Iron Mountain mine*

[Data furnished by the Mountain Copper Co., Ltd.]

Ore and source	Production (short tons)	Grade						
		Gold (ounces per ton)	Silver (ounces per ton)	Copper (percent)	Zinc (percent)	Iron (percent)	Sulfur (percent)	Insoluble (percent)
Copper-zinc (flotation plant)	380,000	0.02	1.00	2.00	3.50	40.5	47.0	6.5
Copper (Old Mine ore body)	1,608,000	.04	1.00	7.50	[1] 2-5			
Copper (disseminated)	820,000	.001	.04	3.50				
Pyrite for sulfur	3,600,000	.02	.80	.40		43.0	50.0	3.5
Gossan	2,638,000	.073	8.21					

[1] Estimated.

figures are not available for the total copper production since that date as copper production was reported only by counties. After 1919, the principal periods of copper production from ore of the Iron Mountain mine were in 1925, 1928–30, and 1943–47. Minor copper production was maintained between these periods by leaching of ore that was mined for sulfur.

Gold and silver have been extracted from the gossan overlying the massive sulfide ore of the Old Mine ore body. From 1889 to 1893, 38,000 tons of gossan that contained 8 ounces of silver to the ton was mined. The gold content of this ore is not known. From 1929 to 1942, 2,600,000 tons of gossan that contained 8.3 ounces of silver and 0.073 ounce of gold per ton was mined.

The tonnage and grade of ore are summarized in table 11. The copper-zinc ore that was treated in the flotation plant was mined from the Richmond and Mattie ore bodies. The disseminated copper ore came from the No. 8 mine and the Confidence-Complex vein system. Most of the pyrite mined for sulfur came from the Hornet ore body, although some came from the Richmond-Complex ore body.

FORMATIONS IN THE MINE AREA

The Copley greenstone, the Balaklala rhyolite, and the albite granite are the only rock units that occur in the immediate vicinity of the Iron Mountain mine. These are shown on the surface map and on the cross sections (pl. 19).

The Copley greenstone in the mine area is composed of chloritized mafic flows and minor pyroclastic material; the greenstone ranges from well-sheared to massive rocks and is pyritized only in small, local areas. It underlies the Balaklala rhyolite.

The Balaklala rhyolite in the mine area probably includes the upper, middle, and lower units, although shearing, folding, original lenticularity of flows and pyroclastic rocks, and an abnormal amount of intrusive rhyolite make the stratigraphy within the Balaklala unusually difficult to decipher at Iron Mountain. It is particularly difficult at this mine to distinguish between the middle and the lower units, and the two units may not be separable as distinct stratigraphic units at this locality. Also, most of the typical coarse-phenocryst rhyolite of the upper unit has been eroded in the mine area; only the transition zone of tuff and coarse-phenocryst pyroclastic material that is characteristic of the material along the contact between the upper and the middle units remains. The Iron Mountain mine was mapped early in the present study, and subsequent work in other parts of the district proved to the writers that the rock mapped as coarse-phenocryst rhyolite at the Iron Mountain mine (Dbx_3 on pl. 9) is not typical of the rhyolite of the upper unit elsewhere in the district, but is a coarser phenocryst facies of the middle unit, and is limited to the vicinity of the Iron Mountain mine. The fourfold lithologic division of the rhyolite based on phenocryst sizes, which was used in the mapping of the Iron Mountain mine in the earlier stages of the work and which could be maintained in the mine area, was later revised to a threefold division for the district as a whole. The small and medium-phenocryst rhyolites of the earlier work at the Iron Mountain mine were grouped together under medium-phenocryst rhyolite in the later work in the West Shasta district because these subdivisions could not be maintained elsewhere.

ORE DEPOSITS

CHARACTER

The two types of ore in the Iron Mountain mine are massive pyrite bodies, that contain chalcopyrite and sphalerite, and zones of disseminated chalcopyrite and quartz-chalcopyrite veins in schistose rock. The massive sulfide is much more abundant than the disseminated ore. Disseminated ore occurs only in the No. 8 mine and the adjoining Confidence-Complex ore bodies. All other ore bodies in the Iron Mountain mine are massive pyritic bodies. The principal minerals are pyrite, chalcopyrite, and sphalerite. Galena, pyrrhotite, tetrahedrite, and magnetite have been seen in a few specimens, and the ore contains recoverable amounts of gold and silver. The only gangue minerals seen are very small amounts of quartz and calcite, both of which occur as interstitial grains to the sulfide minerals and as veinlets cutting the ore.

In addition to the sulfide ores at the Iron Mountain mine, three small, lenticular bodies of magnetite occur about 700 feet south of the Iron Mountain quarry (pl. 9). These are isolated from the main body of pyritized rock and massive sulfide by several hundred feet of unmineralized rock, and contain little or no pyrite and no copper or zinc minerals. Drilling has shown that the base of the lenses is at shallow depths. Part of the lenses are solid magnetite, some of which is polar but they contain some hematite, which is in part a surface alteration product. At the edges of the lenses, the magnetite is disseminated in rhyolite. Maghemite, ferromagnetic ferric oxide, is reported in gossan at Iron Mountain by Sosman and Posnjak (1925, p. 332–333). On the supposition that the maghemite might have been collected from the surface of the magnetite bodies, a sample of the magnetite that contained some brown, rusty material was sent to Charles Milton of the U. S. Geological Survey for study and analysis.

The analysis of the sample of magnetite is given below:

Analysis of magnetite from Iron Mountain mine

[Analysts, Michael Grasso and Leonard Shapiro]

	Percent
Total Fe as Fe_2O_3	92.27
FeO	20.63
Actual Fe_2O_3[1]	69.34
TiO_2	.24

[1] By calculation.

Milton (written communication) found no maghemite to be present and reported on the sample as follows:

Calculations of an analysis [by Grasso and Shapiro] together with optical data indicate that the composition of the ore is about: 70 percent magnetite, 10 percent hematite, 20 percent limonite and gangue. There is no reason, from any data at hand, to believe that maghemite is present in this sample.

However, in other samples, collected from gossan from massive sulfide on Brick Flat (east of the reservoir shown on pl. 9), Milton found maghemite, and reported:

One thin and three polished sections were studied, as follows:

692. Very slightly magnetic. Microscopically shows structure as shown in [fig. 53]; the bright areas are composites mainly hematite, with probably maghemite. The outer zone of the network appears to be hematite (anisotropic bright steel gray) and the interior part maghemite (isotropic, darker gray).

A thin section shows angular (fractured) quartz and opaque iron oxide.

693. Similar to following, more or less.

694. This appears to be somewhat more dense than the other two specimens, and is also darker, as well as much more magnetic. In polished section it shows the same micro-botryoidal texture, apparently consisting essentially of hematite, maghemite, and limonite. A few specks of pyrite are present, and also some pinkish mineral with the reflectivity of magnetite, which it strongly resembles except for being translucent (giving strong internal reflections). It is, therefore, probably rutile.

An X-ray study of the magnetic powder by Fred A. Hildebrand (IWX–314) gave the following:

Hematite and moderate amounts of maghemite and quartz. The maghemite has a unit cell slightly larger than the cells of other maghemite patterns in the X-ray file.

The micro-structure of the iron oxide minerals strongly suggests crystallization from a gel.

No separate bodies of magnetite are known in the area where the samples that contain maghemite were collected, nor is any visible in the ore in the underground workings. Magnetometer readings in the area of Brick Flat showed a considerable range, but whether this is due to maghemite or magnetite in the ore is not known. The maghemite was probably derived from lepidocrocite, as suggested by Sosman and Posnjak (1925, p. 329–342), and not by oxidation of magnetite.

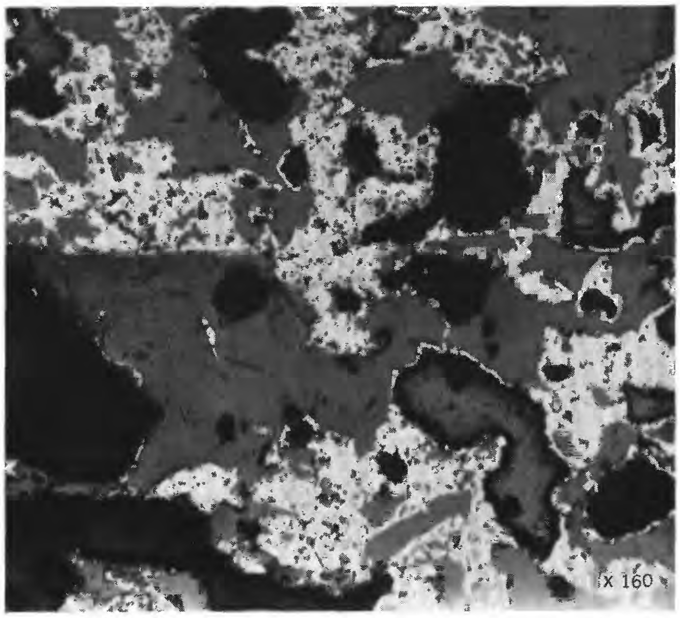


FIGURE 53.—Photomicrograph of polished section of gossan that contains maghemite. Iron Mountain mine, × 160.

DISTRIBUTION OF MINERALS

The mineral content of the massive sulfide and of the disseminated ores is essentially the same but the proportions of minerals differ considerably between the two types. The disseminated ore contains predominantly pyrite and chalcopyrite in sericitic and siliceous rocks, and contains many chalcopyrite-bearing quartz veins. Pyrite and chalcopyrite generally are present in about equal amounts. Only minor amounts of sphalerite, gold or silver occur in the disseminated ore. The massive sulfide ore, on the other hand, is composed almost entirely of pyrite, but contains chalcopyrite and sphalerite in small amounts distributed throughout the massive sulfide bodies, and local concentrations of these minerals occur. Gold and silver are present in recoverable amounts only in the massive sulfide ore bodies.

The detailed distribution of chalcopyrite and sphalerite in the massive sulfide ore is not well known. However, as all the larger concentrations of these minerals were mined as base-metal ore, the location of stopes that were mined for the copper and zinc content shows the location of these concentrations. The record is incomplete, as small bodies of base-metal ore have been found at some localities in the mine where such ore could not be mined separately from the massive pyrite.

The copper-zinc ore is found principally along the edges and bottoms of thick massive sulfide bodies that as a whole contain little chalcopyrite or sphalerite, but it is not everywhere present along such boundaries. In other parts of the mine, chalcopyrite and sphalerite occur in minable quantities throughout massive sulfide ore bodies, but these localities are generally in the thinner parts of the sulfide mass.

DISSEMINATED COPPER ORE

A body of disseminated chalcopyrite and pyrite underlies the Old Mine ore body. Where the disseminated ore lies beneath the Old Mine ore body, it was mined through the workings of the No. 8 mine, but its extension to the northeast of the Old Mine ore body is known as the Confidence-Complex vein system. The location of this disseminated and vein-type copper ore is shown in plates 8 and 11.

Two principal types of ore, which occur together or separately, are present in the No. 8 mine. One type consists of chalcopyrite grains, veinlets, and fairly solid masses of coalesced chalcopyrite veinlets that replaced schistose porphyritic rhyolite. Pyrite is subordinate in amount to chalcopyrite and occurs as scattered anhedral grains. Many small and discontinuous faults and gouge zones are present, and the largest ore bodies occur at intersections of these gouge or fracture zones. The second ore type, quartz-chalcopyrite veins, is less abundant but occurs locally in or near the borders of the disseminated ore. The quartz-chalcopyrite veins of the Confidence-Complex workings occur as fracture fillings. In the southwest end of the Confidence-Complex workings the ore zone contains disseminated chalcopyrite and quartz-chalcopyrite veins. The northeast end of the Confidence-Complex workings contains principally quartz-chalcopyrite veins along a fault that has formed several inches to several feet of gouge.

The shapes of the No. 8 mine ore bodies are shown in figures 54 and 55. The ore in the No. 8 mine occurred along shear zones, particularly along intersecting shear zones or intersecting minor faults. The ore bodies are reported by the mine staff to parallel the schistosity of the replaced rock. Plate 11, which was compiled from stope maps, shows the shapes of the mined ore bodies. The material between ore bodies was, in places, mineralized rock that contained too little copper to be mined. Consequently, the map shows the major ore shoots but not the extent of mineralization.

There are two main ore bodies in the No. 8 mine, and each is arcuate in horizontal section. This curvature of the ore bodies is best shown on the 2,350-foot level in the east ore shoot and on the 2,500- and 2,610-foot levels in the west ore shoot (fig. 54). Sections *A–A′* and *B–B′* (fig. 55) illustrate the echelon pattern of individual ore shoots and indicate that the thickest parts of most of the ore bodies correspond to marked changes in dip.

STRUCTURAL FEATURES OF THE ORE BODIES

The massive sulfide deposits, which make up the bulk of the ore of the Iron Mountain mine, are enormous masses composed almost entirely of pyrite. Except in the vicinity of the Old Mine ore body, the wall rocks are virtually unmineralized. The massive sulfide ore bodies differ in shape and attitude (pl. 10). The Hornet is nearly vertical. The Mattie is cigar shaped and horizontal; its faulted extension has not been located and may have been removed by erosion. The rounded base of the erosion remnant of the Old Mine ore body suggests that before erosion it was a large, gently dipping, lens-shaped or synclinal mass. The Richmond and Complex, taken together, have a synclinal shape, and the Brick Flat also may be in part synclinal, although its shape is determined only by rather widely spaced drill holes. The No. 8 mine and the Confidence-Complex ore bodies (pl. 8) are in zones of chalcopyrite-bearing, sericitic, porphyritic rhyolite, and along quartz-chalcopyrite veins or minor faults.

The Hornet, Richmond, Complex, and Brick Flat massive sulfide ore bodies were one continuous body before they were displaced by the Scott and Camden faults. It also seems possible that the New Camden ore body is a faulted segment of the Complex, but the relationship between these two is not well known. The longitudinal section (pl. 11) suggests that the ore in the gossan area, which occurs updip from the Old Mine ore body and the No. 8 mine, is a faulted part of the Brick Flat. It is probable that all the major ore bodies at the Iron Mountain mine were one continuous deposit before postmineral faulting, but small, isolated ore lenses also occur, such as the Mattie and Okosh, which lie along the side of the main ore body.

The massive sulfide ore that is exposed in the upper Busy Bee adit (pls. 8, 9, and 11) has been explored only to a limited extent. The lower Busy Bee adit is barren but a crosscut 130 feet in from the portal of the upper adit exposed massive sulfide that is 45 feet in width,

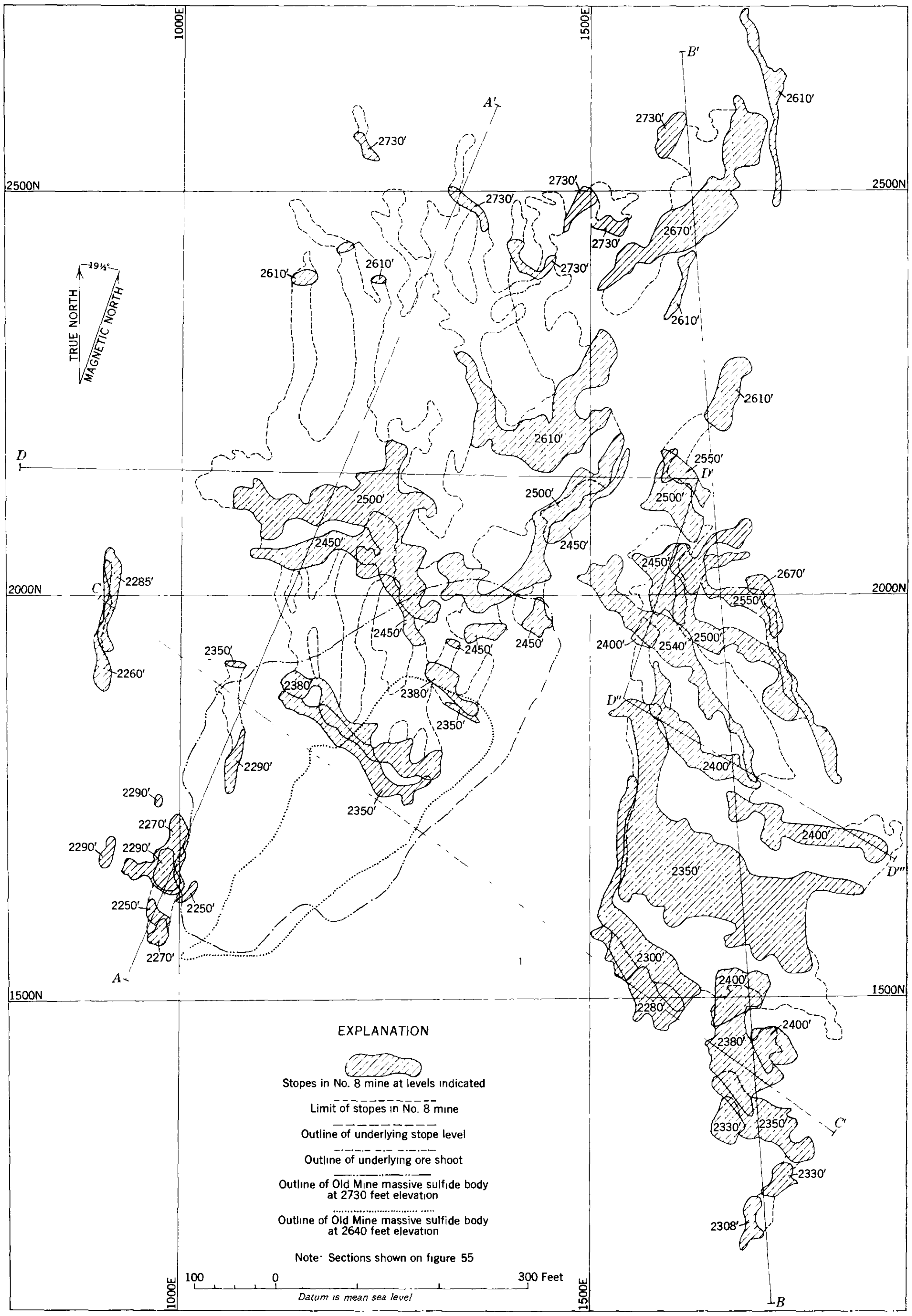


FIGURE 54.—Map of No. 8 mine ore bodies, Iron Mountain mine. (Compiled from maps furnished by the Mountain Copper Co., Ltd.)

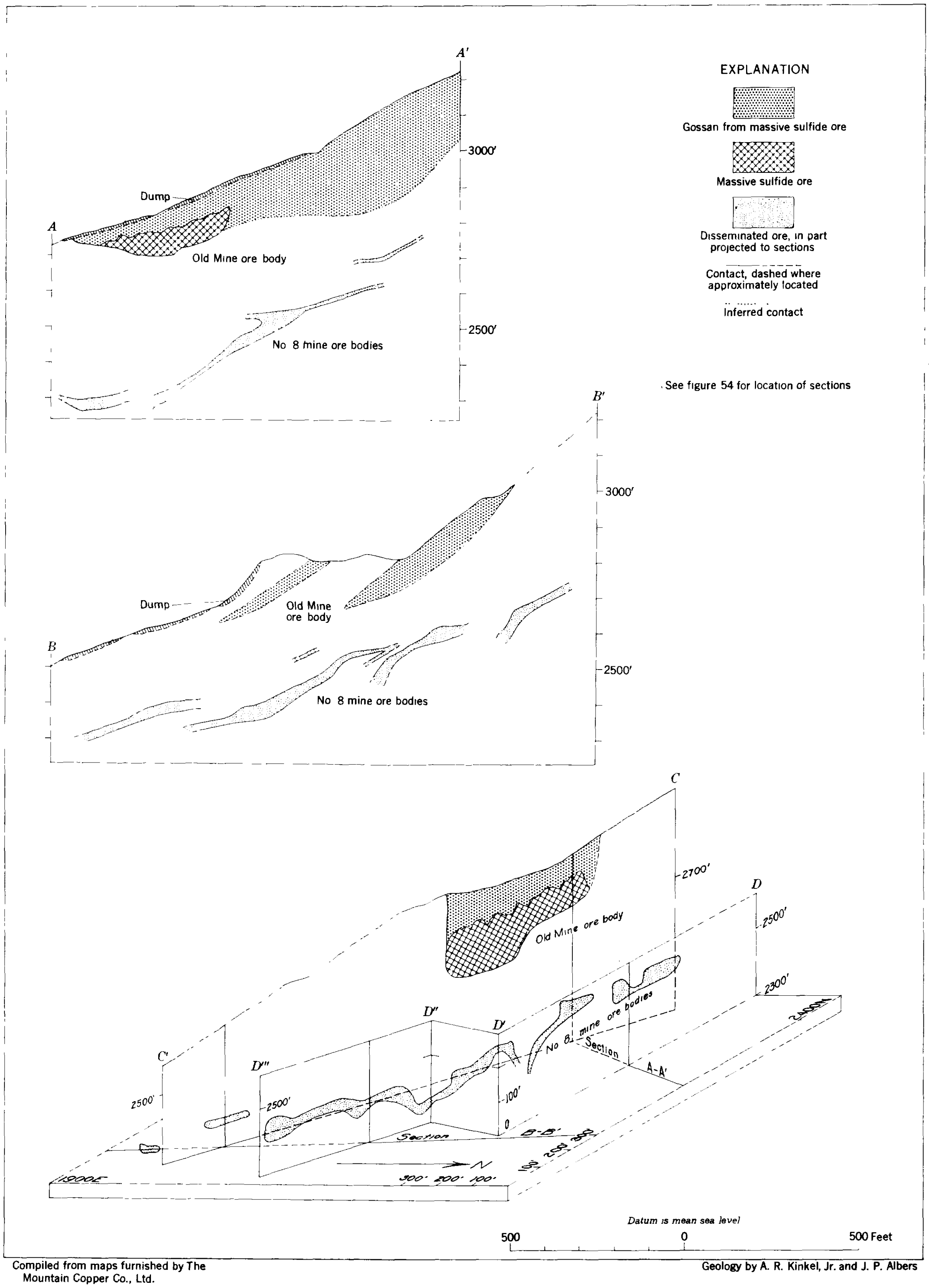


FIGURE 55.—Sections of No. 8 mine ore bodies, Iron Mountain mine. Location of sections is shown on figure 54.

and the face of the adit is in massive sulfide. This massive sulfide resembles the sulfide of the Hornet ore body, and contains little copper or zinc. It is bounded on both sides by steeply dipping faults that have about 1 foot of gouge; these faults are probably the continuation of those that are along the walls of the Hornet ore body.

RELATIONSHIP OF ORE BODIES TO STRUCTURES IN THE HOST ROCK

The relationship between ores and rock structures at Iron Mountain is similar to those described at the Mammoth mine, where the relationship is more clearly exposed in steep-walled canyons. Rock structures that were formed by flexural-slip folding are prominent at the Iron Mountain mine, as at the Mammoth mine. Foliation near the massive sulfide contact is always parallel to the contact, and at no place does the ore contact cut across foliation. The massive sulfide shows no tendency to finger into foliated rocks except in a few minor occurrences at the ends of small ore bodies. On the other hand, zones of disseminated pyrite that parallel the ore bodies and chalcopyrite zones in the No. 8 mine are replacements of foliated rocks along the planes of foliation. The sulfides show no evidence of crushing or rounding and were obviously deposited in a foliated rock, and the foliation controlled the direction of travel of the sulfide-bearing solutions.

The contact between massive sulfide ore and the enclosing porphyritic or nonporphyritic rhyolite is usually abrupt, although several contacts showing what appears to be gradational replacement have been found. At most localities no visible change can be seen in the massive sulfide ore as the contact is approached and the contact has massive sulfide on one side, and unmineralized soft white claylike gouge on the other. This gouge is commonly more than a foot thick, but it grades from structureless white gouge against the ore to strongly sheared, sericitized, porphyritic rhyolite as the more solid wall rock is approached. It consists of highly sheared, altered, porphyritic rhyolite, as relict quartz phenocrysts occur in the claylike part of the gouge. Gouge occurs along all observed contacts, even though it may be less than an inch thick.

The gouge along the contact locally contains some pyrite in the footwall of the Richmond ore body. In these places, the gouge contains less clay and is a sericite schist, but a narrow clay seam separates the massive pyrite from the sheared, pyritized wall rock. The pyrite is in small euhedral grains that show no evidence of crushing, and poorly defined layers of disseminated pyrite in the sericite schist parallel the ore contact and extend several feet into the footwall, diminishing in pyrite content as the distance from the ore increases. The mineralized layers have indistinct boundaries and are a replacement of foliated material parallel to the contact. A few quartz veinlets or irregular bodies of more siliceous material may also parallel the ore contact, but no chalcopyrite has been found in these zones. It should be emphasized, however, that along most ore contacts the wall rock contains little or no pyrite.

Specimens of ore from the Iron Mountain mine that contain streaks of chalcopyrite and sphalerite were seen on dumps, but none was found in place. Members of the mine staff report that the southeast wall of the Hornet and the northwest wall of the Mattie contained a little banded ore, but most of the ore at Iron Mountain is massive and structureless.

FAULTS

Movement has occurred at or near the contact of all the massive sulfide ore bodies. It is not possible to determine how much of the gouge along the ore contact is premineral in origin, having acted as a guide for sulfide replacement, and how much is due to postmineral movement. At many places massive sulfide ore is slickensided—on slips within ore bodies, on walls against contact gouge zones, and on fragments of sulfide within the gouge zones. Gouges that contain crushed sulfide grains and broken or slickensided fragments of massive sulfide ore are postmineral faults, and all faults on which ore bodies are offset contain crushed and slickensided sulfides.

The major postmineral faults in the mine are the Scott, the Camden, and the *J* faults, but only the Camden is both postmineral and premineral. The Scott fault is curved in strike and dip. It is unquestionably a postmineral fault as it contains much crushed sulfide ore and fragments of slickensided sulfides. The fault zone, where it is exposed in underground workings, is 3 to 5 feet wide and contains many anastomosing slickensides and much dark-gray to white gouge. However, it is reported to be less than a foot thick at a few localities at the end of the Hornet ore body. The Scott is a normal fault that dips about 50° NE., but it flattens along the lower part of the Hornet ore body. The movement on the Scott fault was essentially dip slip, and the Hornet ore body has been dropped 250 feet below the corresponding section of the Richmond–Complex. Flat-dipping faults have also been reported along the top of the Hornet and across the northeast part of the Complex ore body near the Scott fault, but little information is available on these faults.

The Camden fault can be seen in many places in the underground workings; in most of these exposures it consists of 1–5 feet of gouge and sheared rock that contains crushed sulfides, but the width of the fault zone is at least 50 feet where it is exposed at the west end of the 2,600-foot level. The fault forms the southeast wall of the Complex ore body and turns to form the south end of the Richmond ore body, which is the offset part of the Brick Flat ore body. The Camden splits into several strands near its west end. The first displacement on the Camden resulted in a dip-slip movement of about 350 feet and occurred along the main Camden and the fault that is now called the Camden North fault, which at the time of the first displacement was a continuation of the main Camden. Movement then occurred on the *J* fault, displacing the western part of the Camden. Renewed movement along the main Camden displaced the *J* fault, forming a new break called the Camden South fault, which is a continuation of the main Camden west of the *J* fault.

The Camden North fault displaced the Richmond from the Brick Flat ore body, and the Camden South fault displaced the Brick Flat from the now oxidized, updip extension of the Old Mine ore body. The displacement along the Camden fault is also shown by the offset of the Copley greenstone and Balaklala rhyolite contact shown on sections 38 and 44, plate 9.

A strong fault zone occurs along the northwest side of the Hornet ore body, and faults along its southeast side are reported to have been seen underground. The alinement of these faults with the Camden suggests that movement along them may also have occurred before and after mineralization, as on the Camden. A continuation of these faults to the northeast into Sugarloaf mountain is known as the Sugarloaf fault.

Many small postmineral faults are found within the ore bodies. These can best be seen along the massive sulfide contact, where they offset the contact as much as 50 feet. They are marked by slickensided surfaces in massive sulfide ore, or by gouge. However, some gouge within the main sulfide mass is so continuous, and so far from the boundaries of the ore, that it can only represent sheared remnants of unreplaced rock or premineral fault gouge.

HYDROTHERMAL ALTERATION OF THE WALL ROCKS

The rocks in the vicinity of the Iron Mountain mine have been altered by sericitization, silicification, chloritization, and pyritization.

Sericite and hydromica are widespread and although at some places they probably were formed by hydrothermal solutions related to the ore-bearing solutions, at other places they have no close spatial relationship to metallization. In places intensely altered zones of sericite and hydromica are closely associated with ore bodies, but they do not necessarily form halos around them. The altered rock is either white or lavender. The alteration of feldspar in rhyolite to sericite and hydromica occurs along the walls of ore bodies and along faults, and as irregular zones in rhyolite near but not necessarily adjacent to ore bodies. Sericitic wall rock is found at the ore boundaries as a gradation between gouge and the less sheared rock, but this sericite may have been deposited before mineralization and may not have been formed by the ore-depositing solutions, as large bodies of sericite schist are found far from known ore zones.

Silicified rocks are common at Iron Mountain; leached bodies of pyritized and silicified rock show all gradations from rock with a few pyrite casts in a silica matrix to a porous silica sponge in which the silica occurs as septa between pyrite grains. Where pyrite did not replace all the rock, relict quartz phenocrysts of the porphyritic rhyolite are found in a rock that is composed almost entirely of secondary silica. Silicification in the mineralized zone is not always coextensive with pyritization, but zones of strong silicification in the porphyritic rhyolite appear to indicate the main solution channelways in the mineral belt.

Pyritization at the Iron Mountain mine is more limited in distribution than any of the alterations previously described. It occurs in broad mineralized zones, but massive sulfide bodies are limited to a small part of the pyritized zones. Massive sulfide deposits in the Iron Mountain area are associated with pyritized rock, but the bodies of pyritized rock are commonly not in immediate contact with massive sulfide ore bodies. On the other hand, large pyritized areas that contain neither massive sulfide ore bodies nor copper or zinc minerals are found in the district. The immediate walls of the massive sulfide ore bodies at the Iron Mountain mine contain little or no pyrite except in the vicinity of the Old Mine ore body. Bodies of disseminated pyrite parallel the sulfide deposits in the same mineralized zone, but disseminated pyrite does not occur as a halo around most of the massive sulfide ore bodies in the Iron Mountain mine.

SUMMARY OF FEATURES CONTROLLING ORE DEPOSITION

The broad controls believed to be responsible for the localization of ore at the Iron Mountain mine are the folds that acted as a guide for solution travel, the thick cover of shale that was present several hundred feet above the ore zone at the time of ore deposition, and the main feeder channels. More detailed controls that served to localize individual ore bodies are preferred layers in the flows and pyroclastic rocks of the Balaklala rhyolite, fractures that served as feeder channels, and premineral fault gouge that confined the solutions to limited areas.

The massive sulfide ore contains few recognizable remnants of unreplaced rock. Exposures showing incompletely replaced rock can be seen in some of the quarry benches, and these exposures suggest that replacement favors massive porphyritic rhyolite that contains 2- to 4-millimeter quartz phenocrysts. Massive nonporphyritic rhyolite, flow-banded rhyolite, and rhyolitic volcanic breccia seem to be unfavorable host rocks in the Iron Mountain area. There is no indication that massive sulfide ore preferentially replaced more foliated parts of the rock at this mine, although the widespread disseminated pyrite tends to concentrate along broad zones of foliation.

Feeder channels and bands of fault gouge guided the mineral-bearing solutions in detailed control. Where these channels were not confined by gouge, zones of disseminated pyrite that lack sharp walls and only partly replace the host rock were formed; but where bands of gouge were present the mineral-bearing solutions were confined and locally replaced all the rock. Evidence of premineral bands of gouge is indicated where some bands extend into the ore in the form of long narrow sheets that show too little movement of the ore bodies on either side of the gouge to account for its formation as dragged wall rock. This gouge was present at the time of mineralization but was not replaced. Although clay gouge occurs at the contact of all the ore bodies, some of this type of gouge may have been formed by minor movement of the sulfide bodies against hydrothermally altered rocks, and does not represent a large amount of movement or necessarily a premineral gouge.

The Camden and Sugarloaf faults, which may be the same fault or may be en echelon in the vicinity of the Hornet ore body, are premineral in age. This is shown by the presence of gossan and of pyritized and hydrothermally altered rocks along both faults.

The Camden fault and its probable continuation to the east, the Sugarloaf, appear to have been the channel for the solutions that formed the Busy Bee, Hornet, Richmond-Complex, and Brick Flat ore bodies, as well as the small areas of gossan shown along the eastern and western extensions of these faults. The Camden probably was not the feeder channel for the formation of the Old Mine ore body and No. 8 mine ore body because these dip away from the fault and because of the upward-branching ore pattern in the No. 8 mine. Unrecognized or concealed solution channels probably are south of these ore bodies.

Concentrations of chalcopyrite and sphalerite in the massive sulfide deposits, with the exception of those in the Old Mine ore body, all lie near the Camden fault, marking this as a feeder. These concentrations are found in the Mattie, the southeast side of the Complex, in the Richmond, and in the Brick Flat ore bodies along the north branch of the Camden fault. The concentration of chalcopyrite at the base of some of the ore bodies adjoining the Camden fault also indicates that upward-moving, copper-bearing solutions were traveling along it.

OXIDATION AND ENRICHMENT

Two types of gossan have been distinguished at the Iron Mountain mine (pl. 9). One of these was derived from disseminated pyrite and ranges from rock containing scattered pyrite casts to rock and silica sponge in which the rock structure is preserved, but which may have contained 50 percent or more pyrite. Rock that is estimated to have contained less than 10 percent pyrite is not shown as gossan on the map. The second type of gossan is that derived from massive sulfide ore. This gossan contains no discernible traces of the original rock structure. It consists of limonite in the form of earthy, spongy, and cellular masses with quartz septa (fig. 45), or of limonite crusts and dense limonite, or a breccia of angular fragments of rock and vein quartz embedded in limonite.

Small relict nodules of massive pyrite have been found in gossan within 10 feet of the surface in the upper part of the gossan quarry at Iron Mountain (fig. 46). However, the high relief and broken, porous character of the ground allows more or less complete leaching and oxidation to a depth of several hundred feet throughout most of the mineral zone. The deepest oxidation extended down along the footwall of the Camden fault, where oxidized material is found locally at a depth of 400 feet.

The copper-bearing disseminated ore does not crop out. No correlation was possible between the types of limonite found in the gossan and the copper content of the massive sulfide bodies.

In gossan derived from pyritized rock, bands of the rock parallel to the schistosity are locally completely replaced by silica and pyrite in ribbonlike structures. The ribbons commonly contain more than 50 percent silica and tend to stand out as resistant ribs of silica sponge on weathered surfaces. Material between the mineralized bands contains less silica and pyrite, or it may be entirely unmineralized.

Supergene enrichment.—The Old Mine ore body is the only body of sulfide ore in the Iron Mountain mine that contained supergene copper minerals. The other ore bodies in the mine area were either completely converted to gossan and the copper has been removed or they were protected from oxidation by gouge along the top of the ore bodies. No secondary copper minerals are visible in the gossan. The gossan above the Old Mine ore body contained 0.4 percent copper and no zinc, but a thin enriched zone in decomposed sulfides along the irregular base of the gossan had a high content of copper and an exceptionally high content of silver. This ore zone is reported to have been several feet thick and to have consisted of clay, and black decomposed sandy sulfides that graded downward into massive sulfide ore. No assay records are available but considerable silver ore was mined from the enriched layer.

A few specimens of ore from the Old Mine ore body have been collected from dumps on the property. They are fine-grained massive sulfide ore that has a bluish color. Thin seams of chalcocite can be seen in the hand specimens, and other secondary sulfide minerals may be present. The following assays compiled by Seager in an unpublished report indicate the amount of copper enrichment in the upper part of the Old Mine ore body. (See table 12.)

The average gold content of the gossan above the Old Mine ore body is 0.073 ounce per ton and that of the massive sulfide ore below the gossan is 0.04 ounce per ton. This is a residual enrichment caused by leaching and oxidation of the sulfide minerals. The average weight of the gossan is 165 pounds per cubic foot and that of the sulfide ore is 275 pounds per cubic foot.

TABLE 12.—*Copper content of the upper part of the Old Mine ore body, Iron Mountain mine*

Ore	Altitude (feet)	Average copper content (percent)
Gossan	Above 2,741	0.25
Massive sulfide	2,741	10.0
Do	2,732	10.0
Do	2,720	8.0
Do	2,710	8.0
Do	2,682	6.0
Do	2,614	6.0
Do	2,601	3.5

Water-soluble gold and mercury have been reported by the mine staff and in publications as occurring in the gossan at Iron Mountain. In describing the cyaniding of the gossan Averill (1938, p. 329) reports:

> One interesting problem solved is the detection of water-soluble gold. Difficulty was at one time experienced in making assay-values check with mill-recovery. Experiments showed that in some samples as much as 44% of the gold would dissolve in distilled water. This gold was at first supposed to be in a colloidal state, but tests proved that it is not. Finally micro-chemical tests with pyridine showed the presence of gold chloride ($AuCl_3$). Of the gold in the ore, it is not unusual for 10% to 13% to be in this water-soluble form. All of the ore does not contain it.

Jackson and Knaebel (1932, p. 125) report that the gossan contained 0.4 percent copper, 50–55 percent iron, 5–10 percent silica, and a little arsenic and mercury, and that from a production of 543 tons per day, 40 pounds of mercury was recovered by a condenser at each monthly cleanup.

It seems improbable that water-soluble gold or mercury occurred naturally in the gossan. It seems more likely that these metals were added to the gossan by losses during early milling operations for gold and silver, in the amalgamating and chloriding plants. In the silver plant, which was destroyed by fire in 1897, silver-bearing gossan was mixed with salt and roasted, and about 200 pounds of quicksilver was added for each 4 tons of pulp before pan amalgamation (Kett, 1947, p. 108).

No gold could be panned from gossan at the Iron Mountain mine, probably because it occurs in extremely fine particles.

Silver and zinc have been removed from the gossan over the Old Mine ore body, and although the silver appears to have been precipitated at the contact between the gossan and the sulfides, no bodies of secondary zinc minerals have been found.

EXPLORATORY DRILLING AT IRON MOUNTAIN

In 1950–51 a diamond-drill hole 1,798 feet deep was drilled at Iron Mountain to obtain additional information about the rock structure and to explore a syncline similar to the structure that localizes the ore bodies in the Iron Mountain mine. This syncline lies parallel to and 1,500 feet southeast of the Iron Mountain ore bodies. Feeder channels were thought to be present south of the Richmond-Complex ore bodies, and the hole was drilled to explore the area where these feeder channels would intersect the large syncline. The outline of the area and the location of the drill hole is shown on figure 56.

No deep exploratory work had been done in the syncline south of the ore bodies before this drilling, and no mineralization was known in the area. Geologic mapping suggested that the favorable zone of porphyritic rhyolite should be at a vertical depth of about 800 feet. In addition, mineralization is known to trend toward this syncline in the disseminated chalcopyrite ore of the Confidence-Complex vein system.

DRILLING DATA

The drill hole was started on an inclination of 70½° and on a bearing of N. 42½° W. It was anticipated that the hole would cut through the Balaklala rhyolite into Copley greenstone at a depth of about 1,200 feet. However, the syncline proved to be much deeper than

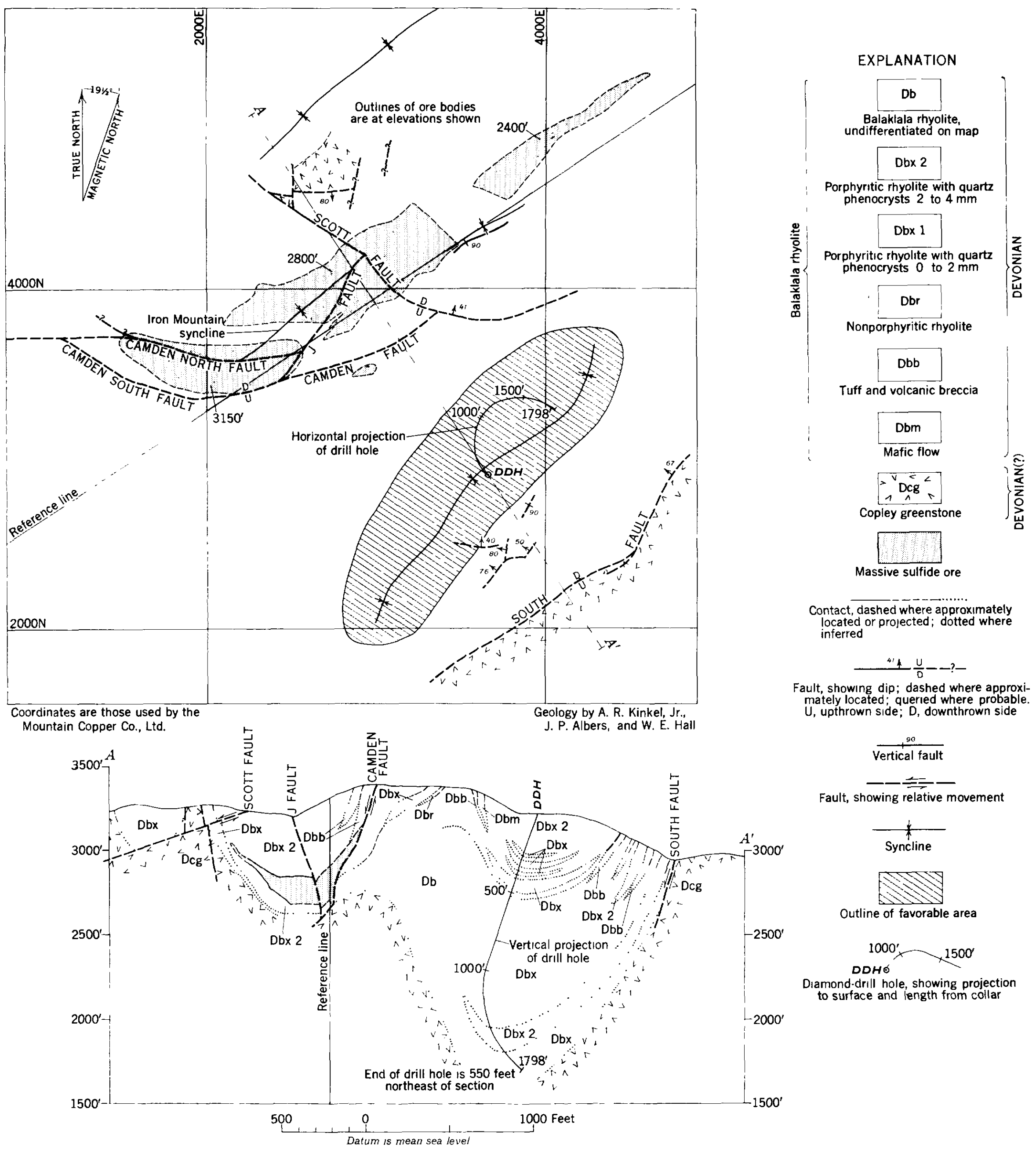


FIGURE 56.—Map of Iron Mountain area showing location of U. S. Geologic Survey diamond-drill hole and geologic cross section through drill hole.

was expected, and the hole was still in Balaklala rhyolite at a depth of 1,798 feet, where it was abandoned because of deflection. From a bearing of N. 42½° W. at the collar the hole deflected to S. 57½° E. at 1,760 feet, and from a dip of 70½° it flattened to 38°.

DESCRIPTION OF THE CORE

The diamond-drill core consisted of porphyritic and nonporphyritic rhyolite, rhyolitic pyroclastic material, and a small amount of chloritic rock. A geologic columnar section of the hole is given in figure 57. Above a depth of 520 feet the core was predominantly porphyritic rhyolite and contained 2- to 3-millimeter quartz and feldspar phenocrysts in an aphanitic groundmass. The porphyritic rhyolite was weathered and contained clay minerals and limonite to a depth of 266 feet. Below 520 feet the core was mainly siliceous nonporphyritic rhyolite and had a mottled appearance due to irregular patches and streaks of chlorite. In three places chlorite was very abundant, and the rock was composed only of

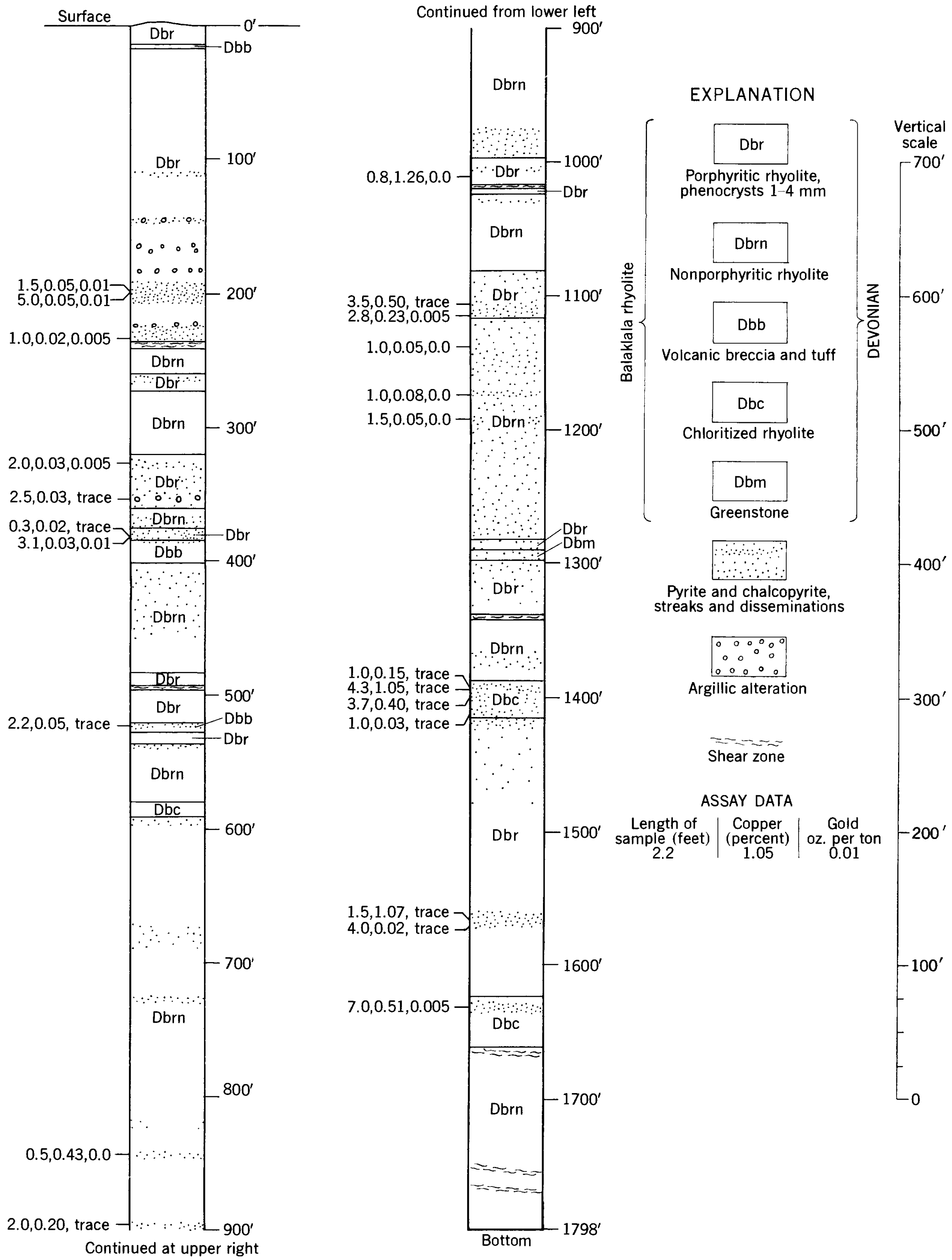


FIGURE 57.—Graphic log of the U. S. Geological Survey diamond-drill hole at the Iron Mountain mine.

quartz and chlorite. Most of the core contains a little disseminated pyrite in the form of tiny cubes. Pyrite was more abundant where the rock was sheared, and some of the core contains abundant disseminations and seams of pyrite as much as 6 inches thick for distances of as much as 6 feet. However, bodies of massive pyrite similar to those at the Iron Mountain mine were not found.

Twenty-two samples of core were assayed for copper and gold; the assays are shown on the columnar section (fig. 57). Parts of the core that contain no visible chalcopyrite were not assayed even though they are pyritized. The rhyolite that contains chalcopyrite generally contains more chlorite than the average rhyolite of the district, and in places the rhyolite that contains chalcopyrite has been largely replaced by quartz and chlorite.

GEOLOGIC RESULTS OF DRILLING

The drill hole at Iron Mountain indicates abundant pyrite and some chalcopyrite at a considerable distance from known ore bodies, but evidence of bodies of commercial ore was not found. The cores show chalcopyrite at depth in the syncline southeast of the known ore bodies, and indicate either that additional feeder channels occur in this area, or that copper-bearing solutions entered the wall rock from known feeder channels, such as the Camden fault, at a considerable depth below the known ore bodies. In addition the drilling demonstrates that detailed geologic mapping of the lenticular, obscurely bedded volcanic rocks will yield sufficient information to predict the occurrence and shape of folded structures, such as the syncline that was drilled.

KEYSTONE MINE

Location.—The Keystone mine is in sec. 14, T. 33 N., R. 6 W., about half a mile southwest of the Balaklala mine. The portal of the Keystone adit is at an altitude of 3,000 feet on the south slope of the canyon of Squaw Creek. The mine was made accessible in 1950 by the repair of a steep dirt road that extends beyond the Balaklala mine.

History and production.—Little is known of the early history of the Keystone mine. Assessment work had been done on the mine by 1902 (Aubury, 1902, p. 90), but no additional work had been done up to 1908 (Aubury, 1908, p. 100). Exploratory work was carried on in 1918, and the Keystone is listed as a producing mine in 1923 in the California State Mining Bureau report for that year. The mine was active until the fall of 1925. It was operated by the United States Smelting Refining and Mining Co. and is still owned by that company. The Keystone mine has not operated since 1925, and the portals of the main adits are caved. The accompanying maps (pls. 5, 6) of the mine are based upon existing mine maps and records and are in part interpretations of such data by the writers. The recorded tonnage and grade of mined ore is 122,000 tons that averaged 0.06 ounce of gold, 2.7 ounces of silver, 6.0 percent copper, 8.0 percent zinc, 30 percent iron, 35 percent sulfur, and 15 percent insoluble material.

Geology of the mine area.—The geologic setting of the Keystone mine is similar to that of the Balaklala mine to the north (pls. 5 and 6, *H–I–J*). The ore bodies are overlain by a gently dipping contact between the upper and middle units of the Balaklala rhyolite. Nothing is known of the rock type in the middle unit of the Balaklala adjacent to the ore, as the underground workings are inaccessible. Where this zone is exposed at the surface near the portal of the Keystone adit, the upper part of the middle unit is porphyritic rhyolite that contains 1- to 2-millimeter quartz phenocrysts, and nonporphyritic rhyolite. A very small amount of yellow rhyolitic tuff and local areas of rhyolitic breccia occur along the contact. The bed of coarse volcanic breccia above the ore at the glory hole of the Balaklala mine 2,000 feet to the northeast lenses out along the contact before it reaches the Keystone portal. The ore bodies of the Keystone mine occur in the updip extension of the productive zone that contains the Balaklala ore bodies. The upper unit of the Balaklala rhyolite near the portal of the Keystone adit is a coarse-phenocryst tuff that contains a few bodies of massive coarse-phenocryst rhyolite.

Character of the ore bodies.—At the time of mine operation, the massive sulfide ore in the Keystone mine was subdivided on the basis of copper content into minable and below minable grade. Both types of ore are massive sulfide bodies, but minable ore occurs in some parts of the mine as separate bodies and in other parts as copper-rich parts of large masses of pyrite that otherwise are very low in copper. Some large bodies of massive pyrite occur that contain a little copper or zinc (pl. 12).

The ore bodies in the Keystone mine are in a productive zone, at least 300 feet thick, in the middle unit immediately below the base of the upper unit of the Balaklala rhyolite. Some of the ore lies just below the contact, but the major part of it was mined 50–80 feet below this contact. The higher grade copper ore is bounded at most contacts by strong shear zones, and may have replaced sheared rock. Stopes outline tabular, gently dipping copper-rich ore bodies that parallel the contact between the upper and middle units of the Balaklala rhyolite, but most of these ore bodies dip 5°–10° more steeply than this contact. The average thickness

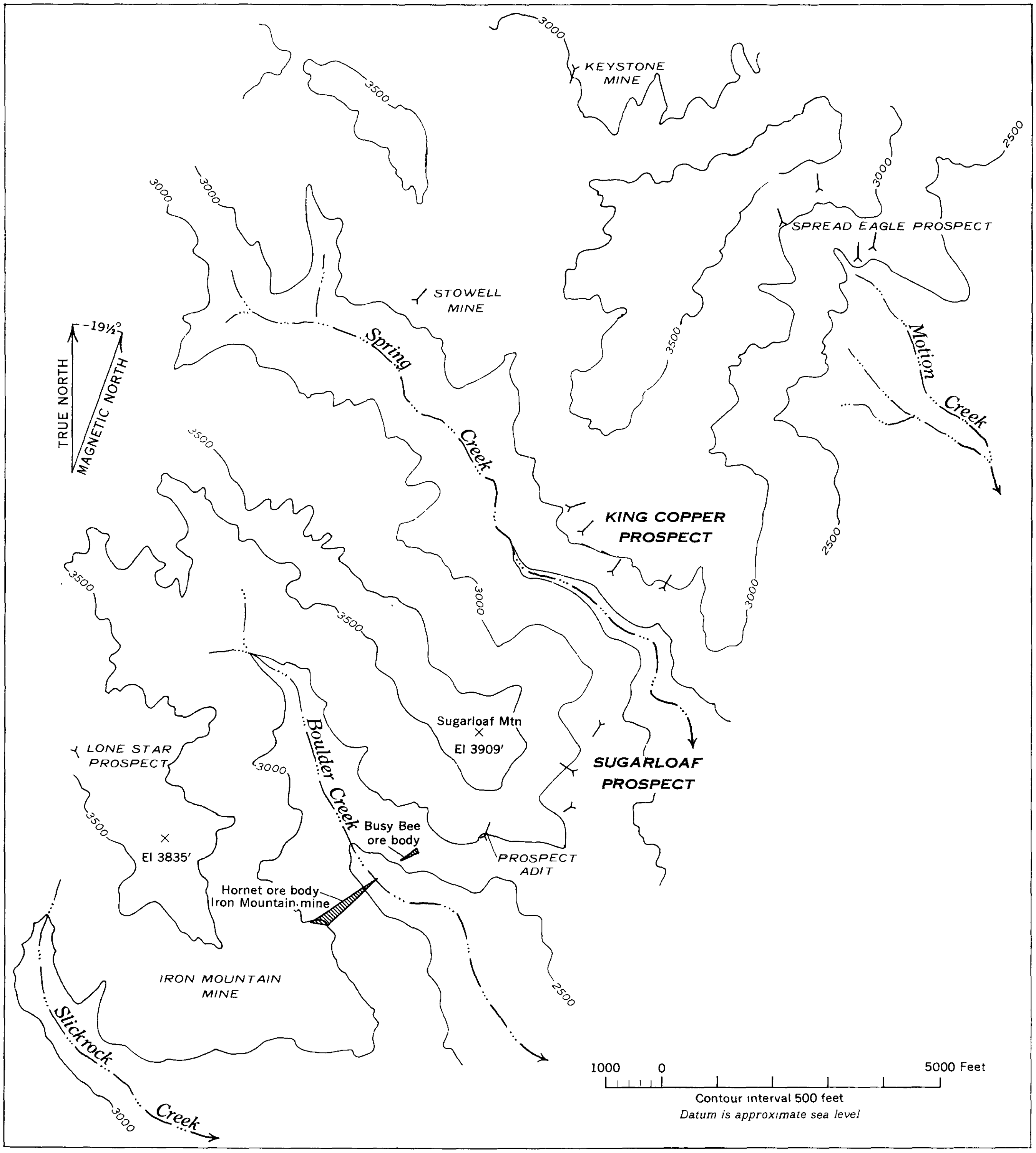


FIGURE 58.—Location of the King Copper and Sugarloaf prospects in relation to nearby mines and prospects.

of ore in the stopes ranged from 10 to 20 feet and the maximum thickness was 50 feet.

The Keystone mine contains large bodies of massive sulfide ore of low copper content. As the ore was direct-smelted at the time of mine operation, these bodies did not constitute ore. Consequently their limits were not completely outlined by drilling. Some bodies of low-grade massive sulfide are cut by mine workings, but what appear to be some of the largest bodies are indicated only by a few drill holes. These low-grade bodies are also bounded by faults or shear zones where they were exposed in the mine workings. Higher grade copper ore bodies occur as parts of lower grade sulfide masses on levels 300 and 318 (pl. 12), but at most places the two types form separate bodies. The massive sulfide bodies that have low copper content are not sufficiently explored to allow an estimate of tonnage, but it probably is large. The grade of the low-copper pyritic ore, from records of the United States Smelting Refining and Mining Co., is 0.01 ounce of gold, 0.3 ounce of silver, 0.54 percent copper, and 43 percent sulfur. The copper content as indicated by available data ranges from 0.1 to 2.5 percent. These massive sulfide bodies are similar to the Hornet ore body at the Iron Mountain mine.

Ore controls and exploration possibilities.—The main ore control at the Keystone mine is the contact between the upper and middle units of the Balaklala rhyolite. The writers have no information on rock types in the mine other than the location of the contact between these two units, and no folds in this contact are known unless the curved contact between the two units that is shown on the 275-foot level indicates a minor fold. A vertical control by the contact seems obvious, but no horizontal controls are known. There is a suggestion, however, that some of the ore bodies may have formed against or along faults. At several places the bodies of massive sulfide ore are terminated by steep faults. Faulted extensions of these ore bodies have not been found in existing mine workings and drill holes. Such steep faults may have been feeder channels for ore solutions, which deposited ore bodies along flat subsidiary shear zones.

The productive zone, that is, the one immediately below the upper unit of the Balaklala rhyolite, is covered in the area between the Keystone and the Stowell mines, a distance of 3,000 feet. The productive zone should extend throughout most of this area at a moderate depth, as the upper unit of the Balaklala is known to be about 500 to 600 feet thick where it has been drilled between the two mines. A limited amount of exploratory drilling has been done between the Keystone and the Stowell mines, but a large area east, south, and west of the Keystone remains to be explored.

KING COPPER PROSPECT

The King Copper prospect is located in secs. 23 and 24, T. 33 N., R. 6 W., on the northeast side of Spring Creek (pl. 4). Many prospect adits were driven on pyritized zones in the Balaklala rhyolite, but no minable ore was developed. Aubury (1902, p. 81) reports that the King Copper prospect was explored by about 1,000 feet of adits by 1902, and apparently little work has been done since that time. The property is now owned by R. T. Walker and W. J. Walker.

Geology.—The rocks in the vicinity of the King Copper prospect are interlayered flows of porphyritic and nonporphyritic rhyolite that contain much flow breccia and some pyroclastic material. They probably belong to the middle unit of the Balaklala, although as at the Sugarloaf prospect to the southwest (fig. 58) their stratigraphic position is obscure because of the difficulty in distinguishing the middle and the lower units of the Balaklala in this area. The Copley greenstone crops out about 3,000 feet southeast of the prospect, and marks the lower limit of the Balaklala rhyolite. The rocks near the prospect have a steep foliation, but no bedding was found.

Ore deposits.—Many pyritized zones are in the vicinity of the King Copper prospect, and areas of strongly foliated rhyolite have been hydrothermally altered to soft crumbly white and lavender rock. Zones of hydrothermal alteration and pyritization range from scattered thin seams to fairly well defined bands several tens of feet in width. Most of the pyritized bands are also silicified. The bands are classed as pyritized, hydrothermally altered rhyolite rather than gossan, as the material consists of scattered euhedral pyrite in silicified, argillized rhyolite, or in secondary silica. The most heavily pyritized material on the dumps of the King Copper adits contains only about 50 percent pyrite. A few small lenticular bodies of gossan, less Balaklala in this area. The Copley greenstone crops out near some of the adits, but the only body of gossan that appears promising at the surface is one that crops out immediately above the second adit from the north as shown on figure 58. This gossan occurs in silicified porphyritic rhyolite; it is 2 feet thick and possibly 50 feet in strike length. It dips flatly to the north and appears to be localized where the normally steep foliation bends and lies flat.

Diller (1906, p. 13) reports that

> The general trend of the small ore body of the King Copper, on the slope of Spring Creek is N. 70°–76° W. It dips 84° SW. and apparently agrees quite closely with the position of the

Spread Eagle ore body, transverse to the general direction of the Iron Mountain lode.

The location of the ore body referred to by Diller is not known.

The area between the Sugarloaf and the King Copper prospects, and northeast of the King Copper, is one in which pyritization is widespread, and these prospects appear to be alined along the trend of a mineralized belt, but no bodies of massive sulfide have been located in the zones of pyritization.

LONE STAR PROSPECT

The Lone Star prospect, which is owned by the Mountain Copper Co., Ltd., is at the southwest end of the West Shasta copper-zinc district in sec. 27, T. 33 N., R. 6 W. It is accessible either by a road from the Iron Mountain mine or by a road that extends north along the ridge from the South Fork Mountain lookout station. The prospect is on a massive sulfide body that is explored by two short adits, but no ore has been mined. An inconspicuous gossan is exposed where the ore body crops out on a steep hillside. The rocks near the ore body are largely covered by slope wash.

Geology of the mine area.—The massive sulfide body at the Lone Star mine is in porphyritic Balaklala rhyolite, but nonporphyritic rhyolite, tuffaceous shale, and amygdaloidal andesite occur near the ore body (pl. 13). Both the porphyritic and the nonporphyritic rhyolite are light-green siliceous weakly foliated rocks. The porphyritic rhyolite contains quartz phenocrysts that average about 2 millimeters in diameter. The thin flow of amygdaloidal greenstone which crops out south of the ore body, is a dark-green chloritic rock. At the surface the fillings of the closely spaced, round vesicles in the greenstone have been dissolved, leaving iron-stained cavities.

The bed of tuffaceous shale north of the ore body is a poorly exposed, buff-colored rock without distinct bedding. The shale apparently lies at about the same horizon as the bedded crystal tuff and tuffaceous shale that occurs at Iron Mountain just northwest of the Brick Flat ore body (Kinkel and Albers, 1951, pl. 1). These shale beds are correlated with the crystal tuff, 800 feet northeast of the Lone Star prospect, in which a fossil was found that was determined by D. H. Dunkle of the National Museum as a fish plate from an euarthrodiran fish close to *Titanichthys* of Middle Devonian age. The tuffaceous shale and crystal tuff in this area are at the horizon of the transition zone between the upper and middle units of the Balaklala rhyolite, but it is doubtful that the upper unit extended this far southwest.

Ore deposit.—The gossan at the Lone Star prospect is a typical massive sulfide gossan composed of dense and cellular limonite, but is in part a collapse breccia, which contains fragments of wall rock cemented by limonite. The gossan forms an arch. At this prospect material formed by the cementation of soil and dump material by iron-bearing waters that issue from the mine portal superficially resembles gossan.

It is apparent from surface exposures and from drilling that the ore body was essentially a flat-lying lens of massive sulfide with sharp contacts. The wall rock, including the fragments in the collapse breccia, is only slightly pyritized. The adits were not accessible, but a lack of gossan in dumps, and underground maps made by L. C. Raymond for the Mountain Copper Co., Ltd., indicate that the gossan extended only a few tens of feet in from the surface. Most of the mineralization in the underground workings was massive sulfide.

The size of the ore body is not adequately determined. Maps and assays furnished by the Mountain Copper Co., Ltd., show that drill hole 5 cut 45 feet of massive sulfide along the length of the ore body, but as indicated in plate 13, the total length of the lens may be about 85 feet. The width in the upper adit is about 20 feet, and the maximum thickness is about 10 feet. The ore body is cut by many small faults, but it is not known whether the small slivers of ore are faulted segments of a formerly continuous massive sulfide body, or isolated lenses of massive sulfide.

Assays of diamond-drill core and of ore from the dump show that this small ore body contains more copper than most of the ore at the Iron Mountain mine, and that it resembles some of the smaller higher grade satellite ore bodies at other mines in the district. The ore contains visible chalcopyrite and sphalerite in massive pyrite and few gangue minerals. Individual assays of the massive sulfide from drill core range in copper content from 0.62 to 6.89 percent, but the average of the ore cut by drill holes is 3.74 percent copper. No assays were made for gold, silver, or zinc. Ore specimens on the dump appear to contain more sphalerite than chalcopyrite.

A prominent, steeply dipping, east-west fault is exposed in surface cuts north of the ore body. The north side is thought to be the downthrown block as the shale that crops out north of the prospect is not found to the south. The amount of vertical (and possibly horizontal) movement is not known, but three holes drilled in the downthrown block failed to locate a faulted extension of the ore body. It seems probable that a faulted extension of the known ore body is north of the fault, either below or to one side of the present drill holes.

MAMMOTH MINE

The Mammoth mine is in the northern part of the West Shasta copper-zinc district in sec. 32, T. 34 N., R. 5 W., on the south side of Little Backbone Creek at an altitude of about 3,000 feet. The mine is owned by the United States Smelting Refining and Mining Co. During mining operations from 1905 to 1925 the mine was accessible by standard gage railroad up Little Backbone Creek and by funicular from the railroad siding to the mine's main adit, the 470-foot level. The flooding of Shasta Lake in 1944 submerged Little Backbone Gulch and the lower part of an unimproved road to the mine. The road is 3 miles long, and climbs 2,000 feet in that distance. In 1950 this road was not passable for a car. The mine was closed in 1925 and the mine plant removed.

The report and the accompanying maps (pls. 14, 15) on the mine were prepared by combining the surface geologic mapping, done by the U. S. Geological Survey, and the underground geologic information obtained from mine records. As many of the workings in the vicinity of the ore bodies, including all the stopes, were inaccessible at the time of study, no underground mapping was done. Information in this report is summarized from a more detailed report on the mine by Kinkel and Hall (1952).

HISTORY, PRODUCTION, AND GRADE

The date of discovery of the Mammoth mine is not known but it is assumed to have been after 1880 and probably before 1890. George Graves is reported to have discovered the ore body, and before 1900 a Mr. Nelson worked the property on a small scale to recover the gold from the gossan. The Mammoth Copper Mining Co., a subsidiary of the United States Smelting Refining and Mining Co. acquired the property in 1904 and began large-scale mining operations in 1905. The mine was operated continuously from 1905 to 1919 but was closed from 1919 to 1923. Operations were resumed late in 1923 but were again suspended in 1925. No ore has been mined since 1925, although some exploratory work has been done. Plate 14 shows the extent of the mine workings.

Table 13 gives the production data of copper and zinc ore from the Mammoth mine, 1905–25. The gross value of the recovered gold, silver, and copper was $51,970,290 and the value of the recovered zinc was $4,525,870.

GEOLOGY OF THE MINE AREA

FORMATIONS

Only the Balaklala rhyolite and the Kennett formation crop out in the vicinity of the Mammoth mine (pl. 15). A small remnant of the lower part of the Kennett formation consisting of interbedded gray shale and water-laid rhyolitic tuff, and arkose is exposed in the southeast corner of the mapped area. The Kennett formation lies conformably on the Balaklala rhyolite and dips gently east.

TABLE 13.—*Annual production and grade of ore from the Mammoth mine, 1905–25*

[Data furnished by the United States Smelting Refining and Mining Co.]

Year	Production (*short* tons)	Grade: Gold (*ounces* per ton)	Silver (*ounces* per ton)	Copper (percent)	Zinc (percent)	Insoluble (percent)	Iron (percent)	Sulfur (percent)
Copper ore								
1905	13, 868	0. 029	1. 94	4. 30				
1906	181, 733	. 048	2. 12	3. 95	3. 30	9. 5	39. 0	42. 2
1907	201, 124	. 039	1. 94	4. 15	4. 70	7. 7	38. 4	44. 5
1908	311, 997	. 033	1. 90	4. 03	4. 70	12. 6	35. 6	41. 9
1909	370, 985	. 036	2. 13	4. 15	4. 70	13. 2	34. 9	42. 0
1910	315, 189	. 038	2. 38	3. 83	4. 40	15. 1	34. 2	40. 7
1911	269, 438	. 032	2. 26	4. 06	4. 20	12. 4	35. 4	44. 7
1912	273, 555	. 034	2. 24	3. 47	4. 20	12. 6	34. 0	40. 9
1913	269, 233	. 036	2. 19	3. 58	3. 80	14. 9	32. 7	39. 8
1914	227, 459	. 043	2. 28	3. 96	4. 10	12. 5	33. 9	40. 3
1915	242, 573	. 038	2. 21	3. 75	3. 80	12. 8	33. 3	40. 0
1916	223, 458	. 042	2. 74	4. 77	4. 20	13. 8	32. 6	37. 6
1917	147, 340	. 040	2. 46	4. 22	3. 80	20. 4	31. 4	35. 6
1918	76, 079	. 040	2. 29	3. 91	4. 30	14. 2	32. 4	37. 6
1919	37, 733	. 040	2. 50	4. 10	3. 70	14. 4	32. 9	37. 8
1923	9, 221	. 050	2. 85	4. 42	5. 00	16. 6	33. 2	36. 9
1924	115, 814	. 045	2. 39	4. 27	3. 70	13. 5	35. 2	40. 2
1925	34, 346	. 037	2. 17	4. 34	4. 30	11. 0	36. 6	41. 4
Total and weighted averages	3, 311, 145	. 038	2. 24	3. 99	4. 20	13. 3	34. 3	40. 4
Zinc ore								
1914–15	84, 000	0. 078	5. 79	2. 40	21. 10			
Total and weighted averages	3, 395, 145	. 039	2. 32	3. 95	4. 62			

The Balaklala rhyolite in the mine area consists of porphyritic and nonporphyritic rhyolitic flows and intercalated coarse and fine rhyolitic pyroclastic material. Parts of the upper, middle, and lower units of the Balaklala are exposed; the ore zone is in the uppermost part of the middle unit, immediately under the base of the upper unit.

The lower unit of the Balaklala in the mine area consists mainly of light-gray to light-green nonporphyritic rhyolite, and rhyolitic tuff and volcanic breccia, but locally it contains a few porphyritic flows. Most of the lower unit is exposed in the deep canyon of Little Backbone Creek north of the mine where it is highly pyritized at many localities.

The middle unit of the Balaklala in the mine area ranges in thickness from 150 to about 300 feet. It consists mainly of light-green to light-gray porphyritic rhyolitic flows and pyroclastic rocks containing 1- to 4-millimeter phenocrysts of quartz and feldspar. The upper part of the middle unit contains abundant, although discontinuous, coarse and fine pyroclastic material and some water-deposited tuff beds.

The upper unit of the Balaklala in the mine area consists of coarsely porphyritic rhyolite containing quartz and feldspar phenocrysts, some of which are more than 4 millimeters in diameter. It is about 1,400 feet thick at Mammoth Butte, 1 mile west of the Mammoth mine, but thins rapidly toward the east over the mine area. Part of the upper unit has been removed by erosion except in the southeast corner of the mapped area (pl. 15). A thin bed of tuff or volcanic breccia is present at many localities at the base of the upper unit. This pyroclastic bed is generally 10 to 50 feet thick, but where it is composed mainly of volcanic breccia, it is as much as 150 feet thick. The upper unit of coarse-phenocryst rhyolite and the tuffaceous bed at the base of this unit form the "cap rock" for the ore deposits, which were deposited just below the base of the "cap rock."

FOLDS

The bedding in the mine area is delineated by the contact between the overlying coarse-phenocryst rhyolite and the underlying medium-phenocryst rhyolite and nonporphyritic rhyolite, or by bedded tuff at the base of the coarse-phenocryst rhyolite.

The structure contour map (pl. 16) shows that a slightly elongate arch trends N. 45° E. and has a culmination in the central part of the group of ore bodies. There are also many smaller arches or elongated topographic "highs," the alinement of which appears to have no relation to the alinement of the main arch.

The sedimentary material at the base of the coarse-phenocryst rhyolite delineates the folded structures. Bedded material is shown at the base of the coarse-phenocryst rhyolite on the underground maps furnished by the United States Smelting Refining and Mining Co. Evidence from correlation with the surface mapping done by the writers and from thin sections, indicates this material is a tuff bed. Dips and strikes in the bedded material on underground maps show that in most places the bedding is parallel to the contact of the coarse-phenocryst rhyolite (pl. 15). Some random orientation of bedding has been observed, but such variations may be caused by close folding, faulting, or minor intrusions of coarse-phenocryst rhyolite.

At the Mammoth mine the degree and type of foliation that is formed is dependent on the competence of the rock units. The nonporphyritic rhyolite, the medium-phenocryst rhyolite, and the thick, massive coarse-phenocryst rhyolite are competent rocks. Pyroclastic beds at the base of the coarse-phenocryst rhyolite form an incompetent layer that covers a considerable area. During the folding in the mine area steep fracture cleavage—with some recrystallization and alinement of mineral grains parallel to cleavage planes—formed in the nonporphyritic and medium-phenocryst rhyolite, particularly near the axes of folds. Movement along the bedding planes, resulting in a secondary foliation parallel to the bedding planes, was concentrated in the pyroclastic layers at the base of the coarse-phenocryst rhyolite. In the overlying massive coarse-phenocryst rhyolite steep fracture cleavage is locally present, but it is rare because of the great thickness and competence of this rock.

FAULTS

Several large fault zones and many small faults are shown on the maps of the mine levels. They are of several ages, as shown by one fault cutting another; most are probably postmineral in age, but some may be premineral. The principal faults are the California, the 313 (and its possible extension, the Yolo), the 12-drift, the Schoolhouse, the Gossan, the Friday, and the Clark. On all these faults, the north or east side has moved down relative to the south or west. The horizontal component is not known except on the California fault, where it is about 250 feet.

Few of the faults can be located or traced for any distance on the surface. Their location as shown on plate 15 is based largely on a projection of the faults from their known position underground. Except for short sections of the California, Schoolhouse, and Gossan faults, where these are exposed at the surface, information on the faults is taken from the underground maps of R. N. Hunt, from R. T. Walker (oral communication), and from the writers' brief underground examination of the California fault.

The California fault zone is a group of subparallel and branching fault planes rather than a single plane, and, as seen underground, the zone does not have sharp boundaries at all localities. At some exposures the rock between the individual fault planes is massive and little altered; at others the fault consists of a zone of gouge and crushed quartz-sericite rock containing many anastomosing shear planes. Hydrothermally altered rock is not prominent in the exposures seen underground although some sericite is present, but hydrothermally altered rock is recognized along the surface exposure of this fault, as weathering has emphasized the difference between altered and unaltered rock. The porphyritic rhyolite along the California fault at the surface is altered to white and pink clay minerals and sericite, but this alteration extends only 5 to 10 feet on either side of the fault.

The California is probably a premineral fault, but postmineral movement has also occurred as shown by offset ore bodies. The north block moved east and down, but the exact amount of movement is not known. The ore bodies plunge at a low angle to the southwest along the fault, so that the present position of offset segments of ore bodies could be due either to mainly dip-slip or strike-slip movement.

The 12-drift fault and the 313 fault, and its possible extension, the Yolo, are shown on the maps of the mine levels (pl. 17) and on the cross sections (pl. 15). The vertical offset on each of these faults is about 40 feet, and the rocks on the north side moved down relative to the south, but nothing is known about a possible horizontal offset. They are probably older than the California fault as the underground maps suggest that they are offset by it.

The Schoolhouse fault is exposed only on the 670-foot level (pl. 17) and for a short distance on the surface. The approximate vertical offset of the base of the coarse-phenocryst rhyolite at the surface appears to be about 100 feet, where the rocks on the north side moved down relative to the south, although exposures are poor where the fault cuts this contact. The offset on the Schoolhouse fault appears to be less in the east end of the mapped area than in the west. The rhyolite along the Schoolhouse fault at the surface is pyritized and altered to clay minerals.

The Gossan fault is exposed on the 670-foot level for a short distance on the surface. The rocks on the east side of the fault are dropped relative to those on the west; on the basis of a doubtful correlation of rock types the vertical offset is about 50 feet. The rhyolite is pyritized and altered to clay minerals along the outcrop of this fault.

Several north-dipping faults that strike N. 60°–80° W. are in the vicinity of the Friday Lowden and Hanley ore bodies at the southwest end of the mine. The Friday fault is one of this group of faults that cut the Friday Lowden ore body and may offset it. The total offset on this group of faults may be considerable, but no data are available.

The Clark fault, which is a minor fault, has an offset of 10–20 feet. The rocks on the north side have moved down relative to the south (pl. 15).

ORE BODIES

CHARACTER AND DISTRIBUTION

The ore bodies of the Mammoth mine are large, flat-lying, tabular bodies of copper- and zinc-bearing massive pyritic ore which extend along a horizontal distance of 4,200 feet. The northeast end of the ore zone has been eroded, and exploration has not delimited the southwesterly extension; the central part has a width of 1,000 feet. Although minable ore bodies occur throughout the ore zone, it is not all ore (pl. 14). Individual stopes reach a maximum horizontal dimension of 900 by 500 feet and the maximum thickness of ore is 110 feet; ore bodies range from these maxima down to small stopes from which only a few hundreds of tons were mined. The ore zone lies along the crest of a broad arch in the rocks, and all known ore occurs at or a short distance below the contact between the coarse-phenocryst rhyolite and the underlying rhyolitic tuff and flows (pl. 15). The ore formed by replacement of rocks in the uppermost part of the middle unit of the Balaklala rhyolite.

Table 14 gives data on the production and grade of stopes and other ore blocks; it also shows the uniformity of the gold, silver, and copper content of all the copper ore bodies, and the sharp distinction between high-grade copper and high-grade zinc ore. The pyritic ore that underlies the copper ore bodies contains a smaller quantity of copper than the massive sulfide that was mined for copper, but the ore is the same type.

Table 15 gives information on the proportion of metals in the zinc-rich parts of the ore bodies. The grade of the material sorted from the ore of high zinc content was calculated from the known assays. It is apparent, for instance, that the discarded material sorted from the zinc ore was rock that contained some copper, gold, silver, zinc, and iron but contained more rock material by volume than is contained in massive sulfide (45.3 percent by weight insoluble). The figures suggest that the discarded material consisted of rock containing stringers of the massive sulfide type of ore in essentially unmineralized rock, rather than that the discarded material contained disseminated minerals.

TABLE 14.—*Production and grade of stopes and other ore blocks of the Mammoth mine*

[Data furnished by the United States Smelting Refining and Mining Co.]

Ore body	Production (short tons)	Grade							
		Gold (ounces per ton)	Silver (ounces per ton)	Copper (percent)	Zinc (percent)	Lead (percent)	Iron (percent)	Sulfur (percent)	Insoluble (percent)
Copper ore									
Gossan (oxidized)	10,000	0.10	15.0	0.5					
Gossan (sulfide)	65,000	.04	7.0	10.0					
Copper Crest	15,000	.04	2.0	4.0					
Main	1,970,145	.03	1.8	3.8					
Winslow	300,000	.04	2.2	3.8					
Graton	2,000	.03	2.0	3.0					
Sierra	100,000	.04	2.2	3.8					
Smith winze	2,000	.03	2.0	3.0					
Clark	385,000	.04	3.0	4.5					
Metcalf	200,000	.04	3.0	4.5					
Starrow	100,000	.04	3.0	4.5					
2784	2,000	.03	2.0	3.5					
Friday Lowden	100,000	.04	3.0	5.0					
Hanley	60,000	.04	3.0	6.0					
Zinc ore									
		Grade of sorted ore [1]							
Yolo	58,000	0.12	5.44	2.20	36.20	3.00	10.1	25.2	20.7
Copper Crest									
473	16,000	.02	1.40	1.80	33.30		9.2	24.7	19.8
313	5,000	.13	6.50	1.00	27.70	1.00	9.5	23.5	31.8
Graton	2,500	.40	4.90	.72	36.35	1.40			
Metcalf									
		Grade of unsorted ore [1]							
Total [1]	84,000	.078	5.79	2.40	21.10		12.6		35.5
Pyritic ore underlying copper ore bodies									
Main (separate blocks)		0.04	1.0	2.0			42.0	45.0	10.0
Do		.03	.4	.5			38.0	40.0	18.0
Do		.03	.2	.5			43.0	45.0	10.0
Shasta		.01	.6	.5				35.0	
Clark		.03	1.0	1.2				40.0	
Friday Lowden (separate blocks)		.04	1.1	2.2				45.0	
Do		.01	.3	.5				40.0	
313, at 390-foot level		.05	1.1	2.0				40.0	
Total		.028	.56	.9				41.4	

[1] See table 15.

Little information is available to the writers on the appearance of the copper-zinc ore underground, or on the nature of the contact between ore and wall rock. Specimens that the writers collected from dumps and along the tramline, and analyses of ore showing the higher content of insoluble material, indicate that although most of the ore was a massive pyrite; some contained more quartz and altered, unreplaced rock material than the massive pyritic ore of the Iron Mountain and Shasta King mines. However, observers who have seen the ore underground have noted that quartz is rarely visible megascopically in the Mammoth mine ore except where a few late veinlets cut both the ore and the wall rocks.

Frozen contacts between ore and wall rocks are reported to be rare; at most contacts, as in the other mines in the district, ore is separated from rhyolite by a clay gouge selvage. R. N. Hunt (oral communication) states that many ore bodies had sharp, smooth outlines. In these the contacts between the massive sulfide and wall rock were sharp; the ore was separated from the wall rock by a thin selvage of gouge at the ends as well as along the top and bottom of the ore bodies, and the wall rock contains little or no pyrite. Contacts of this

TABLE 15.—*Production and grade of hand-sorted zinc ore from the Mammoth mine* [1]

[Data furnished by the United States Smelting Refining, and Mining Co.]

Ore	Production (short tons)	Grade					
		Gold (ounces per ton)	Silver (ounces per ton)	Copper (percent)	Zinc (percent)	Iron (percent)	Insoluble (percent)
Crude ore produced	84,000	0.078	5.79	2.40	21.10	12.6	35.5
Sorted ore shipped	28,800	.111	9.51	2.4	39.6	8.9	16.8
Percent increase or decrease	−65.7	+42.3	+63.6	0	+87.5	−29.4	−52.6
Discard calculated	55,200	.061	3.82	2.4	11.42	14.6	45.3

[1] See table 14.

type are similar to those at Iron Mountain and at the Shasta King mine. R. T. Walker (oral communication), however, reports that the material between some of the stopes (and sometimes beneath them) was massive sulfide containing less copper than the ore that was mined. The ore zone was continuous between these stopes, and some stope outlines mark only the economic limit of mining. Thus, although many stopes were mined to a sharp waste wall at the ends of the ore bodies and on all sides, at some places an individual stope does not represent the extent or continuity of a body of massive sulfide, which may be of considerably greater extent than is indicated by the outline of the stope.

The ore zone at the Mammoth mine extends S. 70° W. for 3,000 feet from the outcrop of the Main ore body to the deepest exploratory workings southwest of the Friday Lowden. The general trend of individual ore bodies—and of the zone as a whole, if the Gossan ore body is included—is more nearly S. 60° W. Part of the zone is eroded between the Gossan and Main ore bodies (pl. 15). Although the Gossan ore body lies at a lower stratigraphic horizon than the main part of the mineralized zone, it seems reasonable to assume that ore bodies were present along the eroded upper part of the zone and its eroded extension northwest of the Gossan ore body. The zone must have been at least 4,200 feet long, and may have been considerably longer.

The plunge of the ore zone ranges from almost horizontal to vertical, but averages only 14° from the outcrop of the Main ore body to the deepest known ore.

A few large and many small faults cut the ore bodies. Small faults are marked by slickensided surfaces. The larger faults appear to have offset ore bodies several hundred feet, and the present distribution of these ore bodies is due in part to postmineral faulting.

OXIDATION AND ENRICHMENT

The gossans at the outcrops of the ore bodies at the Mammoth mine are not large or conspicuous, considering the size and extent of the ore bodies. Gossans about 10 feet thick occur where the tips of the Main and Winslow ore bodies crop out, but most of the gossan is thinner. Rock that contains disseminated pyrite is much more extensive at the surface than massive sulfide ore, and upon oxidation of the pyrite forms considerable areas of rusty rhyolite (pl. 15).

Only a few small nodules of massive sulfide ore were found in the outcrop of the Main ore body, and R. N. Hunt and R. T. Walker (oral communication) state that oxidation extended as much as 150 feet in from the present erosion surface.

Supergene enrichment has not taken place on a large scale at the Mammoth mine, although some secondary enrichment has occurred as a result of oxidation of the ore near the outcrop. The Gossan, and possibly the Main and Winslow ore bodies were partly oxidized and enriched, but figures on the metal content of the oxidized and enriched ore are available only for the Gossan ore body. Specimens of enriched ore were not available to the writers, and the mineralogy of the enriched ore is not known. Seager in his unpublished report noted that chalcocite and covellite are present in specimens of enriched ore from the Mammoth mine.

Production data on the Gossan ore body are given in table 14. The enrichment indicated for it is about the same as that in the Old Mine ore body at the Iron Mountain mine (Kinkel and Albers, 1951, p. 18). The gold content of the enriched oxidized ore is about twice that of the primary ore, owing to loss of weight in the formation of gossan from massive sulfide.

HYDROTHERMAL ALTERATION

Alteration ranges in the mine area from unaltered porphyritic rhyolite to altered rock having relict quartz phenocrysts in a matrix of secondary minerals. The rocks have been altered by regional metamorphism, hydrothermal alteration, weathering, or by a combination of these processes. The altered porphyritic rhyolite is white, lavender, buff, or light green. Hydrothermal origin is evidenced by the conspicuous white alteration of the rhyolite around ore bodies and along fractures in the ore zone; the intensely altered rock is composed of quartz and sericite, and crumbles apart in the hand.

ORE CONTROLS

The four controls that are thought to be of major importance in localizing the ore at the Mammoth mine are: (1) stratigraphic control, (2) structural control by foliation, (3) structural control by the main arch and minor flexures, and (4) location of feeder fissures.

The ore zone is restricted to the rocks that immediately underlie the coarse-phenocryst rhyolite, as shown in the cross sections (pl. 15). This zone contains much bedded pyroclastic material, principally discontinuous tuff beds and lenses of volcanic breccia, but it also contains flows of medium-phenocryst rhyolite. Mapping done by the writers at the Iron Mountain and Shasta King mines shows that the massive sulfide ore preferentially replaced porphyritic rhyolite with 2- to 3-millimeter quartz phenocrysts. Although no direct evidence is available on the nature of the rock that was replaced by the ore bodies of the Mammoth mine, the medium-phenocryst rhyolite in the ore zone at the Mammoth mine suggests that this may be the host rock here also.

The rocks are not strongly foliated in the Mammoth area, probably because the folds are broad, but two types of foliation that have had a considerable effect on ground preparation before ore deposition are evident at the Mammoth mine. The first type is a foliation parallel to bedding planes that is localized along flow contacts, particularly where there is much pyroclastic material between adjacent flows. The second type is a steep fracture cleavage at a high angle to the bedding that forms most strongly near the axes of folds. The foliation ranges from poorly defined sheeting without noticeable reorientation of minerals, through well-formed sheeting with sericite oriented along discrete planes, to a sericite schist in which only lenticles of unsheared rock remain.

The intersection of steep fracture cleavage with foliation parallel to the bedding plane is an important ore control at the Mammoth mine because it formed a zone of fractured rock along the crest of an arch under a relatively impervious cover of unfractured rock.

Most of the ore bodies in the Mammoth mine lie along the culmination of an arch in the rhyolite. This is shown on the structure contour map (pl. 16) and on the level maps. In addition to the main arch extending from the outcrop to the Friday Lowden and Hanley ore bodies at the southwest end of the mine, several smaller arches that appear to localize individual ore bodies are present. Broad minor arches extend over the Clark body, the northwestern extension of the Main, the 473, and the 313 ore bodies. The Copper Crest appears to lie on the northward-plunging nose of a major arch. The ore zone at the culmination of this arch has been eroded.

Faults or shear zones that acted as feeder channels are difficult to recognize with certainty, but faults that appear to have a control on mineral deposition or rock alteration may tentatively be regarded as feeder channels for solutions. The faults along which claylike hydrothermal alteration occurs are the California, the Schoolhouse, the Gossan (these last two are exposed on the surface for a short distance only), and the unnamed fault that lies about 200 feet south of the Copper Crest ore body. All these except the Gossan fault dip steeply and trend northeast.

The California fault cuts the trend of the ore zone at a low angle, and lies 400 feet northwest of the ore bodies at the southwest end of the known ore zone. The localization of ore near the California fault suggests it was a feeder for the whole ore zone, and that the ore solutions left it at a point below the Friday Lowden ore body and traveled upward along the main arch structure rather than along the fault. However, the presence of several small ore bodies on the 200-foot level (pl. 17) that are apparently alined along the California fault indicates either that some solutions continued up along the fault, or that solutions traveled along the arch and deposited ore against the fault where it cuts the arch.

The alinement of the zinc-rich ore bodies (473, 313, Graton, Yolo, and possibly the Copper Crest) along the 313 fault and its probable extension, the Yolo, is conspicuous and suggests that the introduction of zinc into the ore was primarly along the 313 fault. This fault may have reopened slightly after the main period of copper deposition.

The conjunction of the major ore controls is not mere coincidence. Lenses of pyroclastic material initiated local folding because, in contrast to massive flows, they form a less competent layer and because a change in dip, along the boundaries of a lens in this instance, is a favored locus for the start of a fold. Movements along the bedding plane were most pronounced in folded areas, and fracture cleavage formed mainly in the axial regions of folds. The relationship, if any, between the location of feeder fissures and the other ore controls is not known.

SHASTA KING MINE

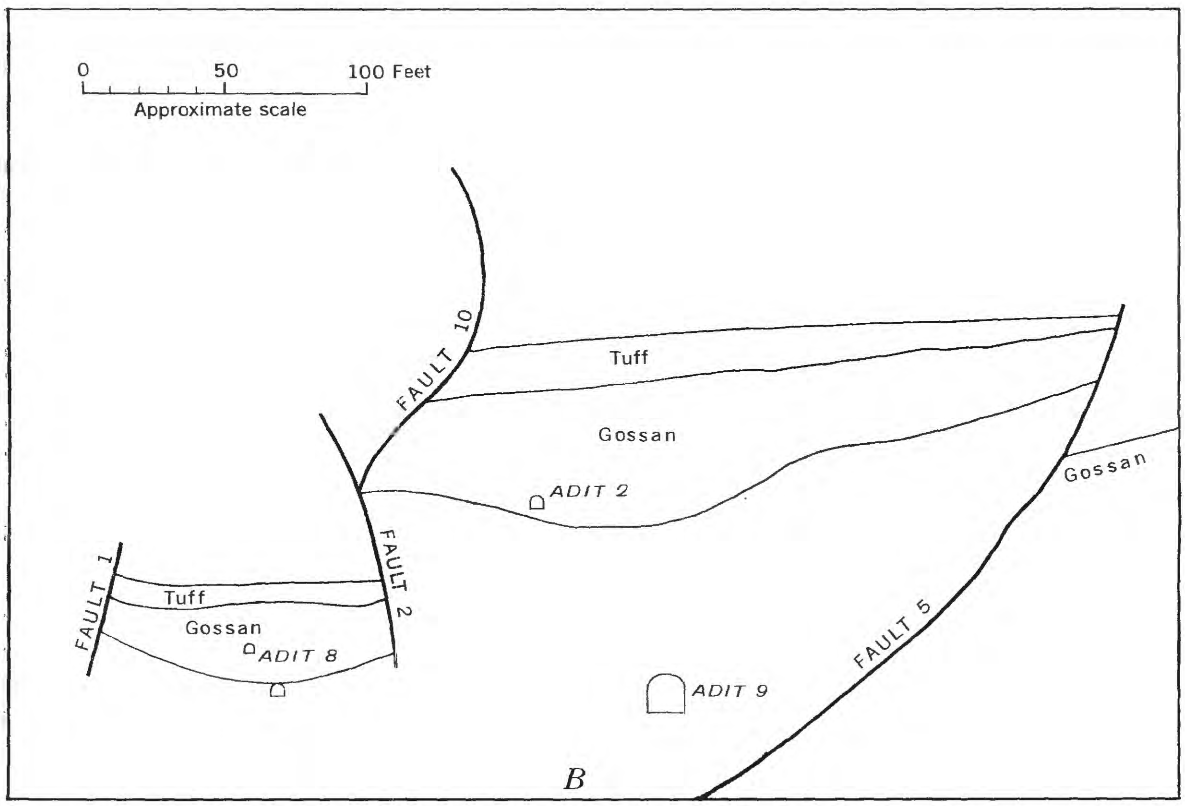
The Shasta King mine is in the central part of the West Shasta copper-zinc district near the bottom of a narrow canyon of Squaw Creek in sec. 12, T. 33 N., R. 6 W., about 19 miles by road from Redding. Access to the mine by automobile was not possible, however, during 1949 because of washouts near the mine. By rebuilding about 1½ miles of road it would be possible to drive from Shasta Dam to the mine. A hard-surfaced road extends from Redding to Shasta Dam.

The topography in the vicinity of the mine is rugged. In the canyon of Squaw Creek few slopes are less than 35° and slopes of 50° are common. The flat-lying ore body of the Shasta King mine crops out about 150 feet above Squaw Creek and has been explored by adits in the canyon wall (fig. 59).

The Shasta King mine was operated from 1902 to 1909 by the Trinity Copper Corp., and the ore was shipped to the Balaklala smelter at Coram, Calif. The property was idle from 1909 to 1917, when it was leased to the United States Smelting Refining and Mining Co., who operated the mine from 1918 to 1919. Operations were stopped at the mine in March 1919, and a fire destroyed the mine camp, Boralma, in 1924, but most of the mine workings were open and accessible in 1949. The mine was purchased by the present (1952) owners, R. T. Walker and W. J. Walker, in 1944.

Data on the amount of copper produced from the Shasta King mine are incomplete. The production figures tabulated in table 16 were furnished by R. T. Walker and W. J. Walker.

A



B

FIGURE 59.—Shasta King mine. *A*, View from the south. *B*, Diagram is sketch drawn from photograph.

The ore contains substantial amounts of zinc, but as it was not recovered in the smelters, no estimate of the content of the mined ore can be made. A probable copper-zinc ratio of 1 : 3 is indicated by assays furnished by Walker and Walker of stope and pillar samples taken in 1948. These assays averaged 2.5 percent copper and 7.61 percent zinc.

TABLE 16.—*Production and grade of ore from the Shasta King mine*

Operator	Production (short tons)	Grade		
		Gold (ounces per ton)	Silver (ounces per ton)	Copper (percent)
Trinity Copper Corp., 1908–09	15, 000	(1)	(1)	(1)
United States Smelting and Refining and Mining Co., 1918–19	68, 889	0. 034	1. 01	2. 92

[1] Unknown.

BALAKLALA RHYOLITE

The ore body of the Shasta King mine has replaced the Balaklala rhyolite; no other formations are exposed in the vicinity of the mine (pl. 18). The rocks adjoining the ore are porphyritic and nonporphyritic varieties of the Balaklala rhyolite and well-bedded rhyolitic tuff and volcanic breccia. The nonporphyritic rhyolite is locally flow banded. Although the nonporphyritic rhyolite contains a few quartz and feldspar phenocrysts less than 1 millimeter in diameter, it is mainly nonporphyritic, which distinguishes it from other flows.

The porphyritic units of the Balaklala rhyolite of the Shasta King mine are subdivided into two principal rock types: (1) medium-phenocryst rhyolite containing quartz phenocrysts 1 to 4 millimeters in diameter; (2) coarse-phenocryst rhyolite containing phenocrysts more than 4 millimeters in diameter. The first type comprises several individual flows, including flow-banded porphyritic rhyolite, rhyolite with abundant small feldspar phenocrysts, and a dark-purple porphyritic rhyolite; it was not possible to map these varieties separately because the underground exposures were inadequate. The second type, the porphyritic rhyolite containing large quartz phenocrysts, appears to have intruded the rhyolitic flows in the mine area.

A bed of tuff and volcanic breccia that lies immediately above the gossan at the surface is also found in the back of the stopes in the southwestern part of the mine. The bed is composed of shaly rhyolitic tuff that is interlayered with fine and coarse volcanic breccia. The tuff is variable in texture along the strike of the bed where it is exposed at the surface. The southwestern part is composed principally of layered tuff; the central part contains mixed shaly tuff and volcanic breccia; and the northeastern part consists almost entirely of coarse volcanic breccia that is underlain by a thin layer of tuff. Columnar sections of the bed at the surface are shown in figure 60.

ORE BODY

The Shasta King ore body is a lenticular body of massive sulfide that contains, in addition to pyrite, copper and zinc minerals and small amounts of gold and silver. The ore body crops out along the steep north slope of the canyon of Squaw Creek and has been partly removed by erosion. The remnant of the ore body has the shape of an elongated shallow basin that trends northeastward. The outcrop of the ore body has a length of 590 feet and a maximum thickness of 42 feet. Underground work has shown its width to be at least 500 feet. Gossan

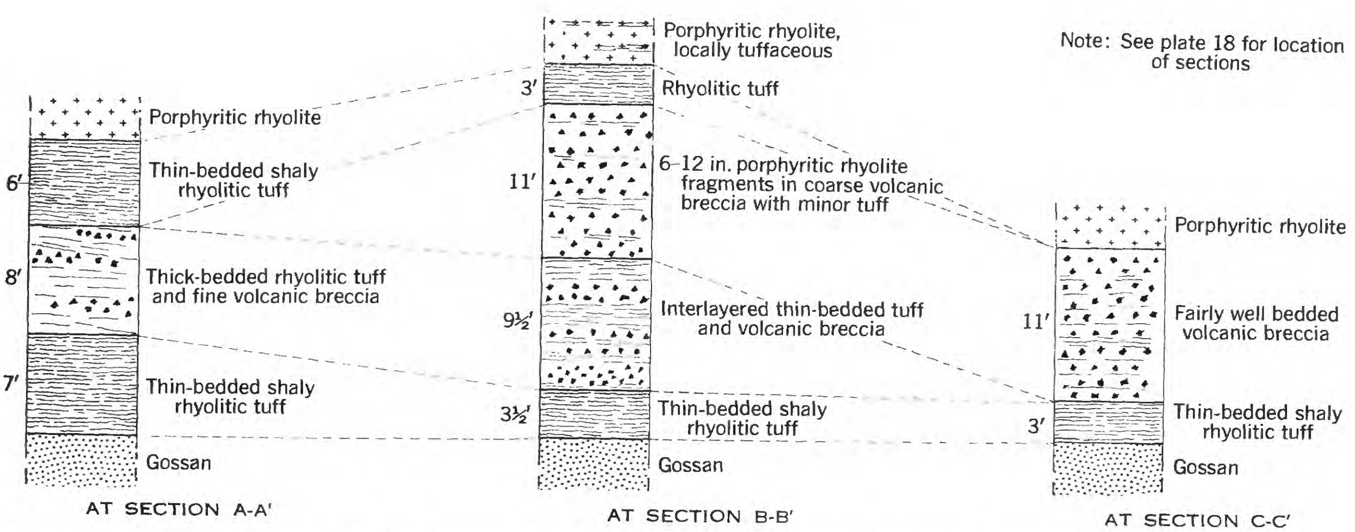


FIGURE 60.—Columnar sections of tuff and volcanic breccia above the gossan, Shasta King mine. See plate 18 for location of sections.

crops out on the opposite side of the canyon southwest of the mapped area, which suggests that the ore body was much larger before erosion. The ore is thickest where it is exposed at the present erosion surface. In places it thins to a few feet toward the northwest, but exploration has not delimited the northern boundary.

The ore body is explored by adits, and most of the mine workings were open and accessible in 1949. The location and geology of the underground workings are shown on plate 19.

The Shasta King ore body is continuous through the mine (pl. 19). The ore is uniform and structureless in appearance, and is composed principally of massive pyrite and lesser amounts of chalcopyrite and sphalerite. The pyrite is anhedral and commonly fine grained, the grains average 1 millimeter in diameter. Megascopic chalcopyrite and sphalerite in small irregular masses can be seen in the massive sulfide ore, but no megascopic veinlets of chalcopyrite or sphalerite were found. The ore contains very little gangue except near the margins of the ore body. A few nodules of porphyritic rhyolite containing 2-millimeter quartz phenocrysts, which are unreplaced remnants of the host rock, occur in the central part of the massive sulfide ore. Such nodules are common at the upper contact of the ore body at several places in the mine. The nodules range in diameter from less than an inch to several feet. Some are rounded and have sharp contacts with sulfide ore; others are irregular and the boundaries are gradational from barren rock, through pyritized rock, to massive sulfide.

The contact between massive sulfide ore and the wall rock is sharp at some places and gradational at others. Where the contacts are sharp, the massive sulfide ore ends abruptly against an unmineralized white clay and sericite schist that constitutes a strong gouge. This type of contact is exposed in the backs of some of the stopes on the 830-foot level, where the ore ends against fault 1, and on the 910-foot level along faults 10 and 11. The gouge and sericite schist are a foot thick in only a few places; they contain no crushed sulfides, and the porphyritic rhyolite outside the gouge zone is unsheared at most localities. Other contacts between ore and wall rock show a gradation from massive sulfide to unmineralized rock, and bodies of partly replaced porphyritic rhyolite remain in the ore. Soft, sandy- or sugary-looking sulfides occur near some edges of the ore body. At these places the ore ranges from 30 to 90 percent pyrite in schistose sericitic rock, and minable ore is determined by assays. Relict quartz phenocrysts that average about 2 millimeters in diameter remain in the partly replaced rock.

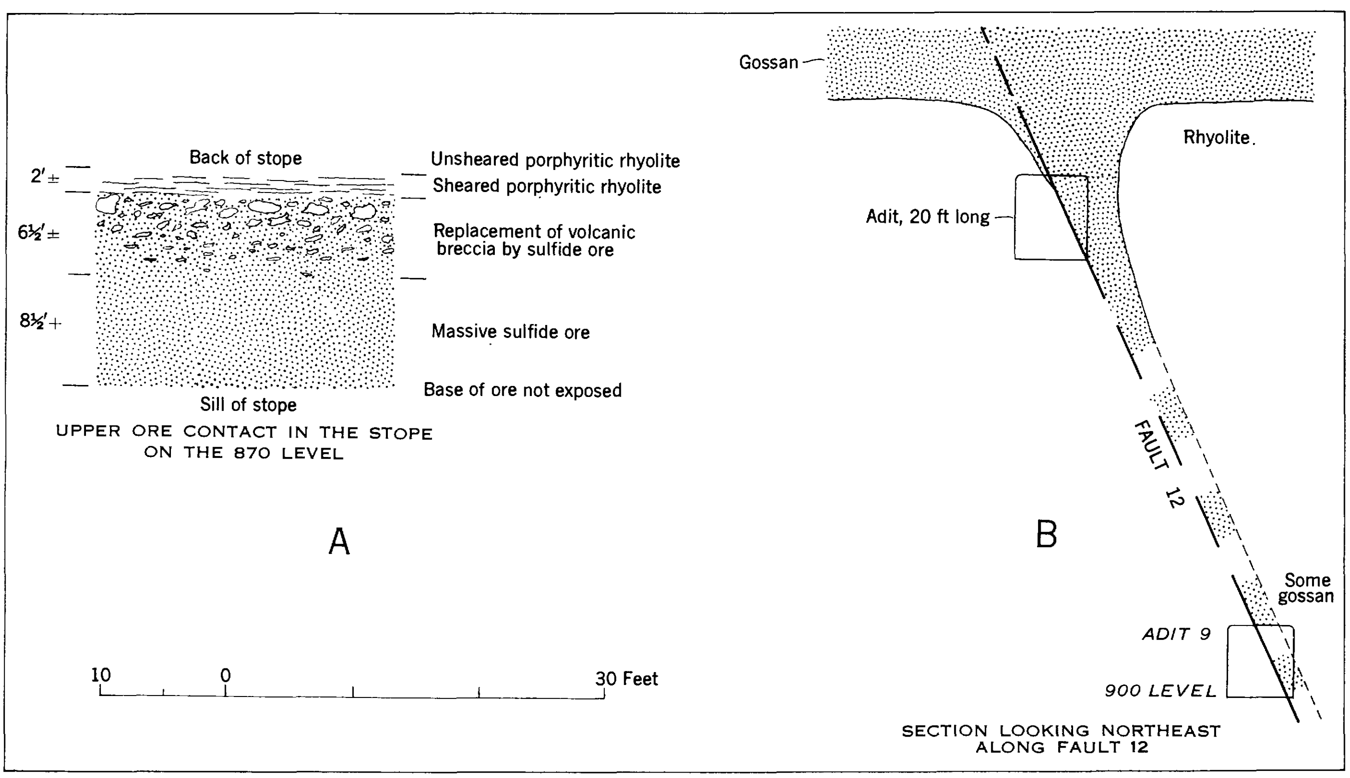


FIGURE 61.—Details of sulfide contacts, Shasta King mine. *A*, Upper ore contact in the stope on the 870-foot level near cross section *F–F'*; *B*, section looking northeast along fault 12.

The rock near the ore contact is in many places hydrothermally altered and schistose. Consequently it is difficult to differentiate the porphyritic and nonporphyritic types of rhyolite, but the porphyritic types can in places be distinguished by the relict quartz phenocrysts. They are present in the less altered facies of the rock above the ore, in nodules of waste, and at a few localities below the ore. Information on the types of rock surrounding the ore body is obtainable only from a few drifts and a few exposures in stopes at the edges of the ore body, or from pieces of rock that have fallen from the backs of stopes. The base of the ore is not exposed in most of the stopes. Although the ore is generally underlain by nonporphyritic rhyolite, it is underlain in a few places by porphyritic rhyolite that was not completely replaced by the ore to the nonporphyritic rhyolite contact. The ore at the west end of adit 6 is underlain by chloritic rock. This rock is a chloritized facies of the nonporphyritic rhyolite. It is not a mafic flow interlayered with the Balaklala rhyolite because it contains a few 1-millimeter quartz phenocrysts and resembles chloritized rhyolite found at a few other places in the district.

The principal ore minerals are pyrite, chalcopyrite, sphalerite, and small amounts of galena, and tetrahedrite. Small amounts of gold and silver were recovered from the ore, although no free gold was seen. Silver is present in ore without tetrahedrite.

Sulfide minerals constitute 85–90 percent of the ore body. The gangue consists of unreplaced nodules of porphyritic rhyolite, unreplaced quartz phenocrysts, sericite, and introduced quartz. Quartz and sericite are present throughout the ore body, whereas the nodules of porphyritic rhyolite are concentrated near the borders.

FAULTS

Faults, which are both premineral and postmineral in age, are conspicuous in surface and underground exposures. Only the most important faults have been given numbers on the geologic maps and sections, but many other faults undoubtedly exist outside the explored area.

The premineral faults are nos. 3, 5, and 12 (pl. 19). The evidence for a premineral age is the presence of pyrite, and pyrite casts filled with iron oxide, along the faults away from the ore bodies, and hydrothermal clay minerals along the faults. Pyrite occurs as much as 75 feet vertically above the gossan on the fault 3. In addition, a small fault in adit 6 (870-foot level) 2,340 feet east on the coordinate system, contains small lenses of massive sulfide ore that apparently formed in place along the fault. Hydrothermal alteration was observed along fault 6; this fault may be premineral in age but may have moved again after mineralization. Faults 3 and 5 also moved both before and after the sulfides were deposited.

All the faults except 12 have some postmineral movement. The direction of displacement on fault 1 was determined from the position of mafic flows that crop out to the west (pl. 18). Geologic evidence outside of the mine area indicates that these mafic flows are lower in the stratigraphic sequence than the rocks exposed to the east. The vertical offset on the fault is several hundred feet or more. The direction of offset on fault 8 is not known with certainty, but the northeast side is probably upthrown relative to the southwest. The direction and amount of postmineral dip-slip movement on faults 2, 5, 6, and 7 are shown by the offset of the gossan at the surface. Faults 2, 5, and 7 must have moved horizontally as well as vertically, as the offset of the gossan at the surface, where the dip is steeper, is greater than the offset of the ore underground, where the dips are at low angles. In addition, the thickness of the gossan is not the same on opposite sides of these faults at the surface. The ore body thins toward the northwest, and horizontal movement along northwestward-trending faults would result in differences in thickness of the ore on opposite sides of the faults.

ORE CONTROLS

The Shasta King ore body formed by the partial to complete replacement of a thin flow of porphyritic rhyolite that lies between a bed of pyroclastic rock above and a flow of nonporphyritic rhyolite below. The ore body is conformable with the contacts of the flow, but it does not everywhere completely replace the flow of porphyritic rhyolite. The porphyritic rhyolite that is the host rock of the ore body contains 2- to 3-millimeter quartz phenocrysts that are distributed through a very fine grained siliceous groundmass. Smaller feldspar phenocrysts are present but are not prominent. The unreplaced remnants of the porphyritic rhyolite in the ore are generally massive and unsheared. Many fragments of the porphyritic rhyolite remain as unreplaced or partly replaced remnants in the massive sulfide ore, and all gradations are present between unreplaced porphyritic rhyolite and massive sulfide ore. Although assays are not available, visual inspection shows that the only sulfide in the transition zones between ore and waste is pyrite, and that copper and zinc minerals are limited to massive sulfide ore.

The ore contacts are sharp against clay and sericite gouge at some places, but are not as sharply defined between massive sulfide ore and hydrothermally altered wall rock. Where a transition zone is present between ore and wall rock, pyrite replaces preferentially the schistose part of the rock. Within the transition zones, anastomosing bands of foliated sericite cut the massive porphyritic rhyolite, but lenticular bodies of unsheared rock a few inches to a few feet in length remain. During the period of ore formation pyrite appears to have completely replaced the foliated (sheared) parts of the porphyritic rhyolite before it replaced the massive nodules. The porphyritic rhyolite above the ore in the central part of the ore body is unsheared. The evidence at the Shasta King mine indicates that the scattered residual nodules of waste in the ore, and the unreplaced part of the flow of porphyritic rhyolite that contains the ore, were not replaced because they were not sheared. The isolated unreplaced remnants of rock in the main body of the massive sulfide ore differ in origin and appearance from the partly replaced volcanic breccia in the back of the stopes at the northeast end of the mine in the stope on the 870-foot level (fig. 61*A*).

Many contacts between the ore and the porphyritic rhyolite are sharp. These contacts are marked by bands of white clay gouge that range in thickness from a fraction of an inch to a foot. The wall rock is locally schistose behind the gouge, but there are no sulfide minerals in the gouge or in the foliated wall rock at these localities. The fact that the sericite bands at ore contacts are foliated parallel to it indicates that this zone is not due to hydrothermal alteration alone. Ore solutions were apparently stopped by premineral foliated bands of gouge and sericite at these points. There is no evidence of postmineral movement, such as crushed or slickensided sulfides, that would be sufficient to orient the sericite parallel to the sulfide contact. A few bands of gouge, ranging from a thin film to a foot in thickness, are found in the massive sulfide ore. The gouge is composed of soft, sticky clay and has a sharp contact with the massive sulfide ore. These bands in the ore appear to be unreplaced bands of premineral gouge.

The ore along fault 6 on the 870-foot level (pl. 19) is in sharp contact with unmineralized fault gouge as much as a foot thick, except in the vicinity of adit 6. At this locality the contact is sharp but irregular and does not lie against the fault, and the massive sulfide ore has smooth, curved contacts with porphyritic rhyolite. There is no gouge at the contact and no evidence of movement, although a claylike alteration of the porphyry a few millimeters thick occurs at a few contacts.

A bed of tuff and volcanic breccia overlies the gossan at the outcrop of the ore body, but is found in the stopes only in the southwestern and (less definitely) in the northeastern parts of the mine. Well-bedded shaly tuff occurs in the back of the stope on the 830-foot level, and material closely resembling the volcanic breccia occurs in the back of the stope on the 870-foot level. In this stope the fragments of porphyritic rhyolite in the ore at its upper contact closely resemble the volcanic breccia exposed at the surface (fig. 60). In addition, the massive sulfide ore, now represented by gossan, replaced tuff and volcanic breccia in the vicinity of cross section *C–C′* at the surface. The fragments of porphyritic rhyolite in the ore (fig. 61*A*) have been interpreted by some observers as a breccia formed by replacement along fractures, but the fragments are not of uniform rock type, nor do they have the same type of alteration. Some fragments are silicified, contain no sulfide minerals, and have sharp, smooth boundaries. Other fragments are soft and are replaced by pyrite, sericite, and clay minerals. The soft fragments commonly have gradational boundaries against massive sulfide ore. The fragments of porphyritic rhyolite in the ore in the lower part of the breccia are alined parallel to the ore contact.

It seems probable that the breccia fragments in the ore represent unreplaced fragments in the volcanic breccia-tuff bed and that this bed caps the ore in both the northeastern and southwestern stopes. The tuff bed can thus be expected to lie a short distance above the ore in the central part of the mine unless it pinches out between the outcrop and the stopes.

The openings that served as channels for the ore-forming solutions have not been located. Faults 3, 5, and 12 are mineralized, but only near the gossan. Fault 12 contained as much as 4 feet of massive sulfide ore, which is now oxidized to gossan along the fault below the base of the main ore body (fig. 61*B*), but the massive sulfide changes to slightly mineralized rock less than 50 feet below the ore body. One or all of these faults may have been feeders to the ore body, but it is equally probable, as suggested by R. T. Walker and W. J. Walker (oral communication), that ore solutions traveling horizontally would tend to work out along premineral faults of this type, which would then simulate feeder channels in appearance.

OXIDATION AND ENRICHMENT

The gossan that formed from oxidation of the massive sulfide ore extends from the outcrop of the ore body into the wall of the canyon, for 30–50 feet. Steep topography and a shaly tuff cover have prevented extensive oxidation of the ore, and relict nodules of massive sulfide ore occur in the gossan less than 10 feet from the surface. The gossan is solid and resistant to erosion. It is composed of dense to cellular limonite that contains minor septa of secondary silica forming a coarse silica sponge in the limonite. No collapsed breccia was seen in the gossan, and there is little transported limonite.

The rhyolite below the gossan is iron stained only along fractures.

No assays are available on the gold content of the gossan. On the basis of data from other mines in the district, it can be assumed that the gossan contains about twice as much gold as the primary ore because of residual enrichment of gold. The weight of the gossan is roughly half that of the massive sulfide ore, and little or no leaching of gold occurs in gossans of this type elsewhere in the district.

Secondary copper minerals are rare, but a little chalcocite occurs in the sulfide ore just below the gossan in adit 8.

EXPLORATION POSSIBILITIES

The stratigraphic zone in which the Shasta King ore body occurs has not been explored north of the mine. The ore body thins to the north, but some massive sulfide ore is found in the most northerly mine workings (pl. 19, *E–E′*). It seems probable that additional exploration to the north along the ore zone would disclose other ore bodies, because the district habit of the ore bodies is to occur as discontinuous bodies of minable ore in a general ore zone. The Shasta King ore zone is present in places for a mile or more north of the mine and the more favorable zone just below the base of the upper unit may also be present north of the mine (pl. 4). The Shasta King ore zone could be explored by following the ore north from the underground workings or by surface drilling.

SPREAD EAGLE PROSPECT

The Spread Eagle prospect is 4,500 feet southeast of the Balaklala mine in sec. 13, T. 33 N., R. 6 W., and lies in a steep-sided amphitheater at the head of Motion Creek at an altitude of about 3,000 feet. It was not accessible by road in 1951, but during the active exploration of the prospect a road extended from the Balaklala Angle Station to the prospect. This road is washed out at many places, but it could be made passable with a small amount of repair.

This prospect was discovered some time before 1902, as Aubury (1902, p. 82) reports:

> About 1,500 feet of tunnels, mainly driven by the Scottish American Syndicate, of Denver, Colorado, under bond, show considerable bodies of ore, including some of excellent grade.

W. C. Onn and sons owned the prospect in 1902. By 1908, 3,000 feet of exploratory work had been done. The property was acquired in 1913 by the United States Smelting Refining and Mining Co., the present owners.

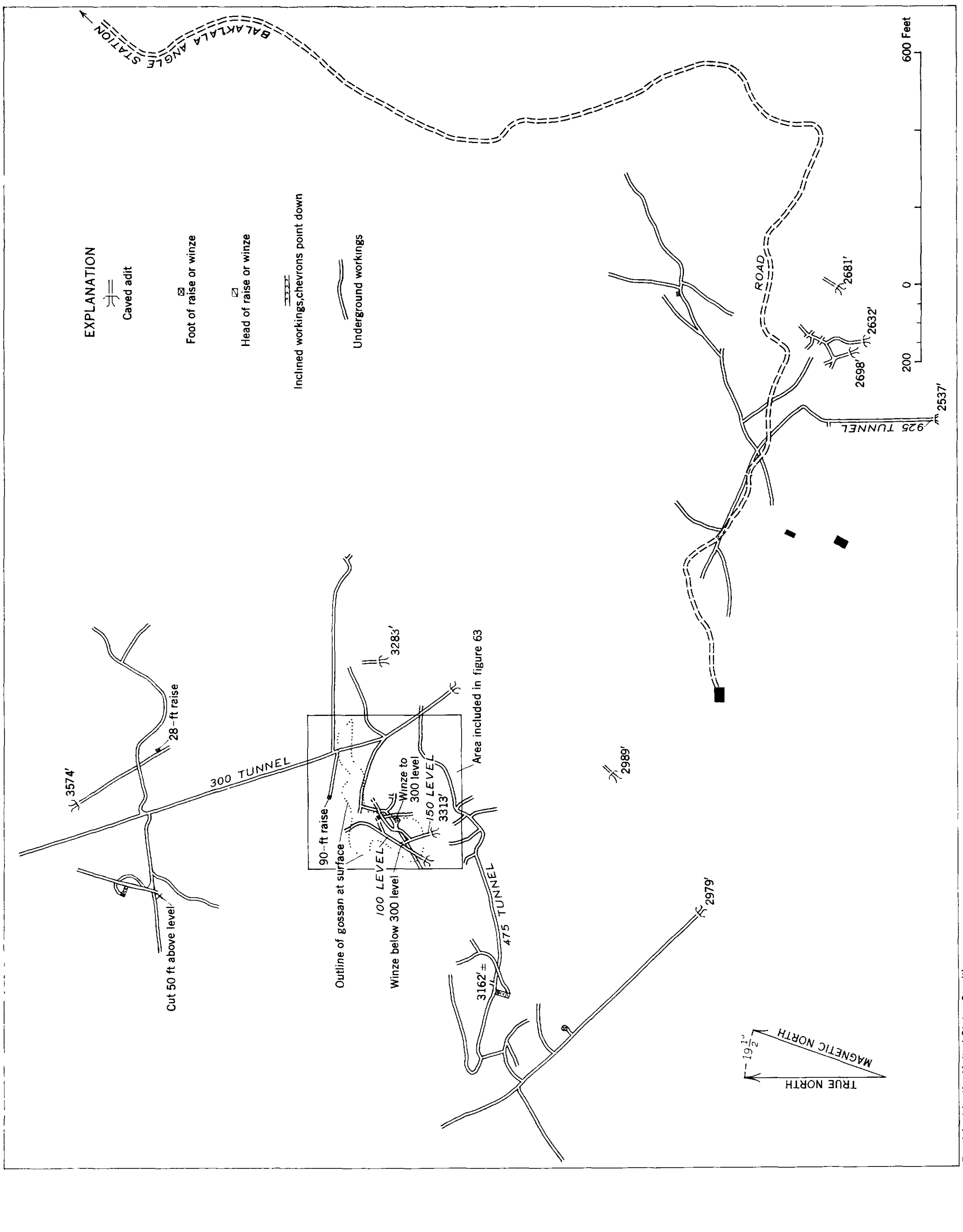
Geology of the mine area.—The Spread Eagle prospect occurs near the gently dipping contact between the upper and middle units of the Balaklala rhyolite. The upper unit in this area consists predominantly of tuff that is poorly bedded, but that locally contains coarser pyroclastic material. Quartz phenocrysts 3 to 5 millimeters in diameter are prominent in most of this unit, and it is considered to be the "cap rock," although very little of the unit originated as a flow. This tuff correlates with the tuff at the base of the coarse-phenocryst rhyolite at the Balaklala and other mines.

The contact between the upper and middle units is gradational, and near the portal of the 475-foot tunnel (fig. 62) this zone consists of a volcanic breccia in which rounded knobs of rhyolite are embedded in a tuff that contains 3- to 5-millimeter quartz phenocrysts. The uppermost part of the middle unit of the Balaklala rhyolite in this area is composed of a fairly persistent bed of coarse volcanic breccia. Below this, beds of pyroclastic rocks alternate with medium-phenocryst (1 to 4 mm) rhyolite.

A layer of rock of doubtful origin occurs in the mine area just below the main gossan. East of the 475-foot tunnel, where this layer is about 100 feet thick, it dips 10°–20° NW. into the ridge. It can be traced for several thousand feet northeast of the prospect, and ranges from a few feet to 100 feet in thickness. The upper part is a green, amygdaloidal, chloritic rock that resembles andesite, but the lower part is light colored and appears to be rhyolite. East and northeast of the gossan the rock is light greenish white with 1-millimeter dark spots, probably chlorite, and is not amygdaloidal. Parts of the layer show sharply defined columnar jointing, and it appears to be a sill of rhyolite that is locally amygdaloidal and is in part chloritized.

Although the upper and middle units of the Balaklala rhyolite are foliated, the upper unit is less foliated than the middle. The foliation strikes northeast and dips steeply. Most of the rocks in the vicinity of the Spread Eagle prospect are heavily pyritized. From the summit of the ridge above the mine at 3,700 feet to the lowest mine workings at 2,537 feet, discontinuous, often lenticular, bands of rock a few feet thick containing as much as 30–50 percent pyrite alternate with thicker bands containing smaller amounts of pyrite. All the rock contains a few percent pyrite, and the entire area above the mine workings is heavily iron stained. Pyritization appears to be controlled by the steep foliation of the rocks; the most foliated rocks are the most heavily pyritized.

Ore deposit.—A gossan from massive sulfide 20–30 feet thick is exposed at the surface, and has been explored by adits. Other than the description of the adits in this ore body, little information is available on the type or amount of mineralization that was found in this exploratory work. It is apparent from old maps and



Furnished by the United States Smelting Refining and Mining Co.

FIGURE 62.—Map of underground workings, Spread Eagle prospect, Shasta County, Calif.

descriptions that heavily pyritized, steeply dipping shear zones were explored at depth in the hope of finding bodies of massive sulfide ore. Diller (1906, p. 13) reports that at the Spread Eagle

> There is a large mass of gossan on the steep slopes at the head of Motion Creek. This has been penetrated by tunnels, which have cut a mass of sulphide ore that appears to strike N. 70°–80° W. and to dip 75° SW. This is traversed by sheared fissures, some nearly parallel to the strike and others transverse. The pyrite ore beneath the gossan is generally wet, soft, and incoherent, and in places it is cemented by quartz gangue in varying proportions. In the lower tunnel some small masses of chalcopyrite are seen.

Material on the mine dumps indicates that, except for the body of sulfide ore under the gossan at the 100- and 150-foot level adits, the pyritized shear zones did not contain minable ore. The lack of stopes confirms this assumption.

The rock on the mine dumps consists either of foliated, slightly pyritized, porphyritic rhyolite (2- to 3-millimeter quartz phenocrysts), or of heavily silicified rhyolite, some of which contains as much as 10 to 60 percent pyrite. Much of the silicified rhyolite is cut by quartz veinlets, and some dump material contains only quartz and pyrite. A few stringers and grains of chalcopyrite are in some of the heavily pyritized rock, but no material of ore grade was seen. One specimen of massive magnetite veined by pyrite was found on the dump from the 475-foot level.

The gossan from massive sulfide ore that crops out above the portals of the 100- and 150-foot levels stands out in sharp contrast to the oxidized material formed from pyritized foliated rock. The massive sulfide gossan contains both dense and porous limonite with quartz septa; it is structureless and nonfoliated and has sharp boundaries against foliated oxidized rock that contained only scattered pyrite. The gossan that crops out at the surface was partly explored by the 100- and 150-foot level adits (figs. 62, 63). This gossan contains an appreciable amount of gold. The surface samples are richer in gold than samples taken in gossan in the tunnels immediately below the surface; it seems probable that there is not only a residual enrichment of gold due to loss of weight in the gossan, but also a residual mechanical enrichment at the surface due to erosion of decomposed gossan. The grade of the unoxidized massive sulfide on the 100- and 150-foot levels is not known.

Some exploratory drilling has been done along the ridge northwest of the mine but no massive sulfide was found. Extensive areas underlain by the favorable zone—the contact between the upper and middle units of the Balaklala rhyolite—remain to be explored between the Spread Eagle and the Keystone and Balaklala mines, and between the Spread Eagle and the Balaklala Angle Station gossan. The favorable zone has been eroded south and east of the Spread Eagle prospect.

STOWELL MINE

The Stowell mine is in the canyon of Spring Creek, about 1 mile southwest of the Keystone mine in secs. 14 and 23, T. 33 N., R. 6 W., at an altitude of 3,100 feet. The property was formerly known as the Grab and also as the Webster Consolidated group. The massive sulfide ore body of this mine was one of the early discoveries in the district, and 600 feet of exploratory adits had been completed by 1902. The mine was purchased by the United States Smelting Refining and Mining Co. in 1916; all the stoping and most of the extensive workings shown on plate 20 was done by this company. Ore was carried from the Stowell mine to the Keystone mine by means of aerial tram.

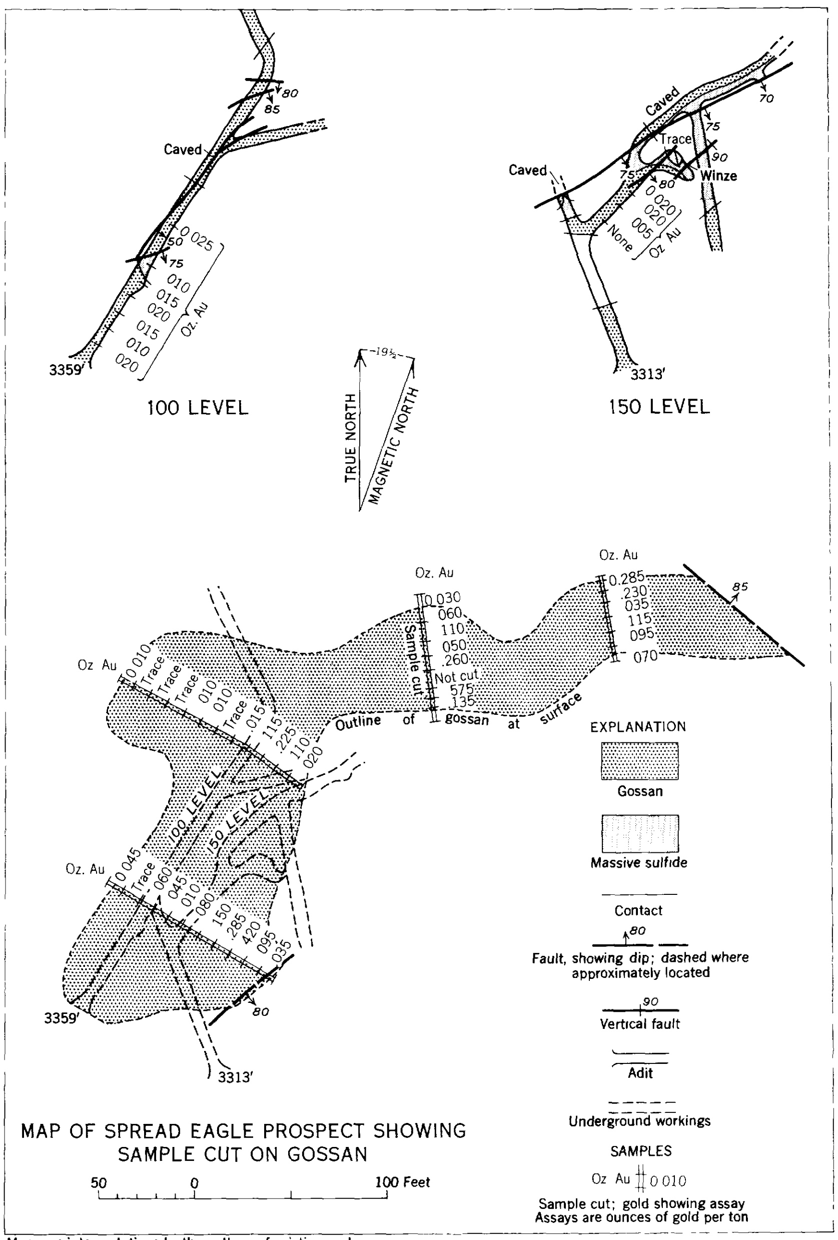
Mining operations were discontinued in 1919 when the Keswick smelter closed. The workings, except for short distances in the Gossan and Pyrite tunnels, are now inaccessible, and the mine plant has been dismantled. The mine was accessible only by a trail from the Keystone mine in 1951.

The Stowell mine has produced 39,538 tons of massive sulfide ore that contained 0.03 ounce of gold, 1.09 ounces of silver, and 3.0 percent copper.[10] The zinc content of the mined ore is not known, but specimens on the mine dumps contain zinc.

Geology.—The ore body at the Stowell mine lies 50 to 170 feet stratigraphically below the base of the coarse-phenocryst rhyolite ("cap rock"). Little or no massive coarse-phenocryst rhyolite is present in the mine area, and the upper unit of the Balaklala rhyolite is represented by tuff beds and coarser pyroclastic material that contain some quartz phenocrysts more than 5 millimeters in diameter. The ridge northeast of the Stowell mine is capped by the upper unit of the Balaklala, but as the bedding dips gently northeast, the base of the upper unit is exposed at only a few places between the Stowell and Keystone mines. The middle unit, and probably the upper part of the lower unit, is exposed in the deep canyon of Spring Creek southwest of the mine.

In the vicinity of the Stowell mine, the rocks crop out in the sequence tabulated below; no information is available on the types of rock in the underground workings except for the location of the base of the coarse-phenocryst rhyolite, which is known from drill-hole logs.

[10] Data furnished by the United States Smelting Refining and Mining Co. Published with permission of the owners.



Maps are interpretations by the authors of existing underground data compiled during the period of the mine's operation. Published with the owner's permission

FIGURE 63.—Levels and sample cuts of Spread Eagle prospect, Shasta County, Calif.

Drill-hole logs	*Feet*
Coarse-phenocryst rhyolitic tuff ("cap rock")	100+
Tan poorly bedded shaly tuff	20–50
Lenticular flow of porphyritic rhyolite (2- to 3-mm phenocrysts)	10–50
Nonporphyritic rhyolitic flow breccia and tuff	50±
Porphyritic rhyolite (2- to 3-mm phenocrysts). The ore crops out in this flow	50±
Nonporphyritic rhyolite	125±
Chloritic amygdaloidal greenstone interlayered in the Balaklala rhyolite	100±
Nonporphyritic rhyolite	150+

Drill-hole data indicate that the dip of lava flows and tuff beds in the mine area is 20°–30° NE., but no reliable attitudes were found on bedded material on the surface. The northeast dip flattens and probably reversed northeast of the mine, as the contact of the coarse-phenocryst rhyolite above the Keystone mine is at an altitude of about 3,100 feet. The middle unit of the Balaklala thins at the Stowell mine, as little porphyritic rhyolite is present. The middle unit at the Stowell mine contains more nonporphyritic rhyolite than is common elsewhere in the district and contains a thin flow of greenstone, as at the Shasta King mine.

Two steep northeastward-trending faults occur in the mine area. On each of these, rocks on the northwest side moved down about 100 feet relative to the southeast. The northernmost fault lies about 100 feet northwest of the portal of the 300-foot level adit, but the fault is poorly exposed as it is largely covered by dumps. The flow of amygdaloidal greenstone is offset by both of these faults, but their location in the underground workings is unknown. These workings should have intersected the northernmost fault.

Ore deposit.—Little information is available on the ore bodies of the Stowell mine. The mine workings are inaccessible, and no geologic maps were located, but a few logs of drill holes were available. The outlines of the stopes shown on plate 20 and figure 64 are from underground maps. Specimens of ore from mine dumps and from spill along the tramline were studied.

The ore bodies are massive sulfide that contains about 3.0 percent copper and considerable zinc. The maps indicate that at least three separate lenses of ore were discovered, but drill holes indicate that there was much pyritized rock, in addition to massive sulfide ore, along the base of the "cap rock." Some of the material between ore bodies probably consisted of pyritized rhyolite, although the sharp contact of the gossan from massive sulfide at the surface suggests that, as in other mines in the district, there was a sharp boundary between pyritized or barren rhyolite and massive sulfide ore. However, the stope outlines may not mark the limits of massive sulfide, because sulfide containing copper below the cutoff grade, or sulfide containing a high percentage of zinc would be left unmined.

The large gossan at the outcrop of the Stowell ore body is a typical gossan derived from massive sulfide and it is composed entirely of dense and cellular limonite containing minor septa of secondary quartz and no relict rock material.

Relict pyrite occurs within a few feet of the surface in the Gossan and Pyrite tunnels, and the part of the gossan that contains more than 50 percent limonite (the rest being pyrite) extends into the hill less than 50 feet. Gossan stringers and irregular areas of gossan are found as much as 100 feet in from the hill slope.

The gold and silver values shown in the gossan samples on plate 20 are confined almost entirely to the true gossan; the adjacent strongly pyritized rhyolite is barren. The boundary between the gossan and the pyritized rhyolite is sharp at most places.

A small amount of surface exploratory drilling has been done around the Stowell mine. In addition, a widely spaced line of holes was drilled between the Stowell and the Keystone mines. These holes cut the mineral zone below the base of the coarse-phenocryst rhyolite, but did not cut massive sulfide ore. This area deserves further exploration, as rocks of the upper unit of the Balaklala rhyolite, which are largely tuff, cover most of the area between the two mines, and the ore zone below the base of the upper unit of the Balaklala is

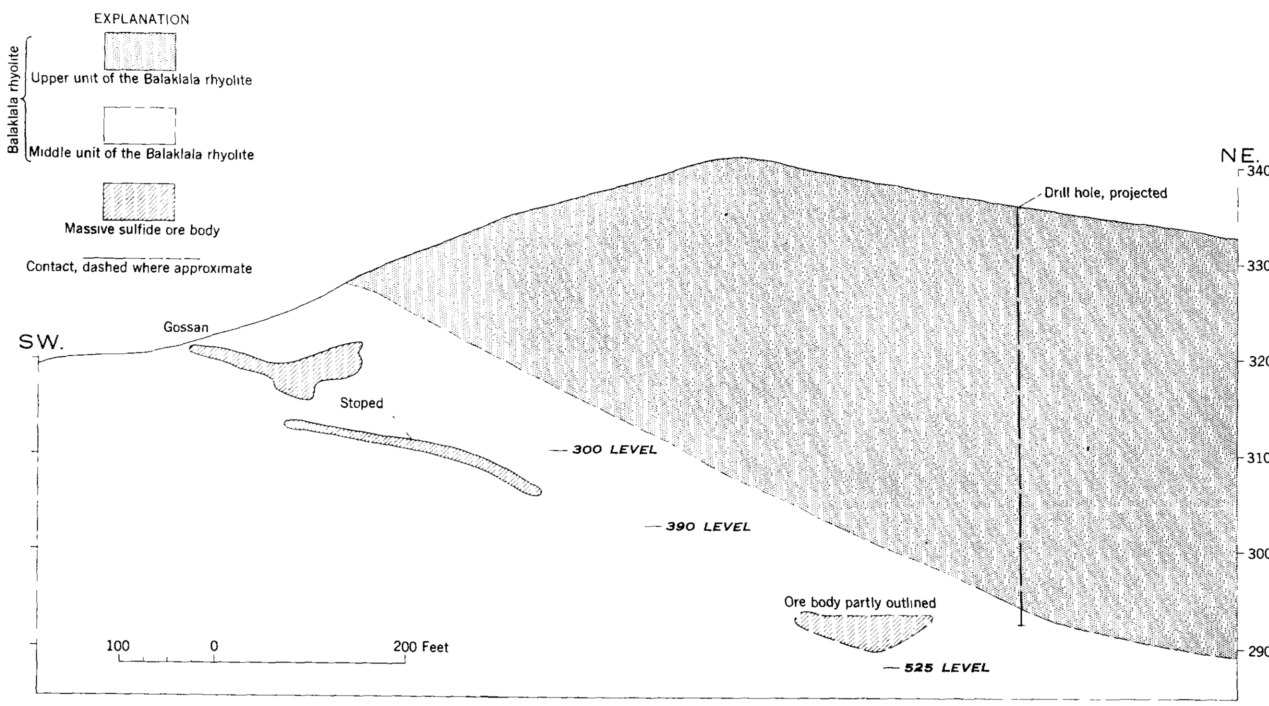


FIGURE 64.—Generalized projection looking northwestward through the Stowell mine showing the relationship of the ore bodies to the base of the upper unit of the Balaklala rhyolite.

present over a wide area to the north, northeast, and east of the Stowell mine. The vertical position of the ore zone is fairly well known northeast of the Stowell, but no features were found that would indicate the horizontal localization of massive sulfide bodies. All this area between the Stowell and Keystone mines is geologically favorable ground for massive sulfide bodies.

SUGARLOAF PROSPECT

The Sugarloaf prospect, formerly known as the Galvin mine, is on the east side of Sugarloaf Mountain between Spring Creek and Boulder Creek. It is about 3,000 feet northeast of the Hornet ore body of the Iron Mountain mine and is in sec. 26, T. 33 N., R. 6 W. The property was located early in the history of the district; 1,300 feet of adits had been driven by 1902. Aubury (1908, p. 84) reports:

> Development work consists of 12 tunnels aggregating 4,365 feet. A very distinct fissure bearing approximately north and south has been followed about 1,000 feet in tunnels No. 2 and 5. Small pockets of sulfide ores were encountered, but no definite ore body has yet been discovered. The property is owned by the Copper Mountain Consolidated Mining Company of Redding.

In 1951 all adits were caved at the portal, except one. The property is now owned by R. T. Walker and W. J. Walker.

Geology.—The rocks in the vicinity of the Sugarloaf prospect are nonporphyritic rhyolite, and porphyritic rhyolite containing 1- to 2-millimeter quartz phenocrysts. A large part of both rock types is flow breccia or pyroclastic material. Most of the rocks are strongly foliated and silicified, and are pyritized locally. The foliation strikes N. 20°–60° E. and dips from vertical to 40° NW. No reliable attitude on bedding could be found, but the trend of flow contacts suggests a steep northwest dip. From a projection of the general structure striking northeastward from Iron Mountain, and from the location of the underlying Copley greenstone to the northwest and to the southeast of the Sugarloaf prospect, it seems probable that the prospect is on the steeply dipping south flank of a syncline. The axis of the syncline should lie north of the portals of the adits but it is not known if the underground workings explored the syncline as their extent and direction are unknown.

No stratigraphic distinction can be made at the Sugarloaf prospect between the lower and middle units of the Balaklala rhyolite. As at Iron Mountain, the rocks consist of interlayered porphyritic and nonporphyritic rhyolite and pyroclastic material. At Iron Mountain these mixed flows and pyroclastic rocks are overlain by crystal tuff and volcanic breccia that probably represent the transition zone between the upper and middle units of the Balaklala, but at the Sugarloaf prospect the crystal tuff beds are absent. At both localities, the mixed flows are underlain by the Copley greenstone, but at the Sugarloaf prospect there is no predominance of flows of nonporphyritic rhyolite at the base of the Balaklala and the stratigraphic relationships are obscure. It is probable that the rocks near this prospect are somewhat lower in the stratigraphic sequence than those that contain the ore bodies at Iron Mountain.

Ore deposit.—The rock in the area of the mine portals is locally well pyritized, but no gossan from massive sulfide was found. The mine dumps contain much material that is 20–75 percent pyrite in a matrix of quartz or siliceous rhyolite. The pyrite is generally in coarse cubes as much as 5 millimeters across, and very little of this material is the fine-grained, solid pyrite that is characteristic of the massive sulfide ore bodies. The coarse-grained pyrite in quartz or siliceous rhyolite is identical with that found in a 300-foot prospect adit that was driven about half way between the Busy Bee workings of the Iron Mountain mine and the Sugarloaf prospect (fig. 58).

Exploratory work on Sugarloaf Mountain, except for the Busy Bee workings, has not disclosed any bodies of massive sulfide similar to those at Iron Mountain or at other mines in the district. However, the trend of the ore zone and geologic structures extending northeastward from Iron Mountain appear to continue through Sugarloaf Mountain. Figure 58 shows the spatial relationship between the Sugarloaf prospect and the nearby mines.

The lens of massive sulfide that is exposed in the Busy Bee workings of the Iron Mountain mine has not been fully explored. A crosscut near the end of these workings exposes massive sulfide 45 feet wide, and the face of the adit is in massive sulfide. The ore appears to be following a steeply dipping shear zone, similar to the Hornet ore body, rather than being localized along flat-lying structures. No flat-lying structures that could act as ore controls were found on Sugarloaf Mountain, probably because of the steep premineral foliation that has been superposed on the rock. It is possible that steeply dipping massive sulfide bodies similar to the Hornet and the Busy Bee ore bodies occur under Sugarloaf Mountain in the area north of the Sugarloaf adit portals, but this area may have been partly explored by underground workings.

SUTRO MINE

The Sutro mine is in secs. 29 and 30, T. 34 N., R. 5 W., about 6,000 feet northwest of the Mammoth mine. The present Sutro mine includes the old Sutro and the Summit group of claims, which later became known as the Stauffer mine.

The Summit mine, discovered before 1902, was originally owned by M. E. Dittmar. By 1902 several adits had been driven; the mine was explored by the Mount Shasta Gold Mining Corp. in that year (Aubury, 1902, p. 89). The Summit mine was sold to the Stauffer Chemical Co. about 1908. Nothing is known of the early history of the original Sutro mine except that it was bonded to the Stauffer Chemical Co. in 1908 (Aubury, 1908, p. 97) (it may be that the Summit mine was meant, instead of the Sutro). The Sutro and Stauffer mines are on different ends of the same ore zone. The Sutro mine is now owned by the United States Smelting Refining and Mining Co.; most of the stoping was done after this company acquired the property.

The mine workings (pl. 21) are not accessible, and nothing remains of the mine plant. The rails were removed many years ago from the railroad that connected the Sutro with the Mammoth mine, and the roadbed is largely washed out.

The massive sulfide ore bodies of the Sutro mine contain much higher values in gold, silver, and copper than most of the other massive sulfide ores of the West Shasta copper-zinc district. Only ore from the Golinsky mine contained more gold, and none of the other ore bodies contained as much copper. The zinc content is unknown. Records show that a total of 35,300 tons of ore was shipped to the Kennett smelter by the United States Smelting Refining and Mining Co. This ore contained 0.08 ounce of gold, 6.40 ounces of silver, and 7.44 percent copper.[11]

The grade of the ore produced varied at different times during the operation of the mine. It is probable that the lack of uniformity in the grade of the ore represents different mining methods and cutoff grades rather than differences in character of the ore. During the time that the United States Smelting Refining and Mining Co. operated the mine, monthly records of shipments show that the ore was remarkably uniform and higher in grade than the average for the total mine production. For several years of operation, the best ore averaged about 0.09 ounce of gold and almost 10.0 percent copper. According to available records, the ore contained from 0.03 to 0.10 ounce of gold, 2.5 to 8.1 ounces of silver, and 2.5 to 11 percent copper during the period that ore was shipped to the smelter. The monthly records of shipments show that when the mine was turned over to lessees, during the latter part of its operation, the grade of ore produced was between 2 and 3 percent copper, probably because only the lower grade ore was available to lessees. Ore was mined continuously from 1913 to 1918, and again from 1923 to 1925.

Geology.—The ore bodies of the Sutro mine are assumed to have formed along the contact between the upper and middle units of the Balaklala rhyolite (pl. 21). The ore bodies are in a light-green siliceous rhyolite containing 2-millimeter quartz phenocrysts that forms the uppermost part of the middle unit in this area.

The middle unit of the Balaklala rhyolite contains an unusually large amount of pyroclastic material in the area between the Mammoth and the Sutro mines, and also to the northeast of the Sutro. At least 300 feet of shaly tuff and coarse and fine volcanic breccia was deposited in a sedimentary basin in this area. Much of the material is well bedded, apparently water deposited, and the coarser material is in part volcanic conglomerate (fig. 17). Many flows of medium-phenocryst rhyolite and a few flows of nonporphyritic rhyolite are interbedded with the tuff. None of the individual beds of pyroclastic material has much lateral continuity; stubby lenses of coarse volcanic breccia are common. The rock in which the ore bodies are assumed to have formed is a lenticular flow that wedges out between the tuff at the base of the upper unit and a bed of volcanic breccia below.

The upper unit of the Balaklala (the "cap rock") in the Sutro area is commonly a soft, punky, tan rock containing quartz phenocrysts as much as 5 millimeters in diameter; some of the phenocrysts are dark smoky gray. The characteristic coarse-phenocryst tuff and tuff breccia is found locally at the base of the upper unit, and the lower part of the coarse-phenocryst rhyolite contains bands of pyroclastic material that diminish in quantity upward in the section. On the ridge 500 feet in elevation above the mine (about 200–300 feet stratigraphically), the upper unit contains no pyroclastic beds.

The rocks in the mine area are cut by many mafic dikes. Only the principal ones are shown on the map, but many small dikes occur also. The largest (pl. 21) is 25 feet wide and is continuous, although its outcrop at the surface was not located because of poor exposures in the mine area. This dike crops out in the gulch west of the mine, and several smaller ones can be seen at the surface. They weather greenish brown and contain flecks of a dark-green mineral, probably chlorite. Unweathered specimens from the mine dumps resemble fine-grained gabbro that is similar to the mafic intrusive southeast of the Stowell mine, and to the small mafic dikes and plugs found in many other parts of the district.

The rocks between the Mammoth and the Sutro mines and at the Sutro mine are gently dipping to flat but

[11] Data furnished by United States Smelting Refining and Mining Co. Published with permission of the owners.

contain several broad warps. The Sutro mine appears to be on the south flank of a large warp, with a structural low point between the Mammoth and the Sutro mines, and a structural high point north of the Sutro mine. However, the thick section of pyroclastic material in this basin implies that the depression may be in part of primary origin.

Many small faults are shown on the underground maps, but they were not seen at the surface because of poor outcrops and because no lithologic offsets can be seen in the uniform "cap rock" over the ore bodies.

Ore deposits.—Gossan derived from massive sulfide is present where the Trounce ore body crops out (pl. 21), and large bodies of rock are heavily pyritized at the northeast and the southwest end of the ore zone. The most heavily pyritized rock lies along the trend of the ore bodies, that is, near the portals of the Stauffer tunnels 2 and 3 and the Old Sutro and *B* adits, but much pyritized rock extends throughout the top of the middle unit of the Balaklala rhyolite in this area. Strongly pyritized rock extends as much as 150 feet below the base of the upper unit. The upper unit contains scarcely any pyrite and is not iron stained.

No information is available on the physical characteristics of the ore bodies. According to the map pattern, they appear to be isolated bodies of massive sulfide that lie immediately below the base of the upper unit of the Balaklala rhyolite. The obvious vertical ore control is the contact between the upper and middle units of the Balaklala. The apparent relationship shown on the sections between the ore bodies and the structurally high and low points along the base of the upper unit suggests a possible control of ore deposition by small warps in this contact, although ore may have been formed along feeder fissures at these points. Old maps indicate that a strong gouge was present at the top of the ore in all the stopes at the northeast end of the mine (pl. 21, *A–A′*), and the ore bodies are presumed to have the sharp contacts to barren wall rock which are characteristic of the ores of this district. However, some low-grade, pyritized rock was present under 1003 raise stope. Old maps show a body of disseminated ore beneath this stope that contains 0.01 ounce of gold, 0.40 ounce of silver, 1.7 percent copper, 55.0 percent insoluble material, 15.0 percent iron, 18.0 percent sulfur, and 10.0 percent alumina.

Some exploratory drilling was done from the surface along the sides of the ore zone and a small amount of underground exploration was done normal to strike of the ore run. None of these exploratory workings extended more than a few hundred feet northwest of the known ore zone, and a considerable area of favorable ground under the "cap rock" remains to be explored northwest of the mine. As the Sutro is on the south flank of a broad warp, the favorable zone can be expected to rise to the northwest with the topography, so that there is less of a topographic handicap to surface drilling than appears.

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