Geology and Mineral Deposits of the Osgood Mountains Quadrangle Humboldt County, Nevada

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Geology and Mineral Deposits of the Osgood Mountains Quadrangle Humboldt County, Nevada

By PRESTON E. HOTZ *and* RONALD WILLDEN

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Stratigraphy and structure of Paleozoic rocks, granitic intrusions, contact metamorphism, and deposits of tungsten, gold, and other minerals

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By Preston E. Hotz and Ronald Willden

ABSTRACT

The Osgood Mountains quadrangle is in north-central Nevada northeast of Winnemucca, the principal town in the region. The quadrangle includes two north-northeast-trending mountain ranges, the Osgood Mountains on the east and the Hot Springs Range on the west, which are separated by a narrow valley and are bounded on the east and west by broad alluviated basins. Large deposits of tungsten and gold have been mined in the northeastern part of the Osgood Mountains; small deposits of quicksilver, lead, zinc, and gold are known in the Hot Springs Range, and some prospecting has been done on barite deposits in the Osgood Mountains. Some quartzite beds are potential sources of silica.

Paleozoic rocks exposed in the quadrangle include strata of Cambrian, Ordovician, Mississippian, Pennsylvanian, and Permian age.

Four units of Cambrian and one of probable Cambrian age have been recognized. The oldest, the Osgood Mountain quartzite of Cambrian(?) age, is exposed only in the southern part of the Osgood Mountains. Most of the formation is a relatively pure crossbedded quartzite with a few thin shaly partings. In places an impure quartzite unit, the Twin Canyon member, forms the upper part of the Osgood Mountain quartzite. The formation is unfossiliferous, but it grades upward into the Preble formation, which contains fossils of Middle and Late Cambrian age. The Preble formation, which also occurs exclusively in the Osgood Mountains, is predominantly shale but includes a few quartzite beds in its lower part and is composed of interbedded limestone and shale in the middle and upper part of the section. A new formation, the Paradise Valley chert of Late Cambrian age, is the next youngest unit, but its stratigraphic relations to the Preble formation are not known because it is restricted to a small area on the northwest side of the Hot Springs Range. The formation is predominantly chert but includes some thin beds of shale and limestone. On the basis of fossil faunas the Paradise Valley chert is correlated with the lower part of the Dunderberg shale in the vicinity of Cherry Creek and McGill, eastern Nevada. The uppermost Cambrian unit is the Harmony formation, which rests in depositional contact on the Paradise Valley chert. The Harmony makes up most of the Hot Springs Range in the quadrangle and occupies a thrust plate in the Osgood Mountains. Feldspathic sandstone and shale are the predominant rock types of the Harmony formation, but the formation also includes some limestone and a little chert. Trilobites of Late Cambrian age were found in the limestone at two widely separated places.

Two Ordovician units having very different lithologies were mapped. The Comus formation of Early and Middle Ordovician age is predominantly an alternating sequence of dolomite, limestone, and shale, with subordinate amounts of chert, siltstone, and tuffaceous(?) material. Sandstone and quartzite are conspicuously absent. In the Valmy formation of Early, Middle, and Late Ordovician age, chert and siliceous shale predominate. The Comus and Valmy formations are not in contact in this quadrangle, but from relations elsewhere the Comus is regarded as autochthonous, whereas the Valmy is believed to have been brought into the area by thrust faulting.

A formation of Early and Late Mississippian age, the Goughs Canyon formation, occupies a thrust plate in the Osgood Mountains. Its stratigraphic relations with other Paleozoic rocks are not known. The formation is composed mostly of altered volcanic rocks of medium to basic composition and coarse-grained fossiliferous limestone, and minor amounts of calcareous shale, siliceous shale, and chert.

Strata of Middle Pennsylvanian to Late Pennsylvanian and Early Permian age rest unconformably on the older Paleozoic rocks in the Osgood Mountains. The beds include the Battle formation, a dominantly terrestrial conglomerate of Middle Pennsylvanian age (Atoka to Des Moines), which underlies and interfingers with the Etchart limestone of Middle Pennsylvanian (Des Moines or older) to Late Pennsylvanian or Early Permian (Missouri or Wolfcamp) age. The Etchart limestone is predominantly a sequence of limestone and sandy limestone, with some interbedded dolomite, minor amounts of calcareous shale, and lenticular beds of conglomerate. A clastic facies partly equivalent in age to the Etchart limestone, the Adam Peak formation, has been thrust over the Etchart limestone. The Adam Peak is chiefly shale, siltstone, dolomitic sandstone, and chert, but the formation includes some limestone and dolomite.

A sequence of sandstone, shale, chert, and altered volcanic rock, the Farrel Canyon formation, crops out in the northwestern part of the Osgood Mountains. No fossils have been found in the Farrel Canyon formation, hence its stratigraphic position is uncertain; but lithologically it resembles parts of the Pumpernickel formation of Pennsylvanian(?) age and the Havallah formation of Middle Pennsylvanian (Atoka) and Permian age which occur elsewhere in north-central Nevada.

One large intrusive body of granodiorite cuts the Preble formation in the Osgood Mountains and three smaller stocks cut the Harmony formation in the Hot Springs Range. A lead-alpha age determination of 69 million years for the Osgood Mountains stock dates the granodiorite as very Late Cretaceous. Except for a slightly more mafic border facies in parts of the Osgood Mountains stock, the composition of the granodiorite is

very uniform. Alteration has affected a small area in the Osgood Mountains stock, however, and large parts of the smaller stocks of the Hot Springs Range have likewise been transformed. The alteration, which was accompanied by introduction of pyrite, resulted in albitization and sericitization of feldspars, destruction of biotite, and some addition of quartz. Small bodies of quartz diorite in the northern end of the Osgood Mountains show similar alteration effects. The granodiorite is cut by aplite dikes and small dikes and veinlets of quartz-feldspar pegmatite. Thin tabular bodies of intrusive dacite porphyry are widespread in the Paleozoic sedimentary rocks of the Osgood Mountains. Probably the dikes are genetically related to the granodiorite, but their relative ages are uncertain.

Remnants of formerly widespread volcanic rocks of Tertiary age (late Miocene(?) to middle Pliocene(?)) are scattered over the quadrangle in the lower parts of the Osgood Mountains and the Hot Springs Range. These are mostly andesitic and some basaltic flow rocks, locally underlain by tuffaceous rocks of rhyolitic composition. Some small remnants of conglomerates that are presumably of Tertiary age were also observed but are not shown on the map.

Surficial deposits of Quaternary age have been divided into older fan gravels, talus, and alluvium. A basalt flow on the southern boundary of the quadrangle is also probably of Quaternary age.

A conspicuous aureole of contact metamorphism surrounds the Osgood Mountains granodiorite stock, and a small but well-defined halo of metamorphism surrounds the largest of the small stocks in the Hot Springs Range. In the Osgood Mountains, the contact metamorphic aureole is as much as 10,000 feet wide; the rocks involved were mainly shales and carbonate rocks of the Preble formation, although some other formations were also affected. The shales were transformed to hornfelses in which biotite, andalusite, and cordierite were formed. The carbonate rocks were converted to marble, light-colored calc-silicate rock, and dark tactite composed of garnet, pyroxene, other lime-silicate minerals, and the economically important tungsten-bearing mineral, scheelite. Paleozoic volcanic rocks were generally too far from the intrusive body to be contact metamorphosed, but at one or two places there has been recrystallization and formation of plagioclase, pyroxene, actinolitic amphibole, and biotite. A narrow aureole of metamorphosed shale and feldspathic sandstone surrounds a small granodiorite stock in the Hot Springs Range. The shale has been darkened and hardened, and where it is most intensely metamorphosed, andalusite and biotite have formed. Some endomorphism of the Osgood Mountains granodiorite stock has taken place along the contact by reaction between carbonate country rock and the intrusive. The granodiorite contains pyroxene as the principal mafic mineral instead of hornblende or biotite, sphene is plentiful, and orthoclase is somewhat more abundant.

The main structural elements strike closely parallel to the northeasterly trend of the ranges. Within both ranges northwest-striking cross faults are about perpendicular to the northeast structures. A major broad anticline involves chiefly the Osgood Mountain quartzite and Preble formation in the Osgood Mountains. This structure is concealed by younger thrust plates and unconformable Pennsylvanian strata in the northern part of the range, and by Tertiary volcanic rocks in the southern part of the range. The younger Paleozoic rocks are steeply tilted and tightly folded at many places; much of the folding is related to thrust faulting. An imbricate thrust zone affecting rocks of Cambrian, Mississippian, and Pennsylvanian and Permian age on the west side of the range is the dominant structural feature of the northern two-thirds of the Osgood Mountains. Other thrust faults are exposed along the crest and east side of the range, involving many of the same rocks. A major thrust fault that probably is equivalent to the Roberts Mountains thrust fault, a major structural feature in north-central Nevada, is postulated to explain the occurrence of two Ordovician formations of different facies—the Valmy and Comus formations—in the same area. The Valmy is believed to have been carried in on a thrust plate.

The rocks in the Osgood Mountains are also cut by north-striking high-angle longitudinal faults, which are more continuous and more important structurally than the high-angle cross faults. The age of most of the high-angle longitudinal faults is not known; however, the Getchell fault on the east side of the range is younger than latest Cretaceous, and some of the longitudinal faults on the east side of the Osgood Mountains are range-front faults of late Tertiary age.

The principal structural features of the Hot Springs Range are a series of asymmetrical, north-plunging, westward-overturned folds in the Paradise Valley and Harmony formations. The faults are normal vertical or high-angle faults in this part of the range; they can be grouped into a north- to northeast-striking set and a northwest-striking set. Many of the faults cut the Tertiary volcanic rocks.

The structural relation between the Osgood Mountains and Hot Springs Range is obscured by an alluvium-filled valley between the two ranges. Two possibilities suggested by the observable data are (1) that the Hot Springs Range is an autochthonous block on the west side of a major anticline whose crest is in the Osgood Mountains; or (2) that the Hot Springs Range is on the upper plate of a thrust fault. High-angle normal faults on the east side of the Hot Springs Range, which postdate the Tertiary volcanic rocks, may be the structural boundary between the mountain blocks. High-angle normal faults also bound the east side of the Osgood Mountains and the west side of the Hot Springs Range.

Mining activity has been directed chiefly toward exploitation of tungsten and gold deposits in the Osgood Mountains. Active mining of tungsten ore began in 1942; by the end of 1955 more than 1,300,000 tons of ore containing more than 590,000 units of WO_3 had been produced. Between 1938 and 1950 production from the Getchell gold mine exceeded $16 million in value. A small production of quicksilver and gold has come from deposits in the Dutch Flat district in the Hot Springs Range. The placers at Dutch Flat have yielded gold valued at about $200,000. Lead-silver and copper deposits of minor importance are known but have yielded little or no production. Nonmetallic deposits include barite prospects in the Osgood Mountains and high-purity quartzite suitable as a source of silica.

Tactites formed by metamorphism of limestone of the Preble formation adjacent to the granodiorite stock in the Osgood Mountains have been the main source of scheelite, the ore mineral of tungsten. Molybdenite has a sporadic distribution in the tactite zone, and accessory amounts of pyrite, chalcopyrite, sphalerite, and galena are present. The tactites are generally tabular bodies parallel to the granodiorite contact. Many of the largest and most productive tactite bodies are situated in troughs and sharp reentrants in the granodiorite contact surfaces.

The Getchell gold deposit is a gold-arsenic association in fractured rocks along the Getchell fault, a high-angle normal fault zone on the northeast base of the Osgood Mountains.

Gold is associated with the arsenic sulfides orpiment and realgar. This epithermal deposit was formed later than the tungsten deposits.

Other gold deposits are on the western side of the Hot Springs Range near Dutch Flat, where gold-bearing quartz veins cut a small granodiorite stock and the surrounding sedimentary rocks of the Harmony formation. The history of mining from these small veins is unknown, but there is no evidence of any important production.

Three small quicksilver deposits are known in the Dutch Flat district in the Hot Springs Range. The only recorded production, from the Dutch Flat mine, has amounted to less than 2 flasks per year since 1942. Cinnabar fills fractures in altered feldspathic sandstone and shale of the Harmony formation and occupies spaces between mineral grains in sandstone.

Placer deposits at Dutch Flat contain significant amounts of gold, scheelite, and cinnabar. Economic recovery of scheelite and cinnabar has not been accomplished.

Three small barite deposits are situated in the Osgood Mountains. None has had any commercial production, and only one appears to be a potentially commercial deposit.

The principal mineral resources of the area are tungsten and gold. Although the reserves of tungsten have been reduced by high production since 1951, and most if not all of the ore that could be readily mined by open pits is gone, there remain important underground deposits of tactite that can be mined. It is very unlikely that new scheelite deposits will be discovered at the surface, but underground exploration might show important extensions of known ore bodies and possibly totally unknown deposits. There is a large reserve of gold-arsenic ore at the Getchell mine, but its exploitation must await more favorable economic conditions and technological improvements. Probably no important production of quicksilver can be expected from the known deposits, and the outlook for discovery of new resources of cinnabar is not good; but the placers at Dutch Flat might yield a moderate amount of quicksilver and tungsten as well as gold. Probably no more than token amounts of galena and sphalerite will ever be produced. Barite and possibly silica may eventually be produced in modest quantities.

INTRODUCTION

LOCATION, CULTURE, AND ACCESSIBILITY

The Osgood Mountains quadrangle is in Humboldt County, north-central Nevada, between lat 41°00′ N. and 41°15′ N., and long 117°15′ W. and 117°30′ W. (fig. 1).

No established towns are within the quadrangle, but Golconda (pop. 430) is about 3 miles south of the southern boundary. Winnemucca (pop. 2,847), the county seat and the most important town in the region, is about 16 miles west of Golconda. Getchell townsite is a semipermanent community with modern facilities at the Getchell mine, but the population varies widely depending on the number of men employed in the mines and mill. A few scattered ranch houses are the only other habitations in the quadrangle. As in most rural areas of Nevada, mining and cattle ranching are almost the only means of livelihood.

Main lines of the Southern Pacific Railroad and the Western Pacific Railroad run on parallel rights-of-way through the southeast corner of the quadrangle. Stations for both lines are located at Golconda. Red House, on the Southern Pacific less than 0.1 mile beyond the southeast corner of the quadrangle, is the main shipping point for the Getchell mine.

The valley areas are readily accessible by automobile road from either south or north, but most of the travel is to and from the south where roads connect with U.S. Highway 40, which passes through Golconda. A good dirt road crosses the full length of the quadrangle from Golconda north through Eden Valley (pl. 1) to the Little Humboldt River; a side road branches off near Stone Corral and leads to the east side of Paradise Valley and the Dutch Flat quicksilver mine. The road along the east side of the Osgood Mountains to the Getchell mine connects with State Route 18, which leaves U.S. Highway 40 about 1 mile east of Golconda. It is paved for 13.5 miles to the junction in NW¼ sec. 20, T. 37 N., R. 42 E. where a good unpaved road (State Route 18) continues eastward to Midas and southeastward to Red House. The paved road is new and is not shown on the topographic map of the quadrangle, but it is about 1 mile east of and roughly parallel to the road that passes near Lone Butte and Hugh Bains ranch. The lower parts of several valleys in the ranges are accessible by poor roads and jeep trails, but the higher parts can be reached only on foot or horseback, except in the northern part of the Osgood Mountains where the steep Burma Road from the Getchell mine crosses the range and connects with the Eden Valley road via Anderson Canyon. All the tungsten mines in the Osgood Mountains are accessible by fair to good roads.

PHYSICAL FEATURES

Two mountain ranges—relatively short in comparison with many Nevada ranges—and a narrow intervening valley are the principal gross topographic features in the quadrangle (fig. 2). The Hot Springs Range trends north-northeast through the northwest part of the quadrangle and ends at the Little Humboldt River, approximately 12 miles beyond the north boundary. The Osgood Mountains, which are only about 24 miles long, commence at the Humboldt River about 2 miles south of the south boundary and continue through the southwest, central, and northeast parts of the quadrangle. The main part of the range terminates about 2 miles beyond the north boundary, and some low hills extend about 8½ miles northeast of the northeast corner of the quadrangle. The Hot Springs Range is flanked on the west by Paradise Valley, which drains

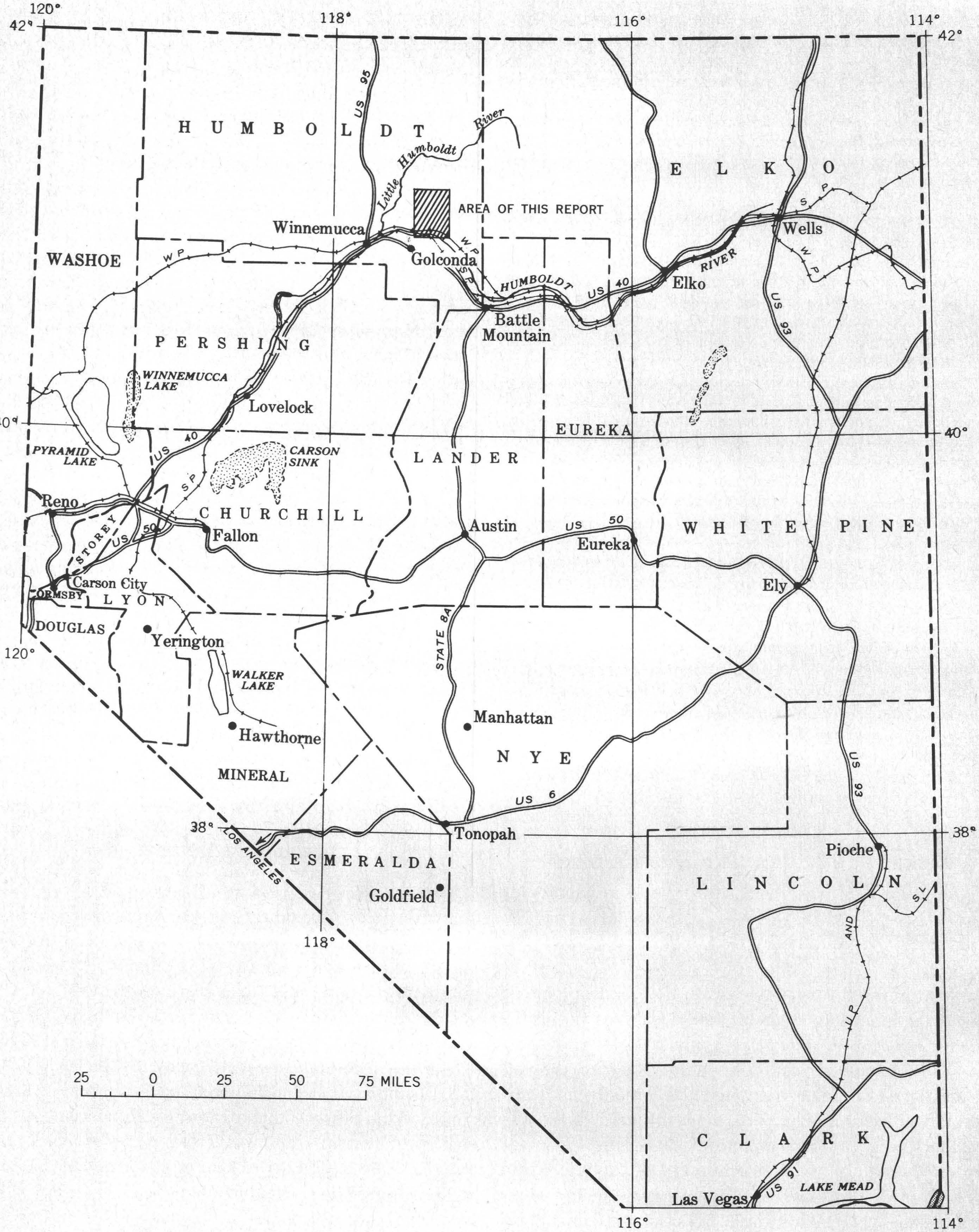


FIGURE 1.—Index map of Nevada showing the location of the Osgood Mountains quadrangle.

to the Little Humboldt River. Eden Valley lies between the Hot Springs Range and the Osgood Mountains; on the topographic map of the quadrangle only the northern part, which drains northward through Eden Creek to the Little Humboldt River, is shown as Eden Valley, but the name is also applied locally to the southern part that drains southward to Humboldt River. The Osgood Mountains are bounded on the east by the broad alluviated valleys of Kelly Creek and the Humboldt River. The main drainage channel of northern Nevada, the westward-flowing Humboldt River, is slightly south of the quadrangle, except for a short loop that enters the extreme southern part near Comus railroad siding. West of Comus the river enters a narrow valley which separates the Osgood Mountains from the Edna Mountains to the southeast in the Golconda quadrangle.

The general altitude of the valleys that border the ranges is from 4,500 to 4,700 feet. The crest of the Osgood Mountains gradually increases in altitude from approximately 5,750 feet at the south end to more than 8,000 feet in the central and northern parts, and declines to 6,000 feet at the north boundary of the quadrangle. The Hot Springs Range is somewhat lower, ranging from a general crestline altitude on the order of 6,750 feet in the northern part to around 6,000 feet in the southern part. Average relief of the Osgood Mountains ranges from 1,250 feet to 3,500 feet above the basins, that of the Hot Springs Range is from 1,500 feet to 2,250 feet. Adam Peak, in the central Osgood Mountains, is the highest point in the quadrangle; its altitude is 8,678 feet. The lowest part is in the southwest corner, where the alluvium-filled valley is about 4,350 feet above sea level. Maximum relief, therefore, is more than 4,300 feet.

The topography of the Osgood Mountains is more rugged and less uniform than that of the Hot Springs Range. In the Hot Springs Range rather uniform dissection has resulted in topography of monotonous aspect. Many short steep-sided canyons, each with a multitude of steeper subsidiary ravines and gullies and narrow smooth-surfaced divides, lead eastward or westward from the narrow north-trending crest to the alluvial aprons at the base of the range. The crest of the range east of Dutch Flat and extending southward for about 2 miles is a rather broad surface of gentle relief; but in general the canyons, ravines, and gullies and the slopes between them resemble their counterparts elsewhere in the range, and there is a general lack of eye-catching forms. The Osgood Mountains, on the other hand, have a considerably varied topography. In the southern part of the range the summit areas are commonly gently rolling surfaces which drop off abruptly to steep slopes below. The central and north-central parts of the range have high narrow ridges, long relatively smooth steep slopes, some bold clifflike features, and several large valleys, amphitheaterlike at their heads, narrow and steeper toward the range front. The north end of the range is thoroughly dissected by many short ravines and gullies. The east front is fairly straight, and in the northern part several of the east-trending ridges are terminated with subtriangular facets; in the southern and central parts the mountain front is fairly steep but descends gradually into gentler pediment slopes veneered with gravel that grades into the valley alluvium. The west front is considerably more irregular and is marked by several reentrants and mountain spurs that project into the valley.

CLIMATE AND VEGETATION

Rainfall records for Winnemucca, Nev. (table 1), show that the annual rainfall averages 8.75 inches. Most of the precipitation is in the winter and spring, chiefly as snow during the winter and early spring months. Occasional thunderstorms bring some rain during the summer.

The mean annual temperature is 49.1°F (table 1). Summer days are warm and the nights cool. The winters are cold, the average temperature during December and January being below freezing.

The Hot Springs Range and the Osgood Mountains are almost devoid of trees, except for a few aspens in the upper parts of some canyons. The most conspicuous vegetation is "sagebrush" and sparse grasses which are food for the range cattle. During the spring small flowering plants of many kinds are plentiful.

TABLE 1.—*Precipitation and temperature data, Winnemucca, Nev., to 1957*

[U.S. Weather Bureau, Annual Summary 1957, v. 72, no. 13, p. 153-154]

	Average temperature (° F)	Average precipitation (inches)
January	27. 8	0. 96
February	34. 5	1. 01
March	39. 4	. 86
April	46. 8	. 73
May	55. 9	. 84
June	64. 0	. 79
July	74. 2	. 31
August	69. 7	. 18
September	59. 9	. 34
October	48. 6	. 79
November	37. 6	. 84
December	30. 0	1. 00
Annual	49. 1	8. 75
Years of record	77	87

PREVIOUS WORK

No detailed geologic mapping was done in the Osgood Mountains quadrangle until 1946 when S. W. Hobbs mapped the northern part of the Osgood Mountains as part of a study of contact metamorphism and tungsten deposits adjacent to the granodiorite stock.[1] Previously Hess and Larsen (1921, p. 300–304) briefly studied some of the tungsten deposits in the northern part of the Osgood Mountains, and Hardy (1941) published a short description of the geology of the Getchell gold deposit. The Dutch Flat placer deposit on the west side of the Hot Springs Range was first described by Smith and Vanderburg (1932). The quicksilver deposits in the Hot Springs Range were first described by Bailey and Phoenix in 1944.

Eugene Callaghan and C. J. Vitaliano prepared planetable maps of some of the tungsten properties for the U.S. Geological Survey in 1940 as part of the Strategic Minerals Program; and during the summers of 1943, 1944, and 1945, Hobbs and S. E. Clabaugh continued the study with revisions of Callaghan's and Vitaliano's maps and preparation of maps of the other deposits. The result of their work was published in 1946 (Hobbs and Clabaugh, 1946). Following World War II a detailed study of the gold deposit at the Getchell mine was made by Peter Joralemon. The results of his field and laboratory investigations are contained in a thesis [2] and a published article (Joralemon, 1951).

Mapping of adjoining areas by others greatly facilitated work in the Osgood Mountains quadrangle. Geologic maps accompanied by brief texts have been published on the Winnemucca quadrangle (Ferguson, Muller, and Roberts, 1951) southwest of the Osgood Mountains quadrangle, and the Golconda quadrangle (Ferguson, Roberts, and Muller, 1952) to the south and southeast.

PRESENT WORK AND ACKNOWLEDGMENTS

Work began in the field in July 1951 and continued during the summer months until August 1955. The areal geologic mapping was done on aerial photographs enlarged 2 × and was transferred by inspection each day to the topographic map. The original compilation was at a scale of 1:40,000 and was replotted at 1:48,000 for reduction to the publication scale of 1:62,500. Planetable maps were made of the surface workings of all the important tungsten deposits and adjoining areas, because maps of the same areas prepared earlier by Hobbs and Clabaugh (1946) were outdated by later intensive mining activity. The underground geology of the tungsten deposits was plotted on level maps furnished by the companies. Geologic maps of the quicksilver and gold deposits in the Hot Springs Range were made by tape-and-compass methods.

Most of the geologic mapping and planetable work were done by Hotz and Willden. Assisting in the areal mapping for short periods at various times were H. R. Craig, D. C. Laub, F. R. Shawe, and G. C. Simmons. Ing. A. Tellez, R., assisted in the planetable work during part of one season. The underground mapping of the tungsten deposits was done by the senior author with the help of his son, R. P. Hotz.

The writers acknowledge the help and cooperation given by Royce A. Hardy, Keith Kunze, and W. J. Newman of the Getchell mine, and by Harry J. Trollope, engineer of the Riley mine (Union Carbide Nuclear Co.). The friendliness of the ranchers in the area did much to make the work pleasant and facilitated it in many ways. We are especially grateful to Mr. and Mrs. Ralph Smith, Mr. and Mrs. William Stevens, and Mr. and Mrs. Gene Christiansen for many courtesies received from them.

In the field we profited greatly from visits and discussions with colleagues whose wider knowledge of Nevada geology supplemented our relatively meager experience in the region. H. G. Ferguson, whose pioneer work serves as a basis for most of the geological knowledge of northern Nevada, was often a helpful and inspiring visitor. We are indebted for paleontologic assistance, both in the field and in the office, to A. R. Palmer on the Cambrian faunas; to Josiah Bridge and R. J. Ross on Ordovician faunas; and to Helen Duncan, Lloyd Henbest, MacKenzie Gordon, Jr., and James Steele Williams on the Carboniferous faunas.

ROCK UNITS

Sedimentary rocks exposed in the Osgood Mountains quadrangle include strata of Cambrian(?), Cambrian, Ordovician, Mississippian, Pennsylvanian, Permian, Tertiary, and Quaternary age. The complete stratigraphic succession is nowhere exposed; in fact most of the units mapped in the Osgood Mountains do not occur in the Hot Springs Range. Nondeposition or erosion have left gaps in the record, and thrust faults have brought into contact rocks that were deposited in different areas and some that are of different ages.

ROCKS OF CAMBRIAN(?) AND CAMBRIAN AGE

Four units of Cambrian and one of possible Cambrian age are exposed in the Osgood Mountains quadrangle (pl. 1). Of these, the Osgood Mountain quartzite, the Twin Canyon member of the Osgood Mountain

[1] Hobbs, S. W., 1948, Geology of the northern part of the Osgood Mountains, Humboldt County, Nevada: Yale Univ. unpublished doctoral thesis.

[2] Joralemon, Peter, 1949, The occurrence of gold at the Getchell mine, Nevada: Harvard Univ. unpublished doctoral thesis.

quartzite, and the Preble formation are exposed only in the Osgood Mountains. The Harmony formation is exposed in both the Osgood Mountains and the Hot Springs Range. The Paradise Valley chert is found only in the Hot Springs Range.

The unfossiliferous Osgood Mountain quartzite is the oldest unit exposed in the area. It is overlain by the Preble formation, which contains Middle and Upper Cambrian fossils. The Paradise Valley chert of Late Cambrian age is the next youngest unit, but its relations to the Preble formation are not known. The Paradise Valley chert is overlain by the Harmony formation, which contains fossils of Late Cambrian age. In the Osgood Mountains the Harmony formation is found only in the upper plate of a thrust, so its stratigraphic relations with units other than the Paradise Valley chert are unknown.

OSGOOD MOUNTAIN QUARTZITE

The Osgood Mountain quartzite was named by Ferguson, Muller, and Roberts (1951) for exposures at the south end of the Osgood Mountains in the Golconda quadrangle. This unit, which is composed predominantly of relatively pure quartzite, is widely exposed in the southern half of the Osgood Mountains. It contains in its uppermost part a discontinuous member characterized by interbedded shale and impure quartzite which has been mapped separately and called the Twin Canyon member. The Twin Canyon member represents beds apparently transitional between the Osgood Mountain quartzite and the overlying Preble formation.

DISTRIBUTION

The Osgood Mountain quartzite is extensively exposed in the central and south-central part of the Osgood Mountains; it extends southwestward for about 9 miles from the latitude of Goughs Canyon and Hogshead Canyon to the southern part of the range, where it is covered by volcanic rocks of Tertiary age. Younger Paleozoic rocks overlap the quartzite on the north. In the south-central part of the quadrangle the quartzite is continuous across the range to the valleys on either side. South of the quadrangle, the formation is exposed west of Emigrant Canyon at the south end of the Osgood Mountains (Ferguson, Roberts, and Muller, 1952), and in discontinuous areas along the east side of the Sonoma Range (Ferguson, Muller, and Roberts, 1951).

The Twin Canyon member of the Osgood Mountain quartzite, here named from typical exposures in Twin Canyon, SW¼ sec. 35, T. 38 N., R. 41 E., crops out in two narrow elongated belts on the east side of the Osgood Mountains. One belt, which is clearly lenticular, occurs in the southern one-fourth of the quadrangle in the low hills northwest of Lone Butte. This belt has a maximum exposed width of approximately 0.8 mile and a length of about 3.5 miles. It pinches out to the northeast along its strike. Its apparently abrupt termination to the southwest probably is due mostly to structural complications caused by faulting and folding. The member is exposed in a second belt on the southeast side of the main ridge of the Osgood Mountains, in the east-central part of the quadrangle. This belt is more than 3.5 miles long and extends from SW¼ sec. 10, T. 37 N., R. 41 E. to Hogshead Canyon in SW¼ sec. 25, T. 38 N., R. 41 E. Its exposed width, which is controlled over most of its length by a reverse fault along its east boundary, is nowhere more than about 0.4 mile. The belt pinches out gradually at its southwest end; it is overlapped by the Pennsylvanian Battle formation on the north.

LITHOLOGY

The bulk of the Osgood Mountain quartzite is composed of white, gray, pale-greenish-gray, pale-brown, and purplish-brown medium- to thick-bedded quartzite. Locally some pure white thin-bedded quartzite is exposed. Beds generally range in thickness from 1 to 10 feet, though some beds or lenses may be as much as 50 feet in thickness, and some strata in the thinly bedded varieties may be only a few inches thick. Crossbedding is characteristic of the quartzite in many exposures. The rocks are commonly massive and stratification is obscure, but many beds are separated by partings, a fraction of an inch to 1 or 2 inches thick, which are composed of platy fine-grained greenish-gray to light-brown micaceous quartzite and silty(?) quartzite. On the parting planes of some specimens sericitic mica is so abundant that the rock has a phyllitic aspect. Where platy partings are absent, bedding in the quartzite may be expressed by faint color variations or minor differences in lithology.

Most of the quartzite is fine to medium grained and very uniform in composition. The quartz grains are subrounded to rounded and generally are well sorted, although in places a few noticeably larger grains are scattered through an even-grained finer matrix. In some specimens individual grains are readily visible with a hand lens; in others the grains are closely packed and indistinguishable from the interstitial siliceous cement, and freshly broken surfaces have a vitreous appearance. Here and there pebbly layers, a fraction of an inch to a few inches thick, are interbedded with otherwise uniformly even grained rock. The pebbles are white quartz, as much as one-fourth of an inch in diameter, set in a fine-grained sandy matrix.

Most of the quartzite is light colored, but some beds are a distinctive dusky reddish purple owing to hematite in the matrix. Some of the light quartzite is traversed by purplish seams of hematite, and in a few places it contains small reddish-purple concentrations of hematite one-fourth of an inch in diameter.

Microscopic examination shows that in typical specimens of quartzite more than 90 percent of the grains are quartz; grains of chert and feldspar are rare. Other primary constituents are limited to a few detrital grains of "heavy" minerals including zircon and tourmaline most commonly, and some sphene. Most of the sphene has been altered to white opaque leucoxene(?). An uncommon variety of quartzite containing a significant amount of feldspar was collected near Soldiers Pass (center NW¼ sec. 20, T. 37 N., R. 41 E.). Approximately 10 percent of the grains is feldspar, of which orthoclase is the most common species, albite is less common, and grains of microcline are rare. The feldspar grains are partly replaced by sericite at the contact with quartz, and also in the body of the mineral. In addition the rock contains some flakes of muscovite. Most grains are rounded to subrounded, and the larger quartz fragments are well rounded; the smallest grains are subrounded and a few are subangular. The "heavy" accessory mineral grains are rounded to well rounded. The clastic grains are firmly bonded by secondary quartz, which is in optical continuity with the quartz grains, so that in many specimens the original grains are impossible or difficult to distinguish and the rock appears to be an interlocking mosaic of quartz. Grain outlines are made visible in some specimens by a film of impurities such as sericite or hematite on the interface between the quartz grains and the silica of the matrix.

Sericite and small amounts of chlorite are common in some specimens and make up several percent of the rock. The sericitic and chloritic material occurs interstitially to the quartz grains. Sericite replaces quartz along some grain boundaries, indicating that the sericite is authigenic (Pettijohn, 1957, p. 305, fig. 77). Hematite is an important constituent in the matrix of some specimens; it forms films on the interface between clastic grains of quartz and the secondary quartz cement or is concentrated in small areas that appear megascopically as purplish spots.

The thin shaly or phyllitic partings between quartzite beds are composed of angular to subrounded quartz grains loosely packed in a matrix of sericite, chlorite, and silica, which probably was originally silty material. In addition to quartz grains there are a few rounded grains of tourmaline and zircon.

LITHOLOGY OF THE TWIN CANYON MEMBER

The Twin Canyon member of the Osgood Mountain quartzite has a greater proportion of silty and shaly material than the rest of the formation. Shaly material in the formation below the Twin Canyon member is represented only by thin partings between quartzite beds, but in the Twin Canyon member the shale units are 100 or more feet thick. Beds of quartzite alternate with phyllitic shale; the beds of quartzite are thicker and more abundant in the lower part of the member, whereas shale predominates and the quartzite beds are thin in the upper part.

In general the rocks are darker than the rest of the Osgood Mountain quartzite. The shales are dark greenish gray to gray and the interbedded quartzite ranges through greenish gray, light brown, and dusky red purple; some beds of quartzite are white to light gray. Dark purple quartzite is abundant in the northern belt of the Twin Canyon member.

The Twin Canyon member is more prominently sheared than the rest of the Osgood Mountain quartzite. The less competent fine-grained units are almost everywhere sheared and are somewhat phyllitic. In many places the shaly units are badly contorted.

Most of the quartzite in the Twin Canyon member is impure, though the unit contains some beds of clean quartzite. Much of the impure quartzite might be more properly classed as subgraywacke or quartzose subgraywacke (Pettijohn, 1957, p. 316–320).

The impure quartzite is fine to medium grained and massively bedded; crossbedded structures are not common. Even in hand specimens the impurity of the rock is apparent. Microscopic examination shows that some specimens contain as much as 48 percent sericitic and chloritic matrix material, which is greatly in excess of the amount of cementing silica. The fragments are in general no more angular than the grains in the purer kinds of quartzite, but the whole bulk of material has not been as cleanly washed, so that detrital grains are comparatively loosely packed in the matrix. Quartz is the principal clastic constituent; plagioclase, potassium feldspar, and a few flakes of muscovite occur in minor amounts. Some lithic fragments, principally chert but also a few pieces of shale, are commonly present. The "heavy" mineral assemblage—zircon, tourmaline, sphene, leucoxene, rutile, and magnetite—is the same as in the quartzite in the rest of the formation but is more abundant. Besides sericite and chlorite, the matrix of many specimens contains scattered irregular grains of hematite and a little magnetite. Probably the sericite and chlorite were mainly derived from the reconstitution of an original "clay" matrix, but in part they may be detrital.

The Twin Canyon member contains some purplish quartzite like that in the main part of the formation, and there are also some beds of very dark purplish-black quartzite that contain as much as 15 percent hematite in the matrix. Hematite occurs as fine-grained structureless interstitial material, as oolitic forms in the matrix, and as shells around quartz grains. The principal bonding agent, however, is silica. The rock is well sorted and contains only a little sericite in the matrix; detrital zircon and tourmaline are very rare.

Some of the fine-grained rock interbedded with the quartzite is shale or silty shale, or their phyllitic equivalents, but much of it is very fine to fine grained silty sandstone that has platy parting parallel to the bedding. Fine flakes of sericite are plainly visible on parting surfaces of most specimens. Commonly, microscopic examination shows the rock to be composed of alternating very fine grained sandstone and chlorite-biotite-sericite layers. The sandy layers contain, in addition to quartz, considerable chlorite, sericite, occasional pyrite and limonite, and some fragments of shale and chert. Grains of tourmaline and zircon can also be recognized. The biotite, which is a common constituent of the fine-grained silty sandstone and phyllite, may be partly detrital, though most of it appears to have formed later than the rest of the minerals.

STRATIGRAPHY AND THICKNESS

The Osgood Mountain quartzite is the oldest formation exposed in this part of northern Nevada. The formation is in conformable succession with the overlying Preble formation. The top of the Osgood Mountain quartzite is drawn at the top of the last quartzite above which the amount of shale exceeds the amount of quartzite. At most places this contact is easily established, but where the Twin Canyon member of the Osgood Mountain quartzite is present the position of the contact is less certain.

The Twin Canyon member is transitional between lithology of the Osgood Mountain quartzite and that of the overlying Preble formation, a predominantly shale and limestone unit. The lower contact of the member commonly is abrupt and conformable, but in some places it seems to be gradational over a distance of a few feet. This contact is drawn at the base of the first prominent shale bed, above which there is an alternating succession of shale and impure quartzite. The upper contact with the Preble formation is arbitrarily drawn at the top of the last quartzite bed, above which the section is predominantly shale. Where the Twin Canyon member is missing, the upper contact of the clean Osgood Mountain quartzite with the shale of the Preble formation seems to be abrupt from the general field relations, but exposures are generally poor along the contact.

The thickness of the Osgood Mountain quartzite is not known because the base of the formation is not exposed. The exposed thickness can be estimated, but without much certainty because the absence of distinctive lithologic units does not permit the working out of the structural complexities within the formation. Ferguson, Roberts, and Muller (1952) estimated more than 5,000 feet of beds in the Golconda quadrangle. A similar thickness of beds is exposed in the Osgood Mountains quadrangle including the Twin Canyon member, which is 1,000 to 1,500 feet thick in most places and has a maximum thickness of 2,500 feet.

AGE AND CORRELATION

No fossils have been found in the Osgood Mountain quartzite, but the formation is in apparently continuous stratigraphic succession with the overlying Preble formation, which contains organic remains of Middle and Late Cambrian age. The lithology closely resembles that of the Prospect Mountain quartzite and other similar quartzites found at the bottom of Cambrian sections in other parts of the Great Basin. The Twin Canyon member, which indicates the beginning of an important change in conditions of sedimentation, was possibly similar in origin to the Pioche shale, which overlies the Prospect Mountain quartzite at some places.

The Prospect Mountain quartzite is generally regarded as Lower Cambrian, and some believe that it may be in part Precambrian (Wheeler, 1943, 1948). At most places where the Prospect Mountain quartzite occurs, it underlies beds in which Lower Cambrian fossils are known. No remains older than Middle Cambrian have been discovered in the Preble formation above the Osgood Mountain quartzite, but because of its position beneath fossiliferous strata of Cambrian age, and the presence between these beds of strata apparently representing nearly continuous sedimentary deposition, we tentatively regard the formation as Early to Middle Cambrian in age, although it is officially considered to be Cambrian (?) in age.

CONDITIONS OF DEPOSITION

The siliceous sediments of the Osgood Mountain quartzite and its equivalents elsewhere in the Great Basin represent the initial deposits of a widespread system of strata of Cambrian age laid down in a gradually encroaching sea (Deiss, 1941, p. 1089–1090, 1098; Wheeler, 1943, p. 1808–1811). These first accumulations represent detritus derived from a landmass that had been subject to a long period of subaerial decay.

The purity of the quartzite, its thick-bedded and persistently cross-stratified character through several thousand feet of section indicate that it was deposited in a gradually sinking basin under persistently shallow water conditions where wave action was an effective sorting mechanism. Probably it represents deposits of sand in a shelf environment where offshore bars and spits were constructed not far from the gradually transgressing strand line. Under conditions like these the quartzite would almost certainly not be time-equivalent from one area to another.

The beds of impure subgraywacke and interbedded shale that constitute the Twin Canyon member at the top of the Osgood Mountain quartzite represent a change in conditions of deposition. Possibly these sediments accumulated without much washing and reworking, although the occasional relatively thin beds of quartzite suggest that from time to time, but with decreasing frequency, the environment reverted to conditions such that sorting by wave action was effective. Possibly this change reflects an increase in rate of sedimentation brought about by accelerated erosion or reflects relatively rapid subsidence of the basin of deposition, or both. The common occurrence of ferruginous quartzite in the section is possibly related to oxidizing conditions on parts of the sea floor that permitted accumulation of iron oxide along with the detrital material. James (1954, p. 272) has suggested that hematite may be "* * * deposited as hydrated ferric oxide in shallow, well-aerated waters."

PREBLE FORMATION

The name Preble formation was given by Ferguson, Muller, and Roberts (1951) to an argillaceous and calcareous section overlying the Osgood Mountain quartzite in Emigrant Canyon near Preble Station, Golconda quadrangle. Rocks belonging to the formation have been followed northward along strike from the type locality into the Osgood Mountains quadrangle, where they are economically important because of the tungsten deposits that have been formed in the limestones adjacent to a granodiorite stock.

DISTRIBUTION

The formation is on the southeast flank of the Osgood Mountains from the southern border of the quadrangle to about Hogshead Canyon (pl. 1). North of Hogshead Canyon it occupies a continuous belt in the main part of the range, partly interrupted by a granodiorite stock. The unit extends to slightly beyond the northern boundary of the quadrangle. Rocks of the Preble formation are also exposed in a small rectangular area in the lower part of Goughs Canyon and in small irregular-shaped areas in Goughs and Perforate Canyons on the west side of the range. The Preble formation is not present in the Hot Springs Range.

The limestones within the formation have been mapped separately because of their importance as host rocks for the tungsten deposits. It can be said, in general, that the limestones of the Preble formation occur in a belt a mile or so wide in the easterly, stratigraphically higher part of the Preble, and a belt of phyllitic shale lies between the limestone belt and the underlying Osgood Mountain quartzite. Limestone is also the predominant rock type in a belt about one-fourth of a mile wide and 1½ miles long at the head of Farrel Canyon. Relatively thin discontinuous and lenticular limestone beds are found in the western part of the Preble outcrop at the north end of the quadrangle, in a section of predominantly pelitic rocks. The spatial and stratigraphic relations between the limestone in the southern and eastern part of the range and that in the north end of the range are not known because of structural complexities and metamorphism.

LITHOLOGY

Along the southeastern side of the Osgood Mountains, the Preble formation is dominantly phyllitic shale in its lower part and interbedded limestone and shale in its middle and upper parts. In the northern part of the quadrangle, the relative stratigraphic position of the beds is uncertain, but shale is more abundant than limestone except at the head of Farrel Canyon. The limestone along the northwest side of the Osgood Mountains occurs as long thin lenses within the shale; in Farrel Canyon the combined width of the limestone lenses exceeds that of the interbedded shale.

The phyllitic shale is most commonly greenish gray or yellowish gray; less commonly it is yellowish brown. Weathered surfaces are lighter in general. A slaty cleavage has been developed in much of the formation, as a rule generally parallel to the bedding, though in places it cuts the bedding at a fairly large angle. Cleavage surfaces are generally coated with flakes of white or pale-colored mica. Some of the shale is calcareous, and some of it is rather siliceous and does not split readily, so that it breaks down on weathering into small rough-surfaced chips. In places where deformation has been severe, the phyllitic shale is intricately crinkled. Locally, strong deformation has converted the phyllitic shale to a phyllite in which microscopic folia and porphyroblasts of yellowish-green biotite have been formed in a quartz-plagioclase-sericite matrix. Retrograde metamorphism has caused partial replacement of the biotite, particularly the porphyroblasts, by chlorite. In some specimens the phyllitic cleavage direction is crossed at a marked angle by a later more widely spaced strain slip cleavage (Williams, Turner, and Gilbert, 1954, p. 213) that deforms the earlier formed mica flakes and reorients them subparallel to the later cleavage direction.

A chemical and a spectrographic analysis of shale from the Preble formation are given in table 2.

TABLE 2.—*Chemical and spectrographic analyses of shale and limestone from the Preble formation*

Chemical analyses [Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C; P. L. D. Ellmore, K. E. White, S. D. Botts, P. W. Scott, analysts, U.S. Geol. Survey]				Spectrographic analysis [Harry Bastron, analyst, U.S. Geol. Survey]	
	1	2	3		1
SiO_2	53.1	5.8	2.4	Cu	0.002
Al_2O_3	26.4	.80	.34	Pb	.004
Fe_2O_3	3.0	.43	.10	Mn	.09
FeO	5.3	.01	.02	Co	.002
MgO	1.7	.26	.42	Ni	.007
CaO	.19	52.7	54.9	Ga	.002
Na_2O	1.0	.07	.06	Cr	.01
K_2O	3.9	.12	.02	V	.007
TiO_2	.87	.04	.02	Sc	.008
P_2O_5	.14	.09	.10	La	.02
MnO	.11	.00	.01	Ti	.4
H_2O	5.1	.01	.02	Zr	.009
CO_2	.05	40.1	42.6	Be	.0004
				Sr	.004
Sum	101	100	101	Ba	.1
Sp. G. (lump)	2.81	2.58	2.34	B	.009
(powder)	2.86	2.71	2.70		

Looked for but not found: Ag, Au, Hg, Ru, Rh, Pd, Ir, Pt, Mo, Re, Ge, Sn, As, Sb, Bi, Zn, Cd, Tl, In, Y, Yb, Th, Nb, U, P.
The above results have an overall accuracy of ±15 percent.

1. Phyllite from NE¼ sec. 36, T. 38 N., R. 41 E.
2. Limestone from NE¼ sec. 1, T. 37 N., R. 41 E.
3. Limestone from NE¼ sec. 1, T. 37 N., R. 41 E.

Because of its tendency to disintegrate rapidly the phyllitic shale does not form prominent outcrops, and areas underlain by shale have a relatively subdued topography. It is characteristic of these areas, however, that the overburden is not thick and the bedrock is exposed whenever there is a small rill or gully, or where a slope is oversteepened.

A few beds of quartzite occur within the formation; however, they are almost entirely restricted to the lower shaly part. These beds, which stand up as bold outcrops because of their greater hardness, are relatively thin—mostly no more than a few tens of feet thick—and probably lenticular, but they are persistent units that can usually be followed for many hundreds of feet. Where the formation grades downward into the Twin Canyon member of the Osgood Mountain quartzite, beds of quartzite gradually become more abundant than shale.

Limestone in the Preble formation is of several kinds, but all of it is dark bluish-gray on weathered and freshly broken surfaces. Most of the limestone is fairly well bedded, though in places it is massive. Some of it has a characteristic rhythmically bedded appearance that is not seen in limestone from the later Paleozoic formations in this area; beds of fine-grained limestone from ½ inch to 2 inches thick alternate with shaly partings that are mostly one-half an inch or less thick. Some of the limestone is platy and on weathering breaks down into small slabs that range from 1 to 3 inches in thickness. Much of the thicker bedded limestone is medium to coarsely crystalline and weathers to a rough surface. Some is cherty, with nodules and irregularly lenticular bodies of chert, 1 inch or so thick, scattered somewhat irregularly through the rock. Some beds are rather sandy, and some show oolitic structure. In places, particularly where the limestone has been recrystallized, the rock is cut by a network of white coarsely crystalline calcite veinlets. Chemical analysis of two specimens of limestone from the Preble formation are given in table 2.

STRATIGRAPHY AND THICKNESS

The Preble formation conformably overlies the Osgood Mountain quartzite. In some places the lower shales rest directly on the quartzite; in other places the Twin Canyon member of the Osgood Mountain quartzite grades upward into the Preble formation. The top of the Preble is not exposed in the quadrangle. On the geologic map of the Golconda quadrangle (Ferguson, Roberts, and Muller, 1952), the Comus formation of Ordovician age is shown overlying the Preble formation with a depositional contact; but on the basis of new data obtained from a reexamination of the area by us with Ferguson, we now believe that this contact is a high-angle reverse fault. The contact between the Preble and Comus formations has also been mapped as a fault in the eastern part of the Osgood Mountains quadrangle. Elsewhere in the area the Preble formation is incompletely exposed because of burial beneath younger rocks or truncation by faulting. On the basis of fossil evidence it is known that the Preble is older than the Paradise Valley and Harmony formations of Late Cambrian age, but there is no stratigraphic evidence for this relation because the formations occur in different parts of the quadrangle.

Detailed studies of the stratigraphy and measurements of thickness cannot be made satisfactorily in the Preble, because the beds are tightly folded and the lithologic units are not sufficiently distinctive to serve as markers in working out the structural complexities within the formation. Probably the structure is also complicated by faulting; there is some minor thrusting along contacts between the shale and limestone beds.

Some general lithologic subdivisions can be made, however, on the east side of the range. The lower part of the formation above the Osgood Mountain quartzite, or above the Twin Canyon member, is composed predominantly of phyllitic shale and siltstone with a few thin beds of quartzite and graywacke. Above this is a unit composed mostly of limestone with some interbedded and interfingering shale, and between the limestone unit and the Comus formation is another section of phyllitic shale. The lower shale probably ranges from 2,800 feet to 4,700 feet in thickness; the intermediate limestone unit may be from 1,300 feet to as much as 1,500 feet thick; and there may be as much as 1,500 feet of shale above the limestone. Following is a measured section of the predominantly limestone unit exposed on the ridge northeast of the lower part of Hogshead Canyon in the NE¼ sec. 1, T. 37 N., R. 41 E., and SE¼ sec. 36, T. 38 N., R. 41 E.:

	Thickness (ft)
Slaty shale, greenish-gray, very fissile	not measured
Limestone, blue-gray, crystalline, rough-surfaced, thick-bedded to massive; many white calcite veinlets and some sandy beds that contain organic fragments	250
Cherty limestone, brown-weathering, rather badly sheared, very rough surfaced; contains blobs and some veinlets of white calcite	200
Limestone, blue-gray, platy-weathering, thin-bedded; contains a little brown chert and some interbedded shaly limestone and greenish-gary phyllitic shale	225
Cherty limestone, blue-gray, rough-weathering, thick-bedded to massive	75
Limestone, bluish-gray, somewhat platy, slightly cherty	100
Platy limestone, bluish-gray; and some interbedded phyllitic shale	200
Quartzite, brown-weathering	10
Phyllitic shale, greenish-gray	50
Platy limestone, gray to purplish-gray; interbedded with calcareous shale	100
Phyllitic shale, greenish-gray; thin calcareous shale and thin beds of limestone	175
Cherty limestone, gray	110
Total thickness of limestone section	1,495
Phyllitic shale, greenish-gray; lenses of quartzite as much as 5 in. thick	not measured

Ferguson, Roberts, and Muller (1952) reported:

Thickness not determinable because of isoclinal folding and minor thrusting; may exceed 12,000 feet.

The present authors estimate that the section in the vicinity of Hogshead Canyon may be about 5,000 feet thick, but both the upper and the lower contacts are faults.

AGE AND CORRELATION

The Preble formation has yielded several collections of fossils which have been reported on by A. R. Palmer, of the U.S. Geological Survey (1953, 1954, written communication).

Four collections have been made from limestone beds on the southeast side of the Osgood mountains:

Collection No. USGS 3136–CO. East of center, SE¼NE¼ sec. 1, T. 37 N., R. 41 E.

Collection No. USGS, 1972–CO. Extreme NE¼SE¼SW¼ sec. 1, T. 37 N., R. 41 E.

Collection No. 1973–CO (field No. f54–W–7). On ridge in NE¼SW¼NW¼ sec. 12, T. 37 N., R. 41 E.

Collection No. USGS 1506–CO. 100 feet east of a prominent draw and about 100 feet above the main valley bottom, SE¼NW¼ sec. 12, T. 37 N., R. 41 E.

Collection 3136–CO had identifiable fossils only in the insoluble residue. Here, a scrap of acrotretid brachiopod that suggests *Acrothele* indicates a probable Cambrian age for the collection.

Collection 1972–CO has a few scraps of trilobite pygidia that suggest an *Ehmaniella*-like trilobite. Trilobites of this type are difficult to identify even when fairly good material is present. The collection is probably Cambrian and possibly Middle Cambrian in age.

Collection 1973–CO contains unusually well preserved Conchostracans, including one specimen possibly referable to *Aluta primordialis* (Linnarsson), a species known from near the Middle-Upper Cambrian boundary in Sweden. Scraps of unidentifiable silicified trilobites were present in the insoluble residue.

Collection 1506–CO was made from a single piece of limestone float that probably came from the same stratigraphic position as 1973–CO. Palmer says of this collection:

The fauna contains six genera of trilobites, inarticulate and articulate brachiopods, a snail and some interesting tentaculites-like problematica. This extends knowledge of the distribution of early Upper Cambrian pre-*Aphelaspis* fossiliferous rocks more than 200 miles westward.

The following trilobite genera are recognized:

Meteoraspis
Coosella?
Tricrepicephalus
Kingstonia
Maryvillia?
Pemphigaspis?

A new species of the snail genus *Strepsodiscus* previously known only from early Upper Cambrian beds of Colorado is also present.

The presence of *Meteorapis* and *Coosella* indicates a correlation to the lower part of the *Crepicephalus* zone of the standard Upper Cambrian faunal sequence.

The unit from which this collection came is equivalent in age to the Hamburg dolomite * * *.

Two collections of trilobites were made from thin limestone beds on the northwest side of the range north of Anderson Canyon:

Collection No. USGS 1372-CO. Northern part NW¼NW¼ sec. 30, T. 39 N., R. 42 E.

Kootenia sp.

Wimanella sp.

Collection No. USGS 1378-CO. Eastern part SE¼SE¼ sec. 25, T. 39 N., R. 41 E.

Kootenia? sp

According to Palmer these two collections

> contain trilobites characteristic of rocks of lower Middle Cambrian age. The trilobites * * * are probably older than those collected from the "Secret Canyon" in the Mount Lewis (Nevada) quadrangle.

The fossil determinations thus indicate that the Preble formation ranges from lower Middle Cambrian to lower Upper Cambrian. Ferguson, Roberts, and Muller (1952) reported finding linguloid brachiopods in limestone in the upper half of the Preble formation in Emigrant Canyon, Golconda quadrangle, which was the basis for assigning a Middle or Upper Cambrian age to the formation at the type locality.

According to the faunal data, the Preble exposed on the southeast side of the Osgood Mountains is younger than the rocks mapped as Preble on the northwest side of the range. Stratigraphic relations between the rocks in these two areas cannot be determined, but rocks on the northwest end of the range are predominantly shale, and the limestones are relatively thin and lenticular, so the section may be roughly equivalent to the shale in the lower part of the formation on the southeast side of the range.

Lithologically the Preble formation does not closely resemble the known Cambrian from other parts of Nevada. Faunally the limestone on the southeast side of the Osgood Mountains is correlated with the Hamburg dolomite of the Eureka district; and beds on the northwest side are probably somewhere below the Secret Canyon shale.

CONDITIONS OF DEPOSITION

The lower phyllitic shale of the Preble formation has been shown to grade down into the Twin Canyon member of the Osgood Mountain quartzite, which marks a change in conditions of deposition from a nearshore environment to one of deeper water farther from the land. The fine-grained clastic rocks in the lower part of the Preble, followed by a section in which limestone predominates, is in keeping with the general picture of a gradually encroaching sea in which deposition that began with the clean sands of the Osgood Mountain quartzite continued with essentially no interruption but with a gradual change to conditions of offshore deposition of mud and carbonate.

PARADISE VALLEY CHERT

The Paradise Valley chert is a new formation, here named for Paradise Valley, the large valley west and northwest of the Hot Springs Range.

DISTRIBUTION

The Paradise Valley chert is exposed along the west side of the Hot Springs Range in two narrow belts in the northwest corner of the quadrangle (pl. 1). The most extensive exposures are in secs. 27 and 28, T. 39 N., R. 40 E.; to the south the belts become narrow and disappear under alluvium in sec. 33, T. 39 N., R. 40 E.; these exposures do not extend much beyond the northern border of the quadrangle, but the formation is found farther north, about a mile north of Stewart Gap (Hot Springs Peak quadrangle).

In general, the formation is not well exposed because of heavy soil cover. Chert forms fairly prominent outcrops, but in many places it can be followed only by fragments in the soil; limestone and siliceous shale, which are interbedded with the chert, are generally poorly exposed.

LITHOLOGY

The Paradise Valley chert consists predominantly of chert, but it also contains minor amounts of siliceous shale and limestone. The chert is light to dark gray, light to dark brown, and black; it is well bedded but extensively fractured, so that it is difficult to get a large piece or fresh surface. Most of the beds are 2 inches to 18 inches thick, but some beds are as thin as one-fourth of an inch and others as thick as 5 feet. Individual chert beds are separated by partings of siliceous shale. The siliceous shale partings are light gray or light brown. Siliceous shale also occurs as a few thin beds; the shale of the thin beds is usually light gray or green. Limestone occurs in small lenses and thin beds which can be followed for only short distances. It is dense, fine to medium grained, and fairly pure to slightly shaly. Fossils are found in the medium-grained slightly shaly limestone lenses and beds.

The chert is composed of microcrystalline quartz and many impurities consisting mostly of minute colorless to faintly greenish rods and fibers with parallel alinement, which appear to be chloritic and sericitic material probably derived from original argillaceous material. In addition there is a little limonitic material, some of which is pseudomorphous after pyrite. A few scattered small ovoid bodies of fine-grained quartz may represent the remains of organisms. The rock is cut by a multitude of veinlets of fine-grained quartz. No textures or structures were observed that would be indicative of replacement of original tuffaceous material or limestone by silica.

Microscopic examination of the limestone shows that it is uneven grained, grain size ranging from 0.05 mm to 1 mm, possibly because the rock has been partly recrystallized. It contains many fragments of fossils and a very few detrital grains. The detrital grains identified in thin section and by a study of the insoluble residue of the limestone include quartz and a little quartzite and chert, and "heavy" minerals. The heavy-mineral suite consists of colorless zircon, blue and green tourmaline, abundant monazite and xenotime, muscovite, leucoxene, ilmenite, rutile, magnetite, and hematite. Pyrite, which floods the heavy concentrate, is an authigenic constituent.

STRATIGRAPHY AND THICKNESS

The base of the Paradise Valley chert is not exposed and its stratigraphic relation to the Osgood Mountain quartzite and the Preble formation is not known. The contact with the overlying Harmony formation is depositional, but exposures are not good enough to be sure whether the contact is conformable or not.

The full thickness of the Paradise Valley chert is not known because the base of the formation is not exposed. Estimates of the thickness are further complicated by uncertainty as to the degree of unconformity with the overlying Harmony formation and by poor exposures that make it difficult to detect repetition of beds by faulting or tight folding. The exposed thickness on the west side of the Hot Springs Range is estimated to be about 300 feet, and the thickness north of Stewart Gap is estimated to be about 500 feet.

AGE AND CORRELATION

The Late Cambrian age of the Paradise Valley chert is well established by fossils collected from two localities, one of this quadrangle and one in the Hot Springs Peak quadrangle.

Collections Nos. USGS 1370–CO, 1380–CO, 2901–CO. Ridge S½NE¼SE¼ sec. 28, T. 39 N., R. 40 E., Osgood Mountains quadrangle.

Collection No. USGS 1374–CO. Extreme NE¼NW¼ sec. 31, T., 40 N., R. 41 E. (Unsurveyed), Hot Springs Peak quadrangle.

A. R. Palmer reports (1960, written communication):

collections 1370–CO, 1380–CO, and 2901–CO were made at the same locality by three different collectors at three different times. Because of the structural complexity and poor exposures, different limestone lenses including in part different faunas were sampled.

Collection 1370–CO and most of 2901–CO contain the following trilobites:

Aphelaspis sp.
Agnostus inexpectans Kobayashi
Glyptagnostus reticulatus reticulatus (Angelin)
Olenaspella regularis Palmer
Pseudagnostus sp.
Homagnostus sp.

Collection 1380–CO and one piece of collection 2901–CO indicate the presence of a slightly older fauna containing:

Kingstonia cf *K. spicata* Lochman
Crepicephalus sp.
Deiracephalus sp.

Collection 1374–CO contains the following trilobites.

Glyptagnostus reticulatus angelini (Resser)
Agnostus cf *A. inexpectans* Kobayashi
Homagnostus sp.
Cheilocephalus sp.
Aphelaspis sp.
Olenaspella separata Palmer

The *Glyptagnostus*-bearing faunas, in collection 1370–CO and 1374–CO have been found in sections near Cherry Creek and McGill, Nevada, respectively. They characterize the lower beds of a nonresistant unit of interbedded limestones, siltstones, and shales that probably represents the Dunderberg shale. These faunas are older, however, than the oldest fossiliferous beds in the Dunderberg shale at Eureka. Trilobite assemblages equivalent to the oldest faunas of the type Dunderberg shale are found in the McGill and Cherry Creek sections about 200 to 250 feet above the *Glyptagnostus*-bearing beds.

HARMONY FORMATION

The Harmony formation is a distinctive lithologic unit that was first described by Ferguson, Muller, and Roberts (1951) in the Winnemucca quadrangle. They tentatively assigned the formation to the Mississippian(?) on the basis of lithologic evidence and structural relations, an age assignment that was followed by Roberts (1951) in the Antler Peak quadrangle, where the Harmony is also exposed. Fossil collections from the Hot Springs Range and the Osgood Mountains have shown that the Harmony formation is of latest Cambrian age.

DISTRIBUTION

The Harmony formation is rather widely distributed in north-central Nevada. The type locality is Harmony Canyon in the northern part of the Winnemucca quadrangle, and rocks belonging to the formation are abundantly exposed in the Sonoma Range (Ferguson, Muller, and Roberts, 1951). The formation is also found on Battle Mountain 30 miles east of the Sonoma Range, and Gilluly has observed breccia blocks of Harmony in the Mount Lewis quadrangle 15 miles southeast of Battle Mountain. Rocks of the same distinctive lithology make up the bulk of the Hot Springs Range in the Osgood Mountains quadrangle; they can be followed northward into the Hot Springs Peak quadrangle, where they become covered by volcanic rocks of Tertiary age of alluvium south of the Little Humboldt River.

Within the Osgood Mountains quadrangle, the Harmony formation makes up the bulk of the Hot Springs

Range (pl. 1), but in the Osgood Mountains it is confined to a rather narrow belt between two thrust faults on the west side of the range, extending for about 3 miles from Goughs Canyon on the south to the ridge between the East Fork of Eden Creek and Cave Canyon on the north.

LITHOLOGY

The Harmony formation is composed predominantly of feldspathic sandstone with some interbedded shale and minor amounts of pebble conglomerate and clastic limestone. A basal shale member 75 to 150 feet thick is included within the formation in this quadrangle, though it has not been found elsewhere.

Perhaps the most characteristic lithologic feature of the Harmony formation is the sandstone, which is very distinctive throughout the section. It is light brown to olive green and ranges in grain size from medium to coarse, some of it having a distinctly gritty appearance. Many specimens, especially the coarser grained and gritty varieties, are characterized by an abundance of prominent bluish-gray to milky-white grains of quartz, light-gray to milky-white grains of feldspar, and less abundant fairly large flakes of muscovite in a considerably finer grained greenish-brown matrix. The sandstone is generally thick bedded to massive—though some thin-bedded units are seen—and the beds are commonly separated by thin partings of shale half an inch to an inch thick. Many of the beds show graded bedding, which in some places is easily seen and in other places is difficult to recognize. The graded beds range in thickness from a few inches to several feet and are continuous along strike for as much as 1,000 feet. The lowermost part of some of the graded beds contain shale fragments; small-scale scour-and-fill structures, usually not very well developed, can be seen in the underlying shaly beds. Crossbedding was not seen in the sandstone.

Microscopic examination of thin sections of the sandstone show that most specimens are poorly sorted and most of the grains are subangular to subrounded; a small percentage are angular or rounded fragments (fig. 2). Commonly, many of the grains show various degrees of etching, and some of the quartz grains have secondary overgrowths. The grains are tightly bonded by a matrix that is largely very fine sercite and chlorite but is also composed of very fine fragments of the same minerals that constitute the sand-size fraction of the rock. Sericite and chlorite probably formed from original clay, though some of the mica may be detrital. Some varieties of sandstone have a calcareous cement or contain secondary quartz in the matrix.

Quartz and feldspar are the predominant mineral grains. Quartz accounts for from 70 percent to 80 percent of the grains, and the total feldspar ranges from approximately 10 percent to 18 percent. The quartz grains, including the coarsest sizes, are mostly fragments of single individuals, but some are composite grains with sutured interlocking borders; some are crowded with fine dust and liquid inclusions; nearly all grains are strained and show undulatory extinction. Some slides contain recognizable fragments of quartzite. The feldspar grains are mostly completely fresh. Both plagioclase and potassium feldspar are present; the ratio of plagioclase to potassium feldspar is about 1 to 2. The plagioclase is a sodic variety, judging by its index of refraction and small extinction angles; orthoclase—commonly perthitic—and microcline are the potassium feldspars.

The remainder of the detrital grains include apatite, zircon, sphene, leucoxene probably derived from sphene and ilmenite, tourmaline, epidote, magnetite, and rutile. Apatite is perhaps the most abundant, but two kinds of zircon are also present, a colorless and a pink variety. The colorless variety occurs as euhedral crystals and as well-rounded grains. The pink zircon is the more abundant and shows hardly any effects of abrasion, being in short prismatic crystals with pyramidal terminations. Pink zircon from sandstone of the Harmony formation in the Antler Peak quadrangle has been determined to be 958 million years old (Jaffe and others, 1959, p. 130).

Additional "heavy" mineral species are undoubtedly present but most of the sandstone is very difficult to break down to sizes that permit satisfactory separation. A specimen of calcareous sandstone was disintegrated with acid, however, and the following additional minerals were observed: monazite, dumortierite, cassiterite, ilmenite, pyrite, pyrrhotite. The grains of monazite are both rounded and euhedral; the dumortierite and cassiterite are euhedral or angular.

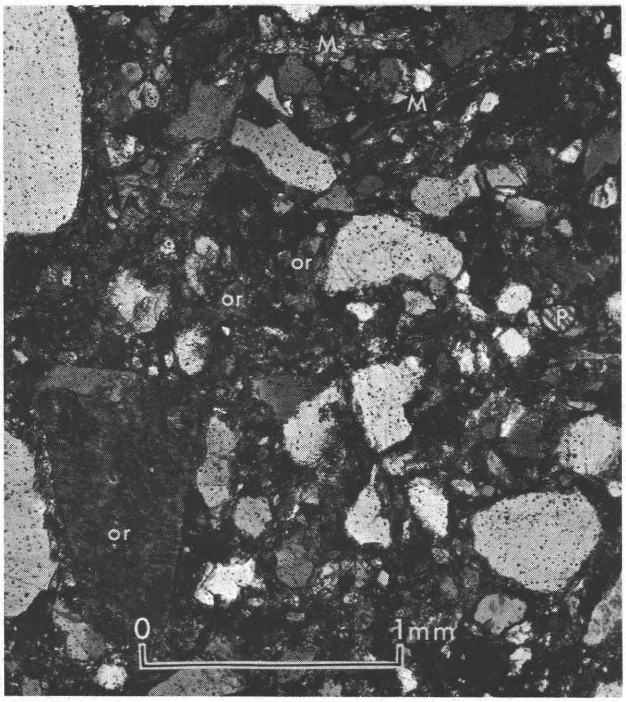
Modal composition of some typical sandstone specimen are given in table 3.

TABLE 3.—*Modes of the Harmony formation*

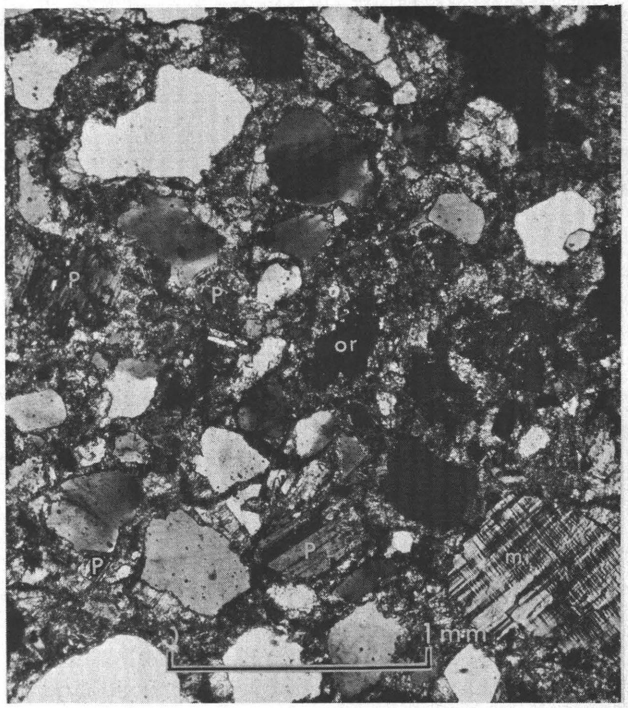
[Volume percent. Tr, trace]

	1	2	3	4
Quartz	72	78	73	38
Plagioclase	2	5	3	8
Orthoclase and perthite	7	11	7	8
Microcline	1	----	Tr.	1
Muscovite and biotite	2	Tr.	1	1
Accessory minerals	----	----	----	Tr.
Matrix	16.0	6	16	44

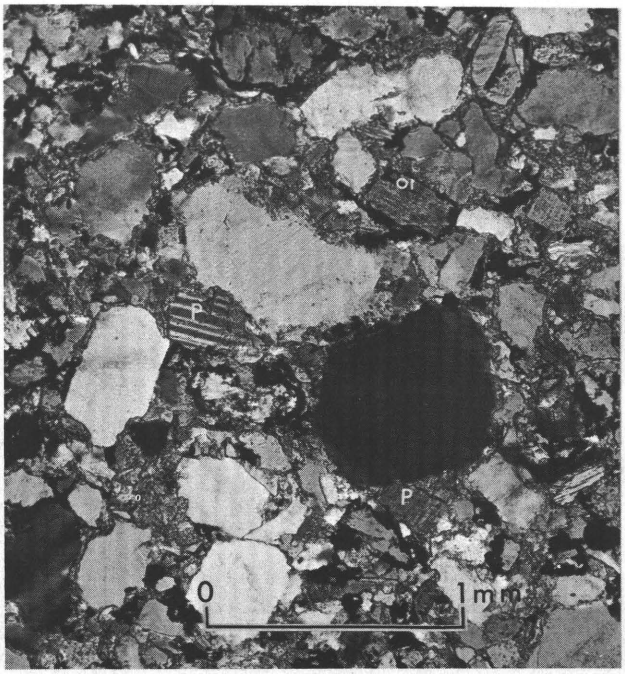
1–3. Feldspathic sandstone.
4. Calcareous sandstone.



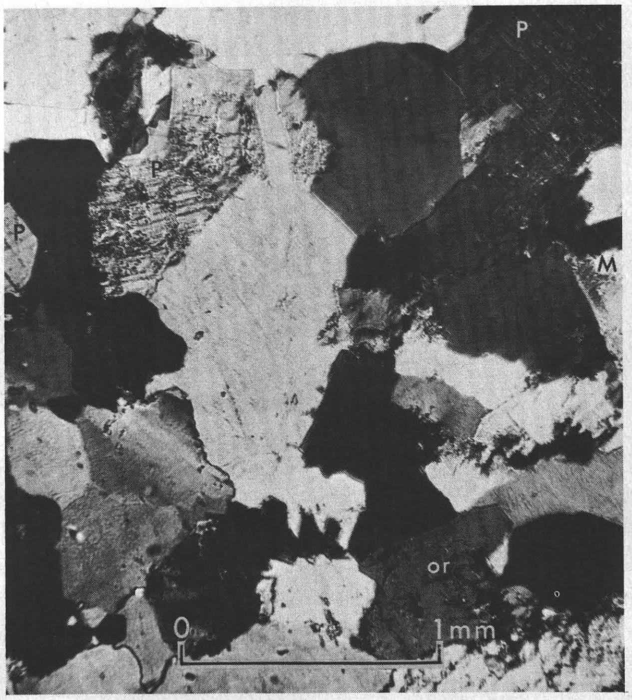
A. Grains range widely in size and angularity; some edgewise flakes of muscovite in a matrix of quartz, sericite, and chlorite.



B. Loosely packed grains cemented by finely crystalline calcite.



C. Closely packed grains ranging widely in size, in a sericite matrix.



D. A quartzitic variety composed of tightly packed angular grains with almost no distinguishable cement.

FIGURE 2.—PHOTOMICROGRAPHS OF FELDSPATHIC SANDSTONE OF THE HARMONY FORMATION.
Unlabeled grains are mostly quartz; others are plagioclase (P), orthoclase (or), mirocline (mi), and flakes of muscovite (M). Crossed polarized light.

The rock type corresponds to Pettijohn's (1957, p. 291) feldspathic sandstone, or to Gilbert's (Williams, Turner, and Gilbert, 1954, p. 291–292) feldspathic wacke.

The chemical composition of two specimens from the Hot Springs Range (1 and 2) and two from Battle Mountain (3 and 4) are shown on table 4.

TABLE 4.—*Chemical analyses of sandstone of the Harmony formation*

[Samples were analyzed by methods similar to those described in U.S. Geological Survey Bulletin 1036–C. Analysts (1, 2), H. F. Phillips, P. L. D. Elmore, K. E. White; (3, 4), K. E. White, H. F. Phillips, P. W. Scott, F. S. Borris, U.S. Geol. Survey, Washington, D.C.]

	1	2	3	4	Avg.
SiO_2	78.2	88.3	84.4	78.2	82.3
Al_2O_3	11.3	5.9	7.0	9.0	8.3
FeO	1.0	.16	1.0	2.0	1.04
Fe_2O_3	1.0	.14	.7	.8	.66
MgO	.82	.14	.72	1.0	.67
CaO	.22	.24	1.0	1.2	.66
Na_2O	1.8	1.2	2.0	1.8	1.7
K_2O	2.0	1.5	.66	2.2	1.6
TiO_2	.52	.32	.25	.36	.36
P_2O_5	.06	.16	.11	.10	.11
MnO	.02	.01	.04	.06	.03
H_2O+	2.2	.67	1.0	1.6	1.4
H_2O-	.31	.03	.05	.06	.11
CO_2	.05	.05	.82	.64	------
Sum	100	99	100	99	99
Total S as S	0.03	0.06	--------	--------	------
Spgr (lump)	1.96	2.49	--------	--------	------
Spgr (powder)	2.65	2.62	--------	--------	------

Throughout most of the section a considerable amount of shale is interbedded with the sandstone. Shale is possibly one-fourth as abundant as sandstone, but no reliable estimates of the sandstone: shale ratio have been made. Except for the lower part of the section, shale ranges from partings less than 1 inch thick between sandstone beds to units as much as 100 feet in thickness. In the northwest corner of the quadrangle, the amount of shale increases downward toward the Paradise Valley chert contact until there are about equal amounts of shale and sandstone. The base of the formation is marked by a very persistent shale member 75 to 150 feet thick.

The shale varies from light olive gray and dusky yellow to grayish red and dusky red; shades of olive gray and yellow predominate. The basal shale member, however, is characterized by an abundance of red beds, which constitute as much as one-half of the section.

Pebble conglomerate occurs throught the exposed section of the Harmony formation, but it seems to be most abundant in the lower half. The pebbles range in size from 4 mm to more than 1 cm; they seldom are as large as 2.5 cm. Sorting is poor; the matrix generally is composed of granules 2 to 4 mm in maximum dimension and interstitial sand. The pebble conglomerate beds commonly show graded bedding. The pebbles consist of subangular to subrounded quartz and microcline and minor amounts of chert and shale fragments. The shale fragments are usually confined to the lowermost part of conglomerate beds that rest on shale.

Beds of limestone are distributed rather unevenly through the formation. Limestone is most abundant about 3,000 feet above the base of the formation, in the northwest part of the quadrangle. Several beds—possibly one or more repeated by folding and faulting—are found at the south end of the Hot Springs Range south of the K and K quicksilver prospect and on the ridge west of the Last Chance quicksilver prospect. Three limestone beds crop out north of Mills Canyon in the southern part of sec. 4 and the NW¼ sec. 9, T. 38 N., R. 40 E. The beds of limestone range from about 50 feet to 150 feet in thickness. Some seem to be lenticular without much lateral continuity; several, however, have been followed for many hundreds and even thousands of feet. The limestone ranges from medium light gray to light yellowish brown, and from fine to coarse grained. The light-gray fine-grained limestone commonly has some light-colored chert (light brown, dusky yellow green to gray) associated with it. Calcareous shale is abundant in the vicinity of Dutch Flat but is generally lacking elsewhere.

The limestone has been recrystallized to varying degrees so that the original calcite grains are no longer visible. The limestone contains varying amounts of the same detrital grains that are constituents of the sandstone: quartz, smaller amounts of feldspar, a few flakes of mica, and some quartzite. Etching of the grains by reaction with the calcareous matrix has considerably modified their original shapes. Commonly some of the quartz grains are broken and veined with calcite. "Heavy" minerals obtained by separation of insoluble residues of the limestone are shown in table 5.

The Harmony formation on the west side of the Osgood Mountains is in a relatively thin sheet between two thrust faults. Strong deformation by shearing has resulted in a unit characterized by highly sheared dark greenish-gray shale containing boudins, boulder-like knobs and blocks as much as 10 feet on a side, of feldspathic sandstone, grit, pebble conglomerate, and limestone. Some beds of feldspathic sandstone are continuous for 100 feet or more, and one sheared lenticular mass of coarsely crystalline gray clastic limestone contains fragments of fossils.

TABLE 5.—*Heavy minerals in limestones of the Harmony formation*

[Separations made in bromoform capable of floating pure calcite. Based on examination of −100 to +150 mesh size fractions.

1. Sandy limestone
2. Limestone
3. Shaly limestone
4, 5. Limestone

	Weight percent
F (flood)	>30
A (abundant)	10–30
C (common)	5–10
R (rare)	1–5
Tr (trace)	<1 (usually one or two grains)]

Sample No.	Pink zircon	Color-less zircon	Blue tour-maline	Green tour-maline	Mon-azite	Xeno-time	Horn-blende	Musco-vite	Biotite	Gar-net	Weight percent non-opaque miner-als in total heavy miner-als	Leuco-xene	Ilme-nite	Ru-tile	Mag-netite	Hema-tite	Py-rite	Weight percent opaque miner-als in total heavy min-erals
1	-----	R	C	C	-----	-----	-----	F	C	-----	30	F	C	-----	C	F	C	70
2	Tr	R	Tr	C	-----	R	-----	F	C	R	10	F	A	C	C	F	-----	90
3	A	R	R	C	-----	Tr	Tr	F	C	-----	10	F	C	-----	C	F	F	90
4	C	A	R	A	R	C	-----	F	------	C	20	C	F	C	C	F	-----	80
5	Tr	C	-----	C	-----	-----	-----	F	F	-----	70	R	C	-----	C	C	F	30

STRATIGRAPHY AND THICKNESS

The Harmony formation rests on the Paradise Valley chert (pl. 1) in the northwestern part of the quadrangle on the west side of the Hot Springs Range. The contact is not visible, but evidence based on the distribution of float and scattered outcrops strongly suggests a conformable relation or only slight and perhaps local angular discordance between the two formations. Locally the contact may be a low-angle thrust fault of minor displacement due to tight folding and difference in competency between the rocks of the two formations. The stratigraphic relation between the Harmony formation and younger Paleozoic formations cannot be demonstrated in the Osgood Mountains or the Hot Springs Range, because the younger formations are in fault contact with the Harmony formation.

The Battle formation of middle Pennsylvanian age rests unconformably on the rocks of the Harmony formation on Battle Mountain in the Antler Peak quadrangle (Roberts, 1951), and in the Edna Mountains (Ferguson, Roberts, and Muller, 1952); the Pennsylvanian and Permian Antler Peak limestone overlies the Harmony unconformably in the Sonoma Range (Ferguson, Muller, and Roberts, 1951).

The stratigraphy of the Harmony formation could not be studied in any detail because the formation is structurally so complex in most places. In the Hot Springs Range the base of the section is marked by a very persistent shale member, 75 to 150 feet thick, above which lies an uncertain thickness of sandstone with interbedded shale and small amounts of grit and pebble conglomerate. Limestone seems to be more common in the upper part of the section on the east side of the Hot Springs Range in this quadrangle, but elsewhere it is also found in beds that may be lower in the section.

The total thickness of the Harmony formation is unknown because the top of the formation is not exposed and tight folding makes precise stratigraphic correlation impossible. The exposed thickness in the Hot Springs Range estimated from structure sections is more than 4,000 feet. Ferguson, Muller, and Roberts (1951) estimated a thickness of more than 5,000 feet at the type locality in the Winnemucca quadrangle, and more than 2,000 feet in the Golconda quadrangle.

AGE AND CORRELATION

Previous studies of the Harmony formation by Ferguson, Muller, and Roberts (1951), Ferguson, Roberts, and Muller (1952), and Roberts (1951) were in areas where, because of faulting, the relations to older formations were unknown, and the rocks were unfossiliferous; the oldest rocks known to overlie the formation are of Early Pennsylvanian age. These relations and the observation that "beds of similar lithology [occur] in the Permian (?) Inskip formation and Permian Edna Mountain formation * * *" led Ferguson, Muller, and Roberts (1951) to "* * * suggest that an age not older than Mississippian is probable for the Harmony formation."

In the Hot Springs Range the Harmony formation overlies the Paradise Valley chert and, therefore, is younger than that formation. Fossils have been found in discontinuous limestone beds or lenses in the sheared shale and sandstone of the Harmony formation in Goughs Canyon on the west side of the Osgood Mountains. A. R. Palmer reports (1954, written communication) Collection No. USGS 1504–CO as follows:

Two identifiable trilobites, *Theodenisia?* sp., and *Leiocoryphe?* sp., were found. Similar trilobites have been described from boulders in the Levis conglomerates in Quebec and have most recently been considered to be upper Franconian (medial Upper Cambrian) to early Trempealeauan (late Upper Cambrian) in age.

In 1960, Palmer (written communication) reported the following forms:

Collection No. USGS 1807–CO. Lens of sheared limestone in Goughs Canyon on the west side of the Osgood Mountains, North edge of ridge, NE¼SW¼SE¼ sec. 22, T. 38 N., R. 41 E.

Pseudagnostus sp.	*Protopeltura* sp.
Agnostus sp.	*Conokephalina?* sp.
Homagnostus sp.	

Of these he says,

The presence of an olenid trilobite resembling *Protopeltura*, and *Conokephalina?* sp indicates that this collection is probably Middle Late Cambrian (Franconian) in age. Associated agnostids could equally well indicate an early Late Cambrian age according to our present inadequate knowledge.

The Harmony formation cannot be correlated lithologically with Upper Cambrian formations elsewhere in Nevada, but the fossil collections indicate that it is approximately equivalent with the Windfall formation (Nolan, Merriam, and Williams, 1956, p. 19–23) in the vicinity of Eureka, eastern Nevada.

No fossiliferous beds have been found in the Harmony formation in the part of the Hot Springs Range that is within this quadrangle; however in the Hot Springs Peak quadrangle, about 2 miles north of Stewart Gap, fossils have been collected from limestone of the Harmony formation at two localities, and the following forms have been identified by A. R. Palmer (1953, written communication):

Collection No. USGS 1376–CO. Hot Springs Peak quadrangle, about 2 miles north of Stewart Gap road. About 0.75 mile S. 21°30′ E. from lat 41°20′ N., long 117°25′ W., at north edge NE¼ sec. 25, T. 40 N., R. 40 E.

Briscoia?	*Idiomesus* cf. *I.* levisensis (Rasetti)
Eurekia sp.	*Illaenurus* sp.
Geragnostus? sp.	*Prosaukia?* sp.
Homagnostus sp.	

Collection No. USGS 1375–CO. Hot Springs Peak quadrangle, on ridge in NW corner SW¼ sec. 30, T. 40 N., R. 41 E. (unsurveyed).

Billingsella sp.	*Prosaukia?* sp.
Dartonaspis sp.	*Ptychaspis* sp.
Eurekia? sp.	

Palmer (1953, written communication) states that the collections indicate a late Late Cambrian age approximately equivalent to the Windfall formation at Eureka, Nevada (Nolan, Merriam, and Williams, 1956, p. 20). The fauna from collection 1375–CO is slightly older than the fauna from 1376–CO.

ORIGIN

The mineralogy of the sedimentary rocks indicates that the Harmony formation was derived from a terrane of acidic granitic-textured rocks with some quartzite or sandstone, or both. The 958 million year age for zircon in the formation in the Antler Peak quadrangle indicates that the source rocks were at least in part of Precambrian age. A rather minor amount of volcanic material and chert were also contributed from the source area.

The material composing the bulk of the formation either has been transported no great distance or was carried in from far away by turbidity currents, as indicated by the predominance of angular grains, unabraded quartz overgrowths, and euhedral or angular heavy minerals. The basal shale unit is evidence of a source area of low relief from which the streams carried mud and silt. The overlying coarser clastic units with abundant fresh feldspar are evidence of rapid uplift of the source area with consequent vigorous erosion and rapid transportation and deposition.

The environment of deposition is postulated to have been a rapidly subsiding basin in which the sandstone and shale were deposited below wave base. The graded beds probably were deposited at times of high seasonal runoff, possibly aided by turbidity currents generated when sediment-charged rivers entered the sea. The occasional beds of clastic limestone and red shale are indicative of shallow-water conditions that prevailed where sedimentation had nearly caught up with subsidence. The rare chert beds were probably formed as chemical precipitates during periods when clastic sedimentation was at a minimum, or by the metasomatic replacement of limestone, or perhaps by both processes,

ROCKS OF ORDOVICIAN AGE

Two formations of Ordovician age have been mapped in the Osgood Mountains quadrangle (pl. 1). These are the Comus and Valmy formations, which were originally defined in the Golconda and Antler Peak quadrangles, respectively (Ferguson, Roberts, and Muller, 1952; Roberts, 1951). The two formations are not found in mutual contact in the Osgood Mountains quadrangle, so the relations between them are not clearly known; however, both formations contain Lower and Middle Ordovician strata. Regional studies suggest that the Comus formation was deposited virtually where it is found, and the Valmy formation, which was deposited somewhere to the west of its present outcrop areas, has been brought into the area on a thrust fault.

COMUS FORMATION

DISTRIBUTION

The Comus formation is exposed on the east side of the Osgood Mountains in a discontinuous belt about 9½ miles long, from the mouth of Hogshead Canyon to about 2 miles northeast of the Penson Ranch. The formation is separated from the type locality, as defined by Ferguson, Roberts, and Muller (1952), in the Edna Mountains by nearly 12 miles of valley fill and some volcanic flows. A more complete section is represented in the Osgood Mountains quadrangle, and the age of the rocks is somewhat better established by fossils. We propose, therefore, that the formation be redefined in terms of its lithology on the east side of the Osgood Mountains.

Metamorphosed sedimentary rocks which we believe belong to the Comus formation occupy a tonguelike thrust plate overlying the Preble formation along the crest and eastern side of the range at the northern end of the Osgood Mountains. Two small klippen lie several hundred and a few thousand feet, respectively, west of the main sheet, in sec. 19, T. 39 N., R. 42 E. Faults also bound the sequence on the northeast and north where it is in contact with the Etchart limestone and rocks of the Farrel Canyon formation.

LITHOLOGY

The Comus formation along the east side of the Osgood Mountains is predominantly an alternating sequence of dolomite, limestone, and shale, with subordinate amounts of chert, siltstone, and tuffaceous(?) material. Sandstone and quartzite are conspicuously absent.

The dolomite varies from grayish orange to moderate yellowish brown and from light gray to grayish black. Its composition varies from nearly pure dolomite to sandy and shaly dolomite, and calcareous dolomite grading into dolomitic limestone. The gray and dark-gray dolomite is commonly thick bedded to massive cut by a network of quartz veinlets and contains many lenses and nodules of chert. The upper part of the section contains a prominent intraformational conglomerate composed of platy fragments of medium dark-gray to grayish-black surgary-textured dolomite. The grayish-orange dolomite occurs as thick beds within a section of brown-weathering light-gray to grayish-brown platy sandy dolomite.

Most of the limestone varies in shades of gray and some beds have a definitely bluish cast. It ranges from fairly pure limestone to shaly limestone and grades into dolomitic limestone and calcareous dolomite. Bedding is clearly defined at most places, and it ranges from thick bedded to thin bedded. Some limestone units contain interbedded chert and shale. Some limestone conglomerate has also been observed.

The shale that is interbedded with the carbonate units is mostly gray, commonly with a tinge of green or greenish yellow. Most of it is well stratified and has a secondary cleavage that intersects the stratification from 20° to 90°. At some places where the beds have been tightly folded, closely spaced intersecting cleavage planes have cut the shale into pencillike fragments. The section also contains a moderate amount of bluish-gray siliceous shale, and some dark yellowish-green siltstone.

Tuffaceous-appearing shale or siltstone is characteristic of the Comus formation in this area. The rock is grayish orange to dark yellowish brown, rather soft, and highly porous, as much as 30 percent of its volume occupied by small (0.05 mm) open cavities of irregular shape. It contains angular to subangual fragments of quartz and feldspar averaging 0.03 mm in greatest dimension, and wisps of sericite in an exceedingly fine grained groundmass composed of feebly birefringent cryptocrystalline material and scattered minute granules of clay and iron oxide.

Nodules, lenses, and thin beds of dark chert are associated with the dolomite. Abundant dark chert has apparently replaced limestone and dolomite in an area of extensive faulting on the low hill south of the mouth of Hogshead Canyon. Here also, scattered bodies of barite replace the carbonate rocks and, to a lesser extent, thin chert lenses in the dolomite.

The Comus formation on the thrust plate at the north end of the Osgood Mountains includes phyllite, calc-silicate hornfels, recrystallized limestone, and subordinate amounts of dark recrystallized chert. Individual units have little resemblance to the unmetamorphosed sedimentary rocks of the Comus formation on the east side of the range, but their gross aspect indicates that originally the sequence was fine-grained calcareous or dolomitic siltstone, shale, and carbonate rocks, probably including limestone and dolomitic limestone or dolomite, and some chert. These rocks also have been intricately folded, probably because of thrust faulting.

STRATIGRAPHY AND THICKNESS

The only contacts of the Comus formation with other Paleozoic formations are fault contacts (pl. 1). Therefore, its stratigraphic position is established solely on its age as indicated by fossils. The Paradise Valley chert and the Harmony formations, which should be present between the Comus and Preble formations, have apparently been cut out by the high-angle fault that separates the Comus and Preble on the east side of the Osgood Mountains. This fault appears to be an extension of the mineralized Basin-Range fault from which gold has been produced at the Getchell mine. At the type locality in the Golconda quadrangle the contact of the Comus and the Preble is a high-angle reverse fault.

Stratigraphic units in the Comus formation cannot be correlated from one area to the next, because the outcrop areas are discontinuous and the carbonate units are lenticular, and because folding and, perhaps to a lesser extent, faulting cause apparent changes in the thickness of strata. Metamorphism has also made correlation more difficult.

A generalized section of the Comus formation south of Granite Creek follows:

	(ft)
Limestone, light-gray to grayish-brown, thin- to thick-bedded; a few thin shale beds and a few thin lenses and beds of brown chert	700+
Dolomite, medium dark-gray to grayish-black, massive. Some grayish-brown to black chert beds. Conspicuous "flat-pebble" intraformational conglomerate in upper part of the unit	1,000+-1,200+
Shale, green to gray; some fine-grained tuff?; some siltstone; minor siliceous shale	350-1,050
Dolomite, medium dark-gray to grayish-black, massive, extensively fractured; upper part interfingers with overlying shale; lenses and nodules of dark chert	450+
Platy dolomite, brown-weathering light-gray to brownish-gray, sandy; some beds of buff medium- to thick-bedded dolomite	200+
Limestone, light- to dark-gray, thin- to medium-bedded; some interbedded shaly limestone and shale	450+
Shale and phyllite, light-gray to dark greenish-gray	100-600
Fault contact with Preble formation.	

The exposed thickness of the Comus is, therefore, on the order of 3,200 to 4,600 feet. Ferguson, Roberts, and Muller (1952) estimated about 3,000 feet of beds at the type locality.

AGE

The age of the Comus formation has been established by collections of graptolites and a single trilobite mold. Two graptolite collections were made from shales that crop out on the small hill south of the mouth of Hogshead Canyon. Mr. Josiah Bridge (1952, written communication) reported on the graptolites as follows:

Collection (field No.) H-7-52. Osgood Mountains quadrangle, brick red tuffaceous shale from southern part sec. 12, T. 37 N., R. 41 E. (0.2 mile N. 25° E. from main barite quarry).

One specimen contains a single, well-preserved fragment of *Didymograptus similis* (Hall). The second specimen contains at least two distinct forms, one of which may be identical with the above. They are, however, so badly distorted by metamorphism that not even a tentative identification can be made.

D. similis is a Deepkill, or Lower Ordovician form, and if this identification is correct this fauna falls somewhere between the two faunas listed by Ruedemann from Summit, Nevada (1947, p. 107).

R. J. Ross, Jr., and W. B. Berry (written communication) examined another collection and identified specimens as follows (written communication, January 1960):

Collection No. USGS 1072-CO (field No. H-83-51). Barite quarry, E. edge NW¼SE¼ sec. 12, T. 37 N., R. 41 E., Osgood Mountains quadrangle.

Climacograptus bicornis (J. Hall)
sp.
Diplograptus? sp.
Orthograptus aff. *O. calcaratus* Lapworth

Age: Probably the zone of *Climacograptus bicornis.*

Ross and Berry (written communication, January 1960) examined a small graptolite collection from a shale unit which overlies thin-bedded and massive chert in the thrust plate at the north end of the Osgood Mountains.

Collection No. USGS 1373-CO (field No. F53-W-78). Just below crest of ridge in SE part SE¼SE¼ sec. 30, T. 39 N., R. 42 E.

Dicellograptus cf. *D. divaricatus* var. *bicurvatus* Ruedemann
cf. *D. sextans* (J. Hall)
Unidentifiable scadent form

Age: Probably zone of *Climacograptus bicornis.*

A single mold of a trilobite was found in beds that are at about the same stratigraphic position or slightly lower than those from which H-7-52 was collected. According to R. J. Ross, Jr. (1955, written communication), who reported on this fossil, the specimen is not good enough for certain identification, but it resembles *Acerocare* and *Cyclognathus.* He said that both "* * * are typically Upper Cambrian genera in the Baltic region but *Acerocare* has been reported in the lowest Lower Ordovician * * *" and that the specimen came from beds that may be "* * * a little lower stratigraphically than your H-83-51 [Coll. No. USGS 1072 (CO)] or H-7-52."

The fossil evidence indicates that the Comus formation probably ranges in age from Early to Middle Ordovician, and may be as old as Late Cambrian.

VALMY FORMATION

The type locality of the Valmy formation is about 20 miles southeast of the Osgood Mountains quadrangle, on North Peak in the Antler Peak quadrangle. According to R. J. Roberts (oral communication) the formation consists of interbedded chert, quartzite, argillite, slate, and greenstone. He subdivided the Valmy into a lower and an upper unit totaling more than 8,000 feet. Only a few small areas of Valmy formation are known in the Osgood Mountains quadrangle, and we have interpreted these as remnants of a formerly much more extensive thrust sheet.

DISTRIBUTION

Rocks assigned to the Valmy formation are exposed at two places along the east front of the Hot Springs Range, and in the low hills east of the Getchell mine (pl. 1). Only the westernmost end of the second locality is within the quadrangle.

The Valmy formation on the east side of the Hot Springs Range is exposed on the hill west of Stone Corral and on the low hills just east of Box Spring, about 6 miles north of Stone Corral. At both places the Valmy formation is in fault contact with the Harmony formation and Tertiary volcanic rocks and is covered on the lower parts of the slopes by alluvium. East of the Getchell mine, limestone of Pennsylvanian age is thrust over the Valmy formation, and the Valmy is probably thrust over rocks that belong to the Comus formation, although the trace of the thrust and the rocks on either side of it are covered by alluvium.

LITHOLOGY

The Valmy formation in the Hot Springs Range consists of chert, siliceous shale, quartzite, and some interbedded altered volcanic rocks. The Valmy east of the Getchell mine consists of greenstone and some interbedded limestone in the lower part of the section, overlain by siliceous shale and chert.

The quartzite is light colored, ranging from almost white to, rarely, medium bluish gray. It is dense, medium grained, and exceptionally pure, containing 95+ percent quartz and 3 to 5 percent quartzite grains. The quartz and quartzite grains range from about 0.1 mm to 1 mm and average about 0.5 mm; they are rounded to well rounded but commonly show incipient development of interlocking borders. The interstitial material is recrystallized silica cement with a very small mount of fine mica shreds. Some of the quartz and quartzite grains contain crystals of apatite, zircon, and rarely, mica. No chert fragments or rock fragments other than the quartzite have been observed.

Chert in the formation ranges from light gray to black and is thin bedded to massive. Some chert specimens show fine mica particles oriented in more or less parallel bands and small (0.2 mm) angular quartz grains with a cryptocrystalline silica cement, suggesting that the chert originated by silicification of shale. Other specimens of chert contain dark fine-grained fragments with shardlike outlines and a few irregular clots of chlorite; these probably are silicified tuffs. Other specimens show no relict structures that might be indicative of origin.

The siliceous shale is light gray, light brown, and light green. It commonly occurs as partings between chert beds, but it also forms beds as much as 2 feet thick interbedded with limestone and altered volcanic rocks. The siliceous shale is composed of very fine angular quartz fragments, fine-grained mica particles, very finely divided moderately birefringent material which may be mica or clay particles, and a cryptocrystalline silica cement. The grain size of most of the quartz and mica fragments is in the lower limit of the silt size range (smaller than 0.01 mm) and very few rocks are shale on the basis of grain size alone, but because of the shaly partings produced by the mica, the rocks are referred to as shale.

Dark greenish-gray fragmental altered volcanic rocks are exposed east of the Getchell mine. Most of the rocks are composed of small fragments in a still finer grained groundmass, but in some places the rocks are composed of pieces of volcanic rock an inch or so across in a limestone matrix. Many of the fragments retain an original amygdaloidal or porphyritic structure. Much of the original texture of the rocks, however, has been destroyed by alteration. The rocks are now composed of masses of pale-green to bluish-green actinolitic hornblende; and the plagioclase, though retaining the former shape of phenocrysts, is recrystallized to a fine-grained mosaic of calcic oligoclase or possibly andesine. Some clear, more coarsely crystalline oligoclase occurs with actinolitic hornblende as cavity fillings. Some pale biotite occurs in scattered interstitial masses and as amygdaloidal fillings and partial replacements of plagioclase phenocrysts. In addition there is some clinozoisite and abundant secondary sphene and magnetite.

Some altered volcanic rock is associated with the chert and siliceous shale in the outcrops west of Stone Corral. These are dense, green and gray, sheared rocks that no longer have much resemblance to igneous rocks. Some of these are composed almost wholly of contorted folia of chlorite and small amounts of magnetite and calcite veins. One specimen has a relict intersertal texture shown by kaolinite and illite derived from feldspar laths, montmorillonite derived from interstitial glass, and chlorite, calcite, and magnetite formed from ferromagnesian minerals.

STRATIGRAPHY AND THICKNESS

The stratigraphy and thickness of the Valmy formation in the Osgood Mountains quadrangle are poorly known because of the limited exposures, complex folding of the beds, and complete ignorance as to the top or bottom of the section. The two localities east of the Hot Springs Range cannot be correlated with each other, and they cannot be matched with the strata exposed east of the Getchell mine.

An east-to-west section on the hill west of Stone Corral consists of: 200 to 300 feet of sheared and mylonitized shale with a phyllitic appearance, known only from float; 100 to 150 feet of thick-bedded to massive quartzite (one unit of which, 25 feet thick, forms a persistent outcrop over a considerable length of the hill); 250 to 350 feet of poorly exposed interbedded dark chert, siliceous shale, altered volcanic rocks, and limestone; 400 to 500 feet of poorly exposed light-gray to black chert with siliceous shale partings and some thin beds of siliceous mylonitized shaly sandstone known mainly from float; about 50 feet of massive quartzite; and, finally, about 150 feet of chert and siliceous shale float, which is covered to the west.

The total exposed section on the hill west of Stone Corral is 1,300 to 1,650 feet thick. The section at the northern locality, which is about 700 feet thick, cannot be matched with that at the southern locality. This implies a minimum thickness of nearly 2,000 feet for the formation as exposed in the quadrangle.

The strata assigned to the Valmy east of the Getchell mine consist of greenstone, a little chert, and limestone in the western part of the exposed area, and interbedded chert and siliceous shale to the east. Top and bottom directions are not known, but the beds dip steeply in a general easterly direction, so that the greenstone section is apparently below the chert and siliceous shale. Possibly 3,000 to 4,000 feet of greenstone and about 2,000 feet of chert and siliceous shale are exposed. The chert and siliceous shale closely resemble the Valmy on the east side of the Hot Springs Range but lack the beds of pure quartzite characteristic of the Valmy exposed there.

In the Antler Peak quadrangle (Roberts, oral communication) the Valmy formation is 8,000 to 9,000 feet thick and can be divided into a lower and an upper unit; the lower unit measures more than 5,200 feet in thickness.

AGE AND CORRELATION

Graptolites collected from the Valmy formation in the Antler Peak quadrangle and in the northern Shoshone Range indicate that the age of the formation is of Early, Middle, and Late Ordovician age (Roberts and others, 1958, p. 2833). Some poor impressions of graptolites were found in the siliceous shales east of the Getchell mine, but these were useless for age determination (R. J. Ross, 1955, written communication); however, clastic limestone interbedded with the greenstones in the same general area contains trilobites that were identified as very early Early Ordovician in age. The locality from which the trilobites were collected is east of the eastern boundary of the quadrangle, approximately 0.65 mile N. 46° E. from the mill stack at the Getchell mine on the south side of a low east-trending ridge (approximately SE¼ sec. 28, T. 39 N., R. 42 E.). R. J. Ross reported the following forms (1955, written communication):

Collection USGS D–151–CO.

Symphysurina cf. *S. brevispicata* Hintze
cf. *S. cleora* (Walcott)
sp.
Hystricurus aff. *H. cordai*
sp.
Leiosteguim?
Remopleuridiella? sp. (a single free cheek)

Ross regarded the forms from the first collection as indicating equivalence with the lettered "B" zone of the Garden City formation and Pogonip group (Ross, 1951; Hintze, 1952). According to him, none of the graptolites from the Valmy are as old as the "B" zone; but, as he points out, the lettered zones are based on trilobites in the eastern carbonate facies and almost no information is available about the corresponding graptolite zones. Beds from which these collections came may be equivalent to the Goodwin limestone of the Pogonip group in the Eureka, Nev., area (Nolan, Merriam, and Williams, 1956, p. 26–27), which is of Early Ordovician age and contains faunas showing relationships to faunas in the Garden City formation.

The rocks on the east side of the Hot Springs Range are correlated with the Valmy on the basis of lithologic similarity, for no fossils have been found. The correlation is regarded as fairly sound, however, because the association of beds of highly pure quartzite with chert and siliceous shale is characteristic of the Valmy at its type locality.

The lithology of the Valmy formation in the Osgood Mountains quadrangle suggests that it be correlated with the lower member of the Valmy formation mapped by Roberts in the Antler Peak quadrangle. There the lower member is pure, generally light-colored quartzite and includes significant amounts of greenstone in addition to chert and siliceous shale; whereas the upper member consists principally of dark thin-bedded chert interbedded with dark shale, and only a little greenstone.

The lower part of the Valmy is correlative in part with beds to the east in Eureka County that were assigned by Merriam and Anderson (1942, p. 1694) to the lower part of the Vinini formation in the Roberts Mountains; the upper part of the Valmy is probably equivalent in age to the upper part of the Vinini as originally defined by Merriam and Anderson.

EQUIVALENCE OF THE COMUS AND VALMY FORMATIONS, AND POSSIBLE FACIES RELATIONSHIPS

The faunas contained in the Comus and Valmy formations are evidence that both formations are of Ordovician age; yet their lithologies are dissimilar and there is no doubt that they are different formations. Neither one, however, is like the dominantly carbonate lithology of contemporaneous rocks in eastern Nevada and western Utah.

The Valmy formation is similar to the Vinini formation, which Merriam and Anderson (1942, p. 1699–1701) recognized as a western clastic facies equivalent in age with Ordovician formations in the eastern part of Nevada where carbonate rocks are predominant. The Comus formation, however, has no lithologic equivalents in either the Valmy and Vinini or the carbonate formations, but it contains a mixture of fine clastic sedimentary rocks, chert, minor amounts of silicic tuff, limestone, and dolomite, and seems to have affinities with both the western and eastern facies formations. We are inclined, therefore, to regard it as a transitional facies (see p. 79) that was deposited in an environment intermediate between those in which the western and eastern facies originated. The implications that a transitional facies has for the depositional and structural history of the region are discussed on pages 70 and 81 of this report, and have been published elsewhere (Roberts and others, 1958, p. 2816–2820).

ROCKS OF MISSISSIPPIAN AGE

A formation of Mississippian age is exposed on the west side of the Osgood Mountains and is here named the Goughs Canyon formation (pl. 2). The formation is on the upper plate of a thrust fault and is overlapped by a higher thrust and Tertiary volcanic rocks; its stratigraphic relations with other Paleozoic rocks are not known. The formation is composed mostly of altered volcanic rocks of medium to basic composition and coarse-grained fossiliferous limestone, with minor amounts of calcareous shale, siliceous shale, and chert. Rocks with a similar lithology in the northern part of the Hot Springs Range north of the Osgood Mountains quadrangle have been tentatively correlated with the Goughs Canyon formation by Willden, although he did not refer to them by that name.

GOUGHS CANYON FORMATION

DISTRIBUTION

The Goughs Canyon formation is named for Goughs Canyon, near the center of the quadrangle on the west side of the Osgood Mountains, where the best fossil collections have been found. South of Goughs Canyon the formation is exposed in discontinuous remnants of a thrust plate for 1½ miles to the upper part of Perforate Canyon; to the north it extends continuously for 5 miles to the north side of Farrel Canyon. The outcrop area is widest in East Fork of Eden Creek, where it is slightly over 1 mile wide, and it narrows to less than half a mile at Farrel Canyon. The formation is limited on the west by a northeast-trending normal fault near the mouth of Goughs Canyon, and by Tertiary volcanic rocks in the Dry Hills and near the mouth of East Fork of Eden Creek; north of East Fork of Eden Creek the Goughs Canyon formation is overridden by a younger thrust. The eastern and southern limit is the Goughs Canyon thrust fault.

The metavolcanic rocks and interbedded limestones form prominent, nearly continuous exposures north of East Fork of Eden Creek, where the slopes are steep and rugged. Exposures are also good on the ridge north of Goughs Canyon, but in Goughs Canyon the slopes are gentler and the soil cover is fairly thick, so that outcrops are scattered and only the siliceous shale and chert and beds of limestone form good exposures; the altered volcanic rocks generally weather to thick soil.

LITHOLOGY

The Goughs Canyon formation is composed predominantly of altered volcanic rocks and interbedded limestone: approximately 60 percent altered volcanic rocks; 30 percent limestone; and 10 percent calcareous sandstone, calcareous shale, siliceous shale, and chert.

The volcanic rocks are easily recognized despite their altered condition, for they contain many of the characteristic structures and textures found in rocks of volcanic origin. The rocks are greenish gray, dark grayish green, and dusky yellow green where freshly broken; weathered surfaces are moderate to dark yellowish brown. In exposures the rocks are broken into many small polygonal blocks by closely spaced joints or very commonly are sliced by closely spaced parallel shear planes. Pillow structures are visible in some exposures, and the rocks in many places are obviously volcanic breccias which are composed of fragments that range from less than an inch in diameter to blocks as much as a foot in greatest dimension. Weathered surfaces of the fragments commonly have a pitted or vesicular appearance due to the leaching out of amygdular fillings of calcite. The fragments are firmly bonded by a finer grained matrix that in many places is calcareous. Some material is a mixture of volcanic rock fragments in a limestone matrix.

The volcanic rocks are fine grained to microcrystalline; in fine-grained varieties gray plagioclase laths in a microcrystalline greenish-gray interstitial groundmass can be seen with a hand lens, and in some specimens small phenocrysts of plagioclase and relicts of pyroxene are visible, but a characteristic of these rocks is their generally nonporphyritic texture. Although the rocks are obviously altered, their original fabric has been preserved without any apparent distortion of flow structures and amygdules. The texture of the groundmass is typically pilotaxitic; phenocrysts of plagioclase and mafic minerals are uncommon. The plagioclase is sodic oligoclase in most specimens studied, though nearly pure albite was also identified. The feldspar is mostly hazy and contains many small inclusions of sericite, chlorite, and possibly some clinozoisite. Clear albite occurs as little pockets associated with magnetite veinlets in some rocks and as amygdule fillings in others. Chlorite and epidote are common constituents of the rock; some of the chlorite obviously is pseudomorphous after primary mafic minerals, possibly pyroxene. An unusual type has phenocrysts of fresh augite in a ground-mass of albitic plagioclase and abundant epidote. Actinolitic hornblende, presumably an alteration product of pyroxene, is present in some specimens. Minor accessory minerals are fine-grained sphene, magnetite, and apatite; secondary leucoxene after sphene is also common. Quartz, calcite, and chlorite occur as amygdule fillings. Calcite is plentiful in some of the fragmental rocks as part of the matrix.

Analyses of two specimens are given in table 6.

Calcareous sedimentary rocks of the Goughs Canyon formation, including clastic limestone and calcareous sandstone and shale, are interbedded with the metavolcanic rocks. Limestone is the more abundant, but sandy and shaly rocks are common. In Goughs Canyon the limestone units are mostly less than 100 feet but more than 10 feet thick and are lens shaped and discontinuous. North of East Fork of Eden Creek however, the limestone units are several hundred feet thick and can be followed continuously for thousands of feet; outcrop widths vary considerably, partly owing to folding but mainly because of original variations in the thickness of the limestone.

The limestone is medium to coarsely crystalline and gray to light gray where it is purest, but some of it contains considerable amounts of volcanic rock fragments which give it a greenish-gray color. In places the limestone is mottled gray and pinkish. Much of it is massive to poorly bedded; however, some is thinly bedded and some has a shaly parting. The limestone is locally fossiliferous; crinoid stems are abundant in some beds, corals and bryozoans are plentiful in some places and brachiopods in others. The fossiliferous units are highly lenticular, so that richly fossiliferous pods are surrounded by more or less barren limestone.

TABLE 6.—*Analyses and norms of altered volcanic rocks from the Goughs Canyon formation*

[Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C; P. L. D. Elmore, K. E. White, S. D. Botts, P. W. Scott, analysts, U.S. Geol. Survey]

Chemical analyses			Norms		
	H-22C-52	52-W-71		H-22C-52	52-W-71
SiO_2	43.9	60.1	Q	------	7.2
Al_2O_3	16.8	15.5	Or	1.4	3.3
Fe_2O_3	2.8	3.5	Ab	24.3	60.7
FeO	7.6	3.6	An	30.8	7.4
MgO	6.2	3.0	C	------	0.4
CaO	10.8	3.2	Ne	1.5	------
Na_2O	3.2	6.5	Wo	7.6	------
K_2O	.24	.56	En	4.7	7.4
TiO_2	2.5	.54	Fs	2.4	3.0
P_2O_5	.63	.76	Fo	7.5	------
MnO	.20	.08	Fa	4.2	------
H_2O	4.8	1.8	Mt	4.1	5.1
CO_2	.07	.57	Il	4.8	1.0
			Ap	1.5	1.8
Sum	100	100	Cc	0.2	1.3
Sp gr (lump)	2.87	2.64	H_2O	4.8	1.8
Sp gr (powder)	2.96	2.77	Sum	99.8	100.4

Some of the best fossil collections have been from "hash beds" containing a rich mixture of coral, bryozoan, and brachiopod remains, many of them broken. These remains probably represent material that was reworked by turbulent water shortly after deposition, transported a short distance, and redeposited without much sorting. Some of the limestone and sandy limestone has a high content of volcanic material including small angular fragments of altered crystalline volcanic rock, plagioclase, chlorite, and epidote, with some chert.

The shale and sandstone contain grains of quartz, flakes of colorless mica, and magnetite, and considerable amounts of chlorite and clayey material in the matrix. Calcite is a common constituent of some of these rocks but minor or lacking in others.

Chert is not plentiful in the formation, but a few small lenticular beds of dark chert are interbedded with the limestone and altered volcanic rocks. However, on the south side of Goughs Canyon, apparently in a part of the section different from that exposed elsewhere, chert associated with fossiliferous limestone and altered volcanic rock is fairly abundant. The chert is thin bedded to moderately thin bedded, and much of it is pale olive and dusky yellow green. Some is also dark gray and moderate bluish gray.

THICKNESS

The exposed thickness of the Goughs Canyon formation in the Osgood Mountains quadrangle is uncertain because of complex folding but may be more than 5,000 feet. How nearly this represents the total thickness is unknown because the formation is bounded above and below by thrust faults and is overlapped by later volcanic rocks and alluvium. For the same reasons the degree of completeness of the section where it is exposed in the northern part of the Hot Springs Range is unknown. The formation has not been recognized outside of these two areas.

AGE AND CORRELATION

The stratigraphic position of the Goughs Canyon formation is unknown because of thrust faulting, except that it is older than the Tertiary volcanic rocks which overlap it. Fossil evidence, however, establishes the age as probably Early and early Late Mississippian (Viséan).

A rather large collection of fossils has been obtained from the limestones interbedded with the altered volcanic rocks. The corals and bryozoa were studied by Helen Duncan, and the brachiopods by Mackenzie Gordon, Jr. Their findings are summarized in the following discussions.

Collection No. USGS 19807–PC (field No. H–76–51). Cherry Canyon, southern part of SW¼ sec. 11, T. 38 N., R. 41 E.

The fossils collected are a peculiar type of caninoid coral which Miss Duncan referred to *Timania*?. She said (written communication, 1952):

> *Timania* is known mainly from the Upper Carboniferous and Permian of Russia. These American corals seem to be closer to *Timania* than to any other genus so far described; however, they probably should be recognized as a distinct taxon. At present, the only definite information we have on the stratigraphic distribution of these corals in North America is their occurrence in the upper part of the Deseret and in the lower part of the Humbug formations of north-central Utah (East Tintic and Oquirrh Mountains).
>
> Certain other fossils in the Deseret fauna have Meramec affinities and suggest that the formation is assignable to the lower part of the Upper Mississippian. Furthermore, we generally find that in the west the larger caninoid corals, a category that includes these specimens, are much more characteristic of the Upper Mississippian than they are of the Lower Mississippian. For these reasons, it seems probable that this Nevada occurrence indicates that the coral-bearing rocks in Cherry Canyon are of early Late Mississippian age.

Collection No. USGS 19808–PC (field No. H–77–51) Cherry Canyon, Osgood Mountains quadrangle. Center SW¼ sec. 11, T. 38 N., R. 41 E. Near locality of H–76–51, and from same limestone unit.

Of this collection, Helen Duncan reported (written communication, 1952):

> This lot contains a recrystallized tabulate coral identified as *Michelinia*? sp. indet., and indications of indeterminate rhomboporoid bryozoans. These fossils are of Carboniferous types, but they are not particularly useful for precise correlation. From general observations on faunas in the region, such fossils are more likely to occur in the Upper Mississippian or in the Pennsylvanian than they are in the Lower Mississippian.

Collection No. USGS 14206–PC (field No. H–13–52), Goughs Canyon, NW¼NW¼ sec. 27, T. 38 N., R. 41 E. Bank north of main stream.

Corals:
- *Michelinia* sp.
- *Cyathaxonia* sp.
- *Zaphrentites* sp.
- *Triplophyllites* sp.
- *Rhopalolasma* n. sp.
- *Permia*? sp.
- *Rylstonia* sp.
- *Trochophyllum* sp.
- Zaphrentoid coral fragments, indet.

Concerning this collection Miss Duncan said (written communication, 1953):

> It was amazing to find such a varied assemblage of coral genera in such a relatively small collection (less than 30 specimens). As I recall the corals were difficult to get out of the rock without breaking them, and we collected very few examples of the larger forms. I suspect that an even larger assortment of genera is to be found at this locality. At least two of the genera have not been reported as of 1953 in the United States. *Rhopalolasma* and *Rylstonia* were described from the Lower Carboniferous of Great Britain and are known only from the upper Tournasian and Viséan of western Europe. *Permia* occurs in the Tournasian and Viséan of Europe and Asia, and possibly in the Keokuk of the Mississippi Valley. Most of the corals in the collection came from a clastic limestone and were fairly well broken up before they were incorporated in the enclosing sediments. For this reason, the material is not entirely satisfactory for investigation or comparison with described species. The coral faunule certainly has a Mississippian aspect, but at this stage of investigation, it is impossible to suggest any precise correlation with the Mississippi Valley section or even with the Rocky Mountain section. Some of the corals suggest correlation with the upper Tournasian and the Viséan of Europe and Asia.

Bryozoa:
- Fistuliporoid bryozoans, genera undet.
- Stenoporoid bryozoans, genera undet.
- *Fenestella*, 2 or more species
- *Hemitrypa* sp., possibly new
- *Polypora* sp. undet.
- *Thamniscus*?, probably 2 species
- *Ptilopora*? sp. undet.
- *Cystodictya* cf. *S. lineata* (Ulrich)
- *Rhabdomeson* sp. undet.

Miss Duncan (written communication, 1953) reported on the bryozoans as follows:

> One bed at this locality contains a diversified and fairly well preserved bryozoan fauna, which helps confirm the Mississippian age of the limestones and may eventually be significant in their correlation. The *Hemitrypa* may be a new species and is probably identical with the one in Collection USGS 14029–PC. The *Cystodictya* is close to or identical with a species that is abundant in the Keokuk and lower Meramec of the Mississippi Valley. On the other hand, the bryozoan *Rhabdomeson* has not been reported prior to 1953 in the Mississippian of this country. The genus was described on material from the Lower Carboniferous of England. The genus is rather common in our Pennsylvanian and Permian, but if it occurs at all in our type

Mississippian it has never been identified. As it is rather easy to recognize, I do not think it is very likely that it has escaped detection. Material for some of the other genera is scant.

The following brachiopods were collected from the Goughs Canyon locality (USGS 14026–PC) and identified by Mackenzie Gordon, Jr.:

Brachiopods:
- "*Productus*" (*Setigerites*?) aff. *P. setiger Hall*
- "*Productus*" (Linoproductus) sp.
- "*Productus*" (Avonia) sp.
- "*Productus*" sp.
- *Schizophoria*? sp.
- *Spirifer* aff. *S.* grimesi Hall
- *Spirifer* sp.
- *Brachythyris* aff. *S. suborbicularis* (Hall)
- *Strophopleura* n. sp.

Gordon reported (written communication, 1953):

This fauna is a rather puzzling one, but it appears to be Mississippian in its general composition. The productid related to "*P.*" *setiger* Hall is represented by poorly preserved specimens and is also reminiscent of the genus *Dictyoclostus* which has a fairly long range through the Carboniferous and Permian. The other productids likewise belong in genera that are long-ranging. The *Spirifer* related to *S. grimesi* Hall is indicative of Mississippian age, but in western America, Spirifers of this type may be found rather generally distributed through the Mississippian. The *Brachythyris* resembles a lower Mississippian type, but the genus ranges through a considerable part of the Mississippian. Perhaps the most interesting form in the collection is the single specimen I have identified as *Strophopleura*, a genus known at present only from Late Devonian and early Mississippian rocks. The species, however, appears to be a new one, and therefore there is no guarantee that it does not represent an extension of range of the genus. The other forms that I have not mentioned specifically have no additional stratigraphic implications.

I have also conferred with Helen Duncan, who has studied the corals and Bryozoa from this collection. Some of them are bizarre types that are closer to European than to American forms. We both agree that this is not the sort of fauna with which we are familiar in other western localities (for example, in west-central Utah). There are enough genera, usually restricted to the Mississippian, among the corals, bryozoans, and brachiopods to suggest that this is indeed a Missisippian fauna and probably one well below the top of the Missisippian. I think that the best age determination we can give at present is that the lens containing the fauna is pre-Chester Mississippian in age.

Subsequent work on Mississippian rocks and faunas in north-central Nevada has furnished a better frame of reference for dating collection No. USGS 14026–PC and for interpreting the age span of fossiliferous beds in the Goughs Canyon formation. Duncan and Gordon (written communication, 1961) stated:

The fauna in collection No. USGS 14026–PC contains several corals and brachiopods that appear to be identical with genera and species found in the Mississippian rocks of western Elko County, where the sequence of Early Mississippian faunas has been worked out (Gordon and Duncan, 1961). This particular assemblage from the Goughs Canyon formation is clearly related to Elko County faunules that suggest early Osage age.

Collection No. USGS 14029 (field No. 52–W–16), SW¼SE¼ SE¼ sec. 28, T. 38 N., R. 41 E.

This collection contains one bryozoan, two corals, and a piece of crinoidal limestone with indeterminate fragments of corals and Bryozoa. The species identified are:
- *Syringopora* aff. *S. surcularia* Girty
- *Lithostrotion* aff. *L. whitneyi* Meek
- *Hemitrypa* sp., possibly the same as the species from USGS locality 14026.

Miss Duncan (written communication, 1953) said of this collection:

The *Syringopora* has smaller, more widely separated, and evenly spaced corallites than the type of *S. surcularia*, which is very common in the Madison and equivalents in the West. I have seen comparable specimens from rocks that are supposed to be of post-Madison age but have not studied the material in enough detail to determine whether the variation is persistent enough to distinguish a subspecies.

The *Lithostrotion* falls within the limits of *L. whitneyi* as it has been interpreted by various authors, but the corallites are smaller than those of the type. *L. whitneyi* occurs in the lower part of the Upper Mississippian in Utah. This specimen is also related to *L. irregulare* Phillips and *L. scoticum* Hill described from the Viséan of Great Britain.

The *Hemitrypa* is probably an undescribed species, at least it does not appear to fit the description for any species described from the Mississippian of North America. This bryozoan is one of the few fossils that have so far been found at more than one locality in the Goughs Canyon formation. It is abundant in one bed at USGS locality 14206.

A Mississippian age is certainly indicated by the fossils collected. *Hemitrypa* is not known to persist beyond the Lower Carboniferous in the United States. The *Syringopora* is related to a common Madison species, but it is not known whether this particular variant has a restricted range. Certain lithostrotionoid corals (*Lithostrotionella* especially) occur in the Lower Mississippian, but the phaceloid lithostrotionoids appeared somewhat later. In Utah and Idaho, phaceloid forms such as the one in this collection characteristically are found in the Upper Mississippian, and we have tentatively considered that their appearance coincided approximately with the beginning of Late Mississippian (Meramec) time. However, corals of this type do occur in the uppermost part of the Mission Canyon and Redwall formations, and it may be that they first appeared in late Early Mississippian faunas in areas where environmental conditions were favorable.

Helen Duncan (written communication, 1961) has contributed the following discussion on the age of the Goughs Canyon formation as a unit:

When collections from the Goughs Canyon formation were first studied in 1952 and 1953, it was concluded that the faunal evidence pointed to a Late Mississippian age for the included fossiliferous rocks. The occurrence of *Timania*? and phaceloid lithostrotionoid corals in some lots favors this assignment, though Gordon and I were very puzzled about the significance of the fauna collected at USGS locality 14206–PC. Re-

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cent work on the faunas from western Elko County (Gordon and Duncan, 1961) has clarified our ideas considerably. Available evidence suggests that the lenticular limestones of the Goughs Canyon formation carry faunas that range in age from late Early Mississippian (Osage equivalent) to early Late Mississippian (Meramec equivalent). The upper beds of the fine clastic sequence in Elko County also contain phaceloid lithostrotionoids and other corals suggestive of Late Mississippian age although *Timania*? has not as yet been identified from the area. The corals obtained from the Inskip and Banner formations, though deformed and recrystallized, seem to belong to the same general faunal sequence.

The Goughs Canyon formation is probably correlative with the Mississippian(?) Inskip formation of the East Range, Winnemucca quadrangle, Nevada, which was regarded by Ferguson, Muller, and Roberts (1951) as Permian(?). The Inskip contains volcanic rocks and limestone that somewhat resemble rocks of the Goughs Canyon formation, and is now known to contain fossils of probable Late Mississippian age (Roberts, and others, 1958 p. 2847). Limestone of Late Mississippian age is common in the eastern part of the Great Basin (Nolan, 1935, p. 27–29; Gilluly, 1932, p. 25–26; Weller, 1948), but no precise correlations with the Goughs Canyon formation are possible. No reliable correlation with Oregon or California Mississippian rocks can be made. The Coffee Creek formation of central Oregon was determined by Merriam and Berthiaume (1943, p. 149–151) to be "* * * Lower Carboniferous, roughly Viséan, in terms of the British succession", and, according to Duncan and Gordon, is probably a little younger than the Goughs Canyon formation. The faunal assemblages are different, however, and the Coffee Creek formation, which consists of limestone, argillaceous to sandy limestone, and calcareous sandstone but does not contain volcanic rocks. The upper part of the Baird formation in the Redding-Weaverville districts, California, is related faunally to the Coffee Creek formation (Merriam and Berthiaume, 1943, p. 163) and has volcanic material associated with the sedimentary rocks (Hinds, 1933, p. 92–93).

CONDITIONS OF DEPOSITION

The clastic limestones with their localized concentrations of fossils comprising several forms are evidence of a shallow marine environment where turbulent waters reworked unconsolidated calcareous sediment shortly after its deposition. Beds of dirty limestone and calcareous, argillaceous, and arenaceous rocks probably can also be attributed to a stirred-up limy sea floor though locally, perhaps in deeper quieter basins, some fine-grained siliceous sediments accumulated. Some of the volcanic rocks were extruded in the sea, where they became brecciated on contact with the water and churned up and incorporated some of the limy deposits on the sea floor. Some of them may have formed shallow sills in the ocean sediments. Probably in places they built up piles that rose above the surface or were temporarily elevated above sea level by earth movements, and further extrusion took place subaerially, for there are thick sections with no indications that they were deposited on a sea floor. Explosive volcanism resulting in the deposition of tuffs seems to have been relatively minor.

ROCKS OF PENNSYLVANIAN AND EARLY PERMIAN AGE

Rocks ranging in age from Middle Pennsylvanian to Late Pennsylvanian or Early Permian rest unconformably on the older Paleozoic rocks in the Osgood Mountains (pl. 1); none are known in the southern end of the Hot Springs Range. Similar rocks elsewhere in north-central Nevada have been called the overlap assemblage (Roberts and others, 1958, p. 2821, 2838–2846). They represent sediments deposited in marine basins or on their margins following a late Paleozoic orogenic episode (Roberts, 1951; Roberts and Lehner, 1955, p. 1661; Ferguson, Muller, and Roberts, 1951; Ferguson, Roberts, and Muller, 1952) that formed a highland area through central Nevada (Dott, 1955, p. 2288). Rocks of the overlap assemblage in this part of Nevada have become known as the Antler sequence (Roberts and others, 1958, p. 2839, 2841), which includes the Battle and Highway formations of Middle Pennsylvanian age, the Antler Peak limestone of Late Pennsylvanian and Permian age, and the Edna Mountain formation of Permian age.

The dominantly terrestrial conglomerate of the Middle Pennsylvanian Battle formation rests unconformably on older Paleozoic rocks. It is overlain by limestone, and westward it wedges out and interfingers with limestone that also rests on older Paleozoic rocks. Ferguson, Roberts, and Muller (1952) named the limestone equivalent of the Battle formation in the Edna Mountains the Highway limestone; in the Antler Peak quadrangle, Roberts (1951) called the limestone above the conglomerate in the Battle, the Antler Peak limestone. Representatives of both the Highway and Antler Peak formations are present in the Osgood Mountains, but they cannot be separated on the basis of lithology; so they have been mapped as one unit, here designated the Etchart limestone. A clastic facies of the Etchart limestone, here named the Adam Peak formation, has been thrust over the Etchart limestone. The Edna Mountain formation does not occur in the Osgood Mountains quadrangle.

BATTLE FORMATION

The Battle formation, which crops out in the central part of the Osgood Mountains, was named and defined by Roberts (1951) in the Antler Peak quadrangle, where it takes its name from Battle Mountain. The formation has subsequently been recognized as far west as the Edna Mountains in the northwest part of the Golconda quadrangle and as far east as the Shoshone Range, Mount Lewis quadrangle, but it is best developed at its type locality. Hague and Emmons (1877, p. 688) briefly mentioned this unit in the 40th Parallel report; and Lawson (1913, p. 328–329), describing it in some detail, regarded it as an ancient fan deposit for which he proposed the now common term, "fanglomerate" (Lawson, 1913, p. 329–330).

DISTRIBUTION

In the Osgood Mountains quadrangle the Battle formation is confined to the central and south-central parts of the Osgood Mountains (pl. 1). In most places the formation is overlain by younger strata, but midway down the eastern side of the range it occurs as gently dipping to nearly flat erosion remnants from which the higher beds have been removed. The crest of the ridge for 3½ miles south of Hogshead Canyon is composed of Battle formation, which in places has been tilted so that it forms prominent nearly vertical "ribs." Along the northern side of Hogshead Canyon, gently dipping Battle formation is very prominently exposed in steep cliffs rising above the Osgood Mountains quartzite (fig. 10*A*). These cliffs are a prominent topographic feature that can be recognized from far out in the valley to the southeast. Narrow, discontinuous exposures of steeply dipping conglomerate of the Battle formation also occur high up on the steep slopes west of Adam Peak, where the formation is partly cut out by a high-angle reverse fault. A small remnant of conglomerate of the Battle formation resting on shale of the Preble formation forms steep cliffs on Lone Butte, east of the main range in the southern part of the quadrangle.

Except for some small patches of conglomerate, the Battle formation is not found much farther west than the crest of the Osgood Mountains. Near the top of the range east of Goughs Canyon in the north-central part of sec. 26, T. 38 N., R. 41 E., the thick bed of conglomerate beneath limestone can be seen pinching out rapidly westward, and limestone rests directly on the Osgood Mountain quartzite. On the ridge in the north-central part of sec. 4, T. 37 N., R. 41 E., the conglomerate of the Battle formation also wedges out beneath limestone.

LITHOLOGY

In the Osgood Mountains the Battle formation is predominantly a poorly bedded boulder conglomerate; but at some places the upper part is composed of pebble conglomerate, some coarse-grained sandstone, and some interbedded white limestone in beds less than 10 feet thick that generally contain some sandy and pebbly material; and in a few places it contains minor amounts of interbedded red shale and sandstone. The conglomerate is composed almost entirely of fragments of Osgood Mountain quartzite, with only a few small fragments of chert. Most commonly the quartzite fragments are of boulder size, but they range from small pebbles about one-fourth of an inch in diameter to rare blocks 10 feet or more in diameter. In most places the fragments are firmly cemented in a quartzite matrix, which, being of much the same color and texture as the fragments of quartzite, may make it difficult to tell the fragments from the matrix, and in places it is difficult to distinguish the quartzite conglomerate from the Osgood Mountain quartzite. The matrix in some places is dark-reddish-brown to dark-purple sand and shaly material. The basal 2 or 3 feet of the formation at many places are composed of angular fragments of randomly oriented quartzite in a sand and shaly matrix. Where the underlying quartzite is thick bedded or massive, the angular fragments are few and as much as several feet in diameter; where the quartzite is thin bedded, the fragments are abundant and not much more than 1 foot in greatest dimension. This basal zone grades upward, with no apparent break, into poorly bedded boulder conglomerate.

STRATIGRAPHY AND THICKNESS

The Battle formation rests unconformably on the Osgood Mountain quartzite in most places, and on the Preble formation locally. At some places where the beds have been steeply tilted, as along the crest of the range south of Hogshead Canyon, the contact with the older rocks is faulted but the faulting appears to be along the contact plane. The contact between the Battle formation and the Osgood Mountain quartzite over a distance of approximately 2 miles on the north side of Hogshead Canyon is nearly parallel with the contact of the Battle formation and the overlying Etchart limestone, and the Battle formation shows only small changes in thickness due to minor channeling of the underlying quartzite. Minor channeling is also visible at several other places in the range where the contact between the Battle and Osgood Mountain formations is well exposed.

At its type locality in the Antler Peak quadrangle, Roberts (1951) noted that the lithology of the Battle formation changes upward from coarse conglomerate to pebble conglomerate with interbedded sandstone, shale, calcareous shale, and limestone. The Battle formation in the Osgood Mountains shows a similar succession at some places, though at others only coarse-grained conglomerate is present. On Lone Butte—sec. 34, T. 37 N., R. 41 E.—a boulder conglomerate approximately 60 feet thick is overlain by limestone which at its base has thin interbeds of quartzite pebble conglomerate. In the Osgood Mountains, limestone generally rests directly on the older rocks; but in places there are patches of coarse-grained conglomerate beneath the limestone, and the base of the limestone commonly contains lenticular beds of coarse-grained quartzite conglomerate and pebble conglomerate.

Owing to lenticularity and faulting, the thickness of the Battle formation varies considerably. In the Antler Peak quadrangle (Roberts, 1951) the maximum thickness of the Battle formation is more than 700 feet; but in the Osgood Mountains the maximum thickness exposed is approximately 400 feet, and at most places it does not exceed 100 feet.

AGE AND CORRELATION

The Battle formation has been established on fossil evidence as being of Middle Pennsylvanian (Atoka to Des Moines) age at its type locality in the Antler Peak quadrangle. At the type section the Antler Peak limestone of Late Pennsylvanian and Permian age overlies the Battle formation with a slight erosional unconformity. In the Edna Mountains, northwest Golconda quadrangle (Ferguson, Roberts, and Muller, 1952), and at some places in the Osgood Mountains, the Battle formation is overlain by limestone of Middle Pennsylvanian (Des Moines or older) age. Ferguson regarded the Middle Pennsylvanian Highway limestone in the Edna Mountains as the offshore facies contemporaneous with the upper part of the Battle formation at Battle Mountain (Antler Peak quadrangle).

Strata of Late Pennsylvanian age, probably contemporaneous with the Antler Peak limestone, also rest on conglomerate of the Battle formation in the Osgood Mountains. Some conglomerate mapped as Battle formation in the Osgood Mountains may be as young as Late Pennsylvanian, however, because on the main ridge east of Etchart Canyon (SE¼ sec. 4, T. 37 N., R. 41 E.) some thin beds of limestone containing fusulinids of Late Pennsylvanian age (Antler Peak limestone) are interbedded with coarse conglomerate.

Gilluly (oral communication) has mapped conglomerates in the Shoshone Range (Mount Lewis quadrangle) that are correlated with the Battle formation on the basis of lithologic similarity and their stratigraphic position beneath the Antler Peak limestone. In eastern Nevada and western Utah, parts of the Ely limestone are equivalent in age to the Battle and Highway formations (Dott, 1955, p. 2282–2287).

CONDITIONS OF DEPOSITION

A. C. Lawson (1913) studied the conglomerate at Battle Mountain and concluded that it was the remnant of an alluvial fan deposited under arid conditions in a region of bold relief. He, therefore, called it a "fanglomerate," a term that has since had wide use in the designation of deposits which more or less come within Lawson's definition. Since Lawson's study, which presumably was confined to a limited area, the formation has been examined and mapped at many places in north-central Nevada, and although it seems very possible that parts of the Battle formation were terrestrial deposits. much of it certainly is of marine orgin. In general, the basal parts of the formation are poorly sorted and show some crossbedding, current bedding, channeling, and abrupt changes in lithology, whereas the middle and upper units contain a large proportion of marine beds. From our knowledge of the regional distribution and variations of the formation and the changes that are apparent within the Osgood Mountains quadrangle, we conclude that the Battle formation interfingered with carbonate marine strata at many places, a relationship that was first recognized by Ferguson (Ferguson, Roberts, and Muller, 1952) in the Edna Mountains in the northwestern part of the Golconda quadrangle.

ETCHART LIMESTONE

The Etchart limestone is a formation composed predominantly of carbonate rocks exposed along the west side of the central part of the Osgood Mountains; it is named for Etchart Canyon on the west side of the range in secs. 4, 5, and 8, T. 37 N., R. 41 E. (pl. 1), where the formation is well exposed. It contains rocks of Middle Pennsylvanian age and Late Pennsylvanian or Early Permian age that are elsewhere assigned to the Highway (Ferguson, Roberts, and Muller, 1952) and Antler Peak (Roberts, 1951) formations, respectively. The two formations are included in a single unit because we could not separate the Highway limestone from the Antler Peak limestone in the Osgood Mountains quadrangle on the basis of lithology.

DISTRIBUTION

In the Osgood Mountains quadrangle the Etchart limestone is almost entirely confined to the west side of the Osgood Mountains and an area north and east of the Getchell mine (pl. 2). It extends northeast beyond the quadrangle boundary about 4 miles. The formation is not exposed in the Hot Springs Range and has not been identified in the mountains farther west in Humboldt County. It occupies two principal areas in the Osgood Mountains: (1) on the southeastern side of Goughs Canyon, on the upper parts of Perforate Canyon, and in Etchart Canyon; and (2) in a narrow belt on the west side at the north end of the range, north of Anderson Canyon. Limestone beds are prominently exposed on the north side of Hogshead Canyon, where they lie with a northerly dip above conglomerate of the Battle formation and are terminated above by a thrust fault that brings in rocks of the Preble formation (fig. 10A). A limestone section is well exposed in the hills east of the quadrangle boundary, 2 or 3 miles northeast of the Getchell mine. Some of the same beds are poorly exposed in the low hills north of the Getchell mine, in the extreme northeast corner of the quadrangle. An isolated remnant caps the small hill known as Lone Butte, more than 1 mile east of the main range, where it rests on the Battle formation.

LITHOLOGY

The formation is predominantly a limestone and sandy limestone sequence with some interbedded dolomite, minor amounts of calcareous shale, and lenticular beds of conglomerate. Sandy and pebbly units and conglomerate are most common in the lower part, though higher beds may have some thin, pebbly members. The higher strata tend to have more pure limestone and dolomite and commonly some calcareous shale. Bedding is thick and indistinct within most of the units; however, some are well bedded. In general, stratification is best defined by the boundaries between lithologic types. The lithology varies rapidly laterally as well as vertically, and individual units tend to be lenticular and discontinuous, so that it is impossible to make precise stratigraphic correlations between exposures in different areas.

Much of the limestone—perhaps 50 percent or more—is sandy. Although sandy limestone may occur anywhere in the section, it is more common in the lower part. Weathered surfaces are light brown to gray, the unweathered rock varies from very light gray to medium gray. The sandy limestone is medium to coarse grained and generally fairly well sorted, but some beds contain a few small pebbles mixed in with the sand. The majority of the sand grains are subangular to subrounded grains of quartz from 0.1 mm to 2 mm in size; many of the larger grains are quartzite. Most specimens contain a few grains of dark chert, and some contain occasional grains of fresh feldspar. The calcite matrix is medium grained, less commonly coarsely crystalline. Light-gray to light-brown well-rounded quartzite pebbles, whose size range from one-fourth of an inch to cobbles 5 inches in diameter, are scattered through the sandy limestone. The pebbles also occur in a few thin beds which commonly are only one pebble thick. In places the sandy limestone is cherty. The chert is dark gray, weathers brown, and occurs as irregularly shaped lenses and elongated nodules parallel with the bedding. In places the chert is fairly continuous in beds no more than 1 or 2 inches thick. At several localities the limestone is cut by a network of thin quartz veinlets which, with the chert nodules, give it a very rough weathered surface. Poorly preserved fragments of bryozoans, corals, and brachiopods found in the sandy limestone are suitable for only approximate age assignments.

The fairly pure limestone is thick bedded to massive, medium gray to light gray, and commonly distinctly granular. Some of it contains brown to black chert that is commonly nodular but locally forms interbedded layers a few inches to a few feet thick. Some calcareous and dolomitic reddish-brown siltstone and shale a few inches to a few feet thick are interbedded with the limestone. Most of the fossils have been found in the fairly pure limestone, where they are rather poorly preserved owing to transportation and abrasion prior to incorporation in the sediment; they are seldom found in sandy limestone. A very fine grained light-brownish-gray variety of limestone forms single massive beds about 10 feet thick at a few places in the formation.

Dolomite and sandy dolomite are interbedded with the limestone. The dolomitic rocks characteristically weather moderate yellowish brown to light brown or yellowish gray but are greenish gray to medium gray on freshly broken surfaces.

The beds of pebble conglomerate are reddish brown on a weathered outcrop and light gray to pale brown on a fresh surface. Individual beds are generally less than 10 feet and no more than 5 feet thick. Calcareous sandstone and, in places, a bed or two of quartzite 1 foot or less thick are interbedded with the conglomerate. Mostly, the conglomerate is composed of subrounded to rounded, fairly well sorted pebbles and small boulders of light-gray to pale-brown and greenish medium- to coarse-grained quartzite that looks like the Osgood Mountain quartzite; small pebbles of dark chert are much less common. The matrix of the pebble conglomerates is brown sand or silica-cemented quartz sand. Some beds are identical in appearance with conglomerates of the Battle formation.

The calcareous shale beds are thin, rarely exceeding 2 or 3 feet in thickness. Some beds weather yellow brown or greenish gray and are brown to light gray on fresh surfaces; others are grayish red on both weathered and fresh surfaces.

STRATIGRAPHY AND THICKNESS

Along the crest of the Osgood Mountains south of Hogshead Canyon and east of the range crest, the Etchart limestone rests conformably or with slight erosional disconformity on conglomerate of the Battle formation. On the west side of the range the limestone sequence lies unconformably on folded Osgood Mountain quartzite at most places, because the Battle formation lenses out westward and is present only as scattered discontinuous lenses beneath the limestone. Thin beds and lenses of quartzite conglomerate that resemble the Battle formation occur in the lower part of the limestone sequence, and thin beds of limestone are known in the upper part of the Battle formation; so it is possible that the lower part of the Etchart limestone is locally a temporal equivalent of the Battle formation.

No well-defined stratigraphic succession can be established, because of the lateral variations in lithology and lenticularity of units within the sequence. The rocks are involved in thrust faulting, which makes stratigraphic relations even more difficult to resolve. But in general, in the southern part of the Osgood Mountains the lower part of the formation—perhaps the lower one-fourth—is composed of light-gray to light-brown sandy and pebbly limestone containing thin beds and lenses of quartzite conglomerate and occasional thin beds of pure, fine-grained limestone. Following these beds are light-gray to light-brown, sandy, medium- to fine-grained limestone with only scattered quartzite pebbles or thin pebbly beds and some cherty units. The upper part of the formation in many places is predominantly a medium-dark-gray, fine- to medium-grained, thick-bedded to massive, rather pure limestone containing some chert and some beds of dolomite and dolomitic limestone. In some places the highest beds are reddish-brown-weathering calcareous shale and siltstone, and some interbedded thin limestone units.

In the southern part of the Osgood Mountains the maximum thickness of the Etchart limestone is about 250 to 300 feet. On the west side in the northern part of the range there may be 1,000 feet of beds, although faulting and tight folding obscure the true thickness.

In the northeast corner of the quadrangle the formation is separated from the underlying greenstone of the Valmy formation by a thrust fault; but probably it was originally deposited on the greenstone, for in places at the contact the limestone contains fragments of greenstone. The most complete and thickest section is 2 to 3 miles east of the northeast corner of the quadrangle, where possibly more than 2,000 feet of beds are exposed. Here, at least 540 feet and probably about 1,400 feet of strata make up a dominantly carbonate section that grades upward into a section more than 600 feet thick of interbedded calcareous shale, limestone, and dolomitic limestone. The individual units are not distinctive and probably they change rapidly laterally, so that no direct correlations can be made between here and the exposures in the northeast corner of the quadrangle. The measured sections are listed below:

Measured section on isolated hill, SE¼ sec. 27, T. 39 N., R. 42 E., approximately 1.7 miles N. 66° E. from mill stack at the Getchell mine.

Tape-and-compass traverse:	Thickness (ft)
Limestone conglomerate, angular to subangular fragments of limestone of pebble to cobble size and pebbles of quartzite in calcareous and sandy matrix	20
Limestone, gray, massive, cherty	5
Quartzite conglomerate, some pebbles of limestone, chert, and greenstone in sand matrix	8
Dolomite, yellowish- and reddish-brown-weathering, thin-bedded, shelf-making	30
Limestone, gray, rough-weathering; moderately well bedded, with very thin sandy laminæ	10
Limestone, tan, with dolomite laminæ	4
Limestone, light-gray, prominent, cliff-making, rough-weathering, massive, with scattered quartz pebbles	11
Dolomite, tan and gray, brown-weathering, well-bedded; interbedded ribbed gray limestone and medium-gray massive limestone	24
Limestone, gray, shelf-making; some sandy and dolomitic layers	20
Limestone, gray; many thin brown-weathering dolomitic layers; cliff making; contains scattered isolated quartz pebbles. "Hash beds" of small smooth coiled and straight gastropods and brachiopod fragments	14
Limestone, gray platy-weathering, sandy	62
Dolomite	12
Limestone, light-gray, massive	16
Limestone, gray, moderately well bedded to shaly, sand to locally conglomeratic	47
Dolomite, dark-brown-weathering, shelf-making	21
Limestone, medium-gray, massive to thick-bedded; few sandy laminae; contains abundant smooth coiled and straight gastropods and fragments of brachiopods	19
Limestone and dolomite, shelf-making, interbedded	24
Limestone, gray, slightly ribbed, cliff-forming, massive to faintly bedded	9
Limestone and dolomite, interbedded	34
Limestone, gray, cliff-making, medium-grained; few sandy laminae	12
Limestone, medium-light-gray; many sandy laminae	71
Covered	25
Largely covered, except for small outcrops of gray limestone with sandy layers	43
Total	541

Possible thrust fault.

Greenstone of the Valmy formation .... not measured

Measured section on long ridge, in approx. SE¼ sec. 15 and NE ¼ sec. 22, T. 39 N., R. 42 E., approximately 2 to 3 miles northeast from mill stack at the Getchell mine.

	Feet
Pace-and-compass traverse:	
Limestone, dark-gray, thick-bedded, coarse-grained, sandy; some thin lenticular chert beds	110
Covered: Dolomite and limestone; brown-weathering, shaly	35
Largely covered: Pale-red-weathering shaly limestone and brown-weathering dolomitic limestone. Some gray cherty limestone	215
Covered: Float of reddish and brownish shale	162
Covered: Pale red-weathering shaly limestone; some brown dolomitic shale	100
Limestone, pale-red-weathering, shaly, few thin medium-gray beds	32
Tape-and-compass traverse:	
Covered: Pale-red-weathering shaly limestone and thin medium-gray limestone beds	52
Limestone, tan, thick-bedded	154
Covered: Brown sandy dolomite	30
Limestone, medium-gray, thick-bedded to massive	73
Limestone, gray, thick-bedded, somewhat sandy	67
Limestone, yellow- to brown-weathering, sandy; some interbedded gray limestone	67
Covered: Same as above	65
Limestone, gray, cliff-forming, massive; some sandy laminae	12
Covered: Platy limestone float	37
Quartzite conglomerate	4
Limestone, gray, massive; some sandy laminae and 1-ft chert beds	3
Limestone, gray, sandy	55
Limestone, cliff-forming, massive	60
Covered: Limestone?	173
Limestone, medium-gray, massive	14
Covered: Limestone?	28
Limestone, gray; sandy laminae	62
Subtotal	1,610
Break in section; separation or possibly duplication between sections.	
Limestone, medium-gray, massive, sandy; thin sandy laminae	26
Limestone, light-gray, massive, slightly sandy	37
Limestone, medium-gray, thick-bedded to massive; some thin dolomitic interbeds	76
Covered: Brown fine-grained sandy dolomitic limestone	41
Limestone, gray, massively bedded; few thin interbeds of brown calcareous sandstone; 4-inch pebbly bed	69
Limestone and sandy limestone, gray, rather massive; sandy laminae and some chert	114
Covered: Limestone?	62
Subtotal	425
Total	2,035
Possible thrust fault.	
Greenstone of the Valmy formation	not measured

AGE AND CORRELATION

Collections of fossils show that the Etchart limestone ranges in age from Middle Pennsylvanian (Des Moines or Atoke) to Late Pennsylvanian or Early Permian (Missouri and Virgil or Wolfcamp).

In general, most of the collections of Middle Pennsylvanian age come from limestones in Goughs Canyon, where conglomerate of the Battle formation is missing or occurs only locally in thin lenses; collections from limestones resting on the conglomerate were determined to be of Late Pennsylvanian or Early Permian age. Exceptions to these generalizations were found, however, and Late Pennsylvanian fossils have been collected from beds near the base of the section that cannot be distinguished from strata containing Middle Pennsylvanian forms.

Furthermore, much of the fossil material is abraded fragments which, with the lithology, indicate that they were introduced into the sediments as clastic particles and therefore are not necessarily of the same age as the enclosing rock.

Several collections from the Etchart limestone in Goughs Canyon have yielded forms that were identified as Middle Pennsylvanian in age. Helen Duncan, James Steele Williams, and Mackenzie Gordon, Jr., of the Geological Survey made the fossil determinations. Miss Duncan reported on the bryozoans and corals, and Mr. Gordon and Mr. Williams studied the brachiopods. The following notes are taken from their reports on collections from the Etchart limestone in Goughs Canyon:

USGS Collection No. 19809–PC (field No. H–9–52). Goughs Canyon, hill near center NE¼ sec. 27, T. 38 N., R. 41 E. Medium dark-gray limestone containing bryozoan, coral, and a few brachipod remains.

Miss Duncan reported the following forms (written communication, 1953):

- Horn coral fragment, indet.
- *Rhombotrypella* sp.
- *Fenestella* sp.
- *Polypora*, 2 spp.
- *Archimedes* sp. indet. (1 fragment)
- *Cystodictya* sp.
- *Rhabdomeson* sp.
- "*Ascopora*" sp.

Of the bryozoans, she said (written communication, 1953):

Cystodictya is extremely abundant at this locality, and *Fenestella* and *Polypora* are abundant in some beds. Very few examples of *Rhombotrypella* were found. The species of *Cystodictya* is considerably smaller than the one found in the Mississippian at the Goughs Canyon locality (USGS loc. 14206–PC), and the *Rhabdomeson* in these lots is distinctly different from the species found in the Goughs Canyon formation or the one found in the Antler Peak limestone. The bryozoan called

"*Ascopora*" is also different from the *Ascopora* found in collections from the Antler Peak and from beds that I think are probably Antler Peak equivalents. The makeup of this bryozoan faunule is obviously quite different from that found in the Antler Peak limestone at its type locality or at Edna Mountain.

I consider the fragment of *Archimedes* to be rather significant. In this country, the genus is confined to the Morrowan and older rocks in the central interior region, but it occurs in Middle Pennsylvanian rocks in north-central Utah (Oquirrh formation). In Russia it occurs in the Middle and Upper Carboniferous (C_{II} and C_{III}) and Permian but apparently not in older rocks. Of course, one cannot absolutely depend on the genus being restricted to the Middle Pennsylvanian of this region, but present knowledge of its occurrence suggests that these beds are likely to be of Middle Pennsylvanian age.

Miss Duncan further stated (written communication, 1961):

All described North American species correctly assigned to *Cystodictya* came from rocks that are older than Late Pennsylvanian; and in 1953, it was generally assumed that the genus did not range above the Middle Pennsylvanian. More recent work on bryozoan faunules obtained farther south in Nevada has revealed that *Cystodictya* occurs in association with Early Permian fusulines. Inasmuch as we now have good evidence that the genus has a longer range than was previously thought, the occurrences of *Cystodictya*-like forms is faunal assemblages of Late Pennsylvanian and Early Permian age is not so anomalous. The Etchart limestone bryozoan faunule from USGS locality 19809–PC seems to be older than the Late Pennsylvanian faunule known from the lower part of the Antler Peak limestone; however, one cannot be certain on objective evidence that it is not Late Pennsylvanian.

Mr. Williams (written communication, 1953) said:

The collection contains a fragment of a brachiopod that might be a piece of *Meekella*?, another fragment that is probably a *Hustedia*? sp. indet., but that does not show the punctate structure, typical of the genus, an indeterminate *Chonetes*?, remains of a spiriferinoid brachiopod and a fragment of a very large coarse-ribbed *Dictyoclostus*? sp. indet. The age is Carboniferous or Permian and the general assemblage looks to me to be of Late Pennsylvanian or Early Permian age, but I cannot be sure of which, if in fact either. The *Hustedia* does not seem to be the same species, H. *phosphoriensis* Branson, that I found in the Edna Mountain. Whether it is another species of the same age or one of the species of *Hustedia* that occur in both the Upper or Middle Pennsylvanian or in older rocks, I cannot tell from the specimen. The large *Dictyoclostus* is of a general type that I would expect more to find in Upper Pennsylvanian or Lower Permian; but it is represented by only a fragment and cannot be specifically identified, and a few instances of the occurrence of somewhat similar forms in middle Pennsylvanian rocks are known. I would be inclined to defer to Miss Duncan's analysis of the significance of the Bryozoa as she seems to have more diagnostic material than I have.

Collection No. USGS 19815–PC (field No. 52–W–22). Southeast corner SW¼NE¼ sec. 27, T. 38 N., R. 41 E. Gray limestones containing brachiopod fragments.

- Caninoid corals (fragments of a small species).
- Stenoporoid bryozoan, sp. indet.
- Fenestellid bryozoans, genus and sp. indet.
- "*Ascopora*" sp. cf. species in collection No. 19809–PC.
- Other rhomboporoid bryozoans, indet.
- *Cystodictya*? sp. indet.

Helen Duncan (written communication, 1953) said:

I am fairly certain that the bryozoan I am calling '*Ascopora*' in this collection is the same species that occurs at USGS locality 19809–PC, where I think the rocks are of Middle Pennsylvanian age. This species is very small but quite distinctive, and it certainly is entirely different from the *Ascopora* I have identified in the Antler Peak. The fenestellids and the probable *Cystodictya* also are more suggestive of the fauna that is presumably of middle Pennsylvanian age than they are of the Antler Peak limestone fauna as now known.

Collection No. USGS 19812–PC (field No. 52–W–27) SE corner W½SW¼ sec. 23, T. 38 N., R. 41 E.

- *Fenestella* sp.
- *Polypora*, 2 sp.
- *Cystodictya* sp.
- *Rhabdomeson* sp.
- *Rhomboporoids*, genera indet.

Of this collection, Miss Duncan (written communication, 1953) said:

This bryozoan assemblage contains species that appear to be identical with the species that are found in some other collections that are believed to be of Middle Pennsylvanian age. They are certainly very much like the bryozoans that were found at USGS locality 19809–PC.

The strata that contain fossils suggestive of Middle Pennsylvanian age probably are equivalent to the Highway limestone which Ferguson (Ferguson, Muller, and Roberts, 1951) described in the Edna Mountains, northwest Golconda quadrangle. The Highway limestone contains fossils of Middle Pennsylvanian age, and Ferguson regards it as the offshore facies of the upper part of the Battle formation.

Other beds, indistinguishable lithologically from strata containing Middle Pennsylvanian fossils, have yielded fossils that have been determined as Late Pennsylvanian, partly equivalent in age to the Antler Peak limestone.

Collections of fusulinids from two localities in the southern part of the range, where limestone either rests on Battle type conglomerate or is interbedded with it, were determined as Late Pennsylvanian by L. G. Henbest and R. C. Douglass (written communications 1952, 1953, 1961):

Collection No. f–9471. North side of hill, east part, SW¼ sec. 8, T. 37 N., R. 41 E. Rests on Battle type conglomerate.

- *Climacammina*? sp.
- *Endothyra* or *Endothyranella* sp.
- *Bradyina* sp. (small, thin-shelled form)
- *Ozawainella*?? sp.
- *Waeringella*?? sp.
- *Triticites* sp.

The preservation is very poor. The determination of the specimens of *Waeringella*?? is extremely uncertain. The species identified as *Triticites* is determinable with fair assurance as to genus; but the wall structure is poorly preserved, and it is barely possible but seemingly unlikely that it may actually represent a species of *Fusulinella* of Atoka (lower middle Pennsylvanian) age.

The species of *Triticites* is probably not older than the middle part of the Missourian and not younger than the earliest third of the Virgil. Early Virgil or late Missourian seems most likely. Permian age seems to be definitely ruled out.

Collection No. f–9514a. From ridge a little east of center SE¼ sec. 4, T. 37 N., R. 41 E. Interbedded with conglomerate.

The microfauna includes:

Climacammina sp.
Bradyina sp.
Tetrataxis sp.
Triticites sp.

Mr. Douglass (written communication, 1961) believes that this collection is of Late Pennsylvanian age (Missouri).

Another fusulinid locality is a small remnant of limestone and conglomerate east of the Osgood range near the south border of the quadrangle. Henbest reported (written communication, 1955):

Collection No. f–12070a. Nevada, Osgood quadrangle; SW¼ NW¼ sec. 8, T. 36 N., R. 41 E.; approximately 600 feet southeast of powerline road.

Endothyra sp.
Bradyina sp.
Globivalvulina? sp.
Millerella-like foraminifer
Schubertella or juvenarium of microspheric generation of *Triticites* (?) sp.
Ozawainella sp.
Triticites cf. *T. secalicus* (Say)
Triticites aff. *T.* ventricosus (Meek and Hayden)

According to Mr. Douglass (written communication, 1961) this collection is also Late Pennsylvanian but younger than collection f–9514a, and represents Virgil equivalents. Neither f–9514a nor f–12070a suggests Permian age.

Two types of corals are also abundant at this locality (USGS 15778–PC)—a large caninoid and a syringoporoid. According to Miss Duncan (written communication, 1956):

The caninoids were broken up, decorticated, and eroded before deposition. They cannot be accurately identified but resemble generally the caninoids that occur in the Antler Peak limestone on Antler Peak and at other localities. Most of the horn corals collected are examples of these poorly preserved caninoids, but I found one specimen that I think probably belongs to the genus *Gshelia*. I do not recall having identified this genus in other collections from North America. The genotype came from late Upper Carboniferous (C_{III}) beds in Russia that are probably the temporal equivalent of rocks classed as Early Permian (Wolfcamp) in the United States.

The syringoporoid corals are crudely silicified and their internal structures mostly destroyed. A few random tubes exhibit structures that indicate relationship with *Syringopora*, but the species in hand is certainly not *Syringopora multattenuata* McChesney, the species that has been found in the Antler Peak limestone at several places. At present this particular form of syringoporoid furnishes no good evidence on age of the beds.

By themselves, none of the corals in this lot provide good objective evidence for dating the collection. However, comparable species have not been found in any collections that we know are Battle or Middle Pennsylvanian in the region, and the large caninoids do occur in the Antler Peak at several localities in association with the forms of fusulinids that were identified by Henbest in this collection. Therefore, the caninoids might be used as presumptive evidence that the rocks are Antler Peak equivalents.

In Goughs Canyon a limestone bed a few feet above sandy limestone that rests on Osgood Mountain quartzite, and which is like beds from which Middle Pennsylvanian fossils were obtained, contains Late Pennsylvanian fossils:

Collection No. USGS 19810–PC (field No. 52–W–20). Center, sec. 28, T. 38 N., R. 41 E.

Bryozoans:
Fistuliporoid sp. indet.
Fenestella sp. indet.
Polypora? sp.
Penniretepora? sp.
Cystodictya? sp.

Brachiopods:
Wellerella? cf. *W. osagensis* (Swallow)
Ambocoelia? sup. indet.
Cleiothyridina? sp. indet.
Chonetid brachiopod, gen and sp. indet.
Hustedia, prob. 2 species, one which is close to *H. phosphoriensis* Branson.

Miss Duncan (written communication, 1953) at first believed that the bryozoans were probably of Middle Pennsylvanian age, but on restudy of the collection said that she had seen the same sorts of bryozoans in collections (from elsewhere) known from other fossil evidence to be of Antler Peak age. Mr. Williams said (written communication, 1953) that although data from the brachiopods were insufficient to make a positive age reference, what data there are faintly suggest the possibility of Late Pennsylvanian or of Early Permian age.

A collection of bryozoans from limestone above Battle type conglomerate was identified as probably of Antler Peak age by Miss Duncan (written communication, 1953).

Collection No. USGS 15777–PC (field No. 52–W–25). Just north of the saddle north of center, sec. 26, 26, T. 38 N., R. 41 E.

The rock is dolomitic and the bryozoans are crudly silicified. One form is a rhomboporoid that suggests *Rhabdomeson*. Another specimen is a fenestrate type with a coarse network; the size of the branches suggest *Polypora*, but finer structures are destroyed and its generic identity cannot be proved. A third type is represented by cross sections that are comparable to those of the *Hexagonella* I found in the Antler Peak limestone (Roberts' N unit).

This meager and poorly preserved bryozoan assemblage is more like that found in the Antler Peak than like the fauna that occurs in the older rocks.

Westward a few thousand feet, the thick conglomerate beneath limestone from which this collection came pinches out, so that along the strike, this limestone would lie with no distinguishable hiatus above limestone beds which rest on the Osgood Mountain quartzite and from which fossils identified as Middle Pennsylvanian have been collected.

The limestone unit in the northeast corner of the quadrangle is continuous to the east with a thick section from which fossils definitely identified as Late Pennsylvanian to Early Permian have been collected. Here no Battle type conglomerate underlies the section and no Middle Pennsylvanian limestones have been recognized. The strata have been dated as equivalent in age to the Antler Peak limestone.

Miss Duncan (written communication, 1956) reported on the bryoza:

Collection No. USGS 15777–PC (field No. FH–22–54). From approximately 2.6 miles N. 37° E. from stack at Getchell mill; in NE¼ sec. 18, T. 39 N., R. 42 E. Limestone, approximately 550 ft. above base.

This collection contained a large representation of very poorly preserved bryozoans. The following genera have been identified:

Hexagonella sp.
Rhomboporella sp.
Rhabdomeson sp. (abundant)
Other indeterminate rhomboporoid bryozoans
Fenestella sp.
Polypora sp.
Septopora sp.
Penniretepora sp.

In general aspect this bryozoan assemblage suggests Late Pennsylvanian or Early Permian in this region. Coarse-meshed fenestrate genera, such as occur in this collection, appear to be especially characteristic of this faunal zone. The presence of *Hexagonella* also is interpreted to indicate post-Des Moines age. Even without the supporting evidence provided by the brachiopods, reported on by Mackenzie Gordon on December 15, 1955, I should have felt it was relatively safe to determine the collection as Antler Peak equivalent.

Mr. Gordon reported (written communication, 1955) the following forms:

Spirifer (*Choristites*?) sp.
Neospirifer cf. *N. texanus* (Meek)

The *Spirifer* resembles the species which is abundant in the lower part of the type Antler Peak section. The *Neospirifer* has been compared with Texas specimens of *N. texanus* from the Graham formation (Virgilian), Texas and looks close enough to be conspecific.

A collection of mollusks from a nearby section, approximately 250 feet above the lower contact, was reported by E. L. Yochelson and R. L. Batten (written communication, 1954):

Collection No. USGS 15381–PC (field No. FH–20–54. 1.7 miles N.66° E. of stack at Getchell Mill; SE¼ sec. 27, T. 39 N., R. 42 E.

The collection appears to be of Permian age, but the possibility of Permian? age cannot be ruled out. The following fossils have been identified:

Plagioglypta canna (White)
?*Schizodus* species indet.
Bellerophontid species
Glabrocingulum new species

Paleozoic scaphopods are poorly known, but *Plagioglypta canna* is relatively common. To date it has been reported only in rocks of Permian age. There are no reports of the species in beds classified as ?Permian by the Survey, but the scaphopods of these have not been studied.

The gastropod *Glabrocinglum* is relatively high spired and lacks reticulate ornamentation. These characters are suggestive of Permian rather than Pennsylvanian on the basis of our present knowledge of the genus and the specimens available for study.

ADAM PEAK FORMATION

A predominantly clastic sequence including shale and siltstone, dolomitic sandstone, chert, and limestone that overlies a coarse conglomerate along the west side of the Osgood Mountains has been separately mapped and named the Adam Peak formation for Adam Peak, the highest peak in the range. Fossil evidence indicates that the rocks are Late Pennsylvanian to Early Permian in age, though a younger age for the upper part of the unit is not definitely ruled out.

DISTRIBUTION

The Adam Peak formation occupies a belt approximately 2½ miles long on the high, western slope of the range west of Adam Peak, from the head of the East Fork of Eden Creek on the north to the ridge east of Goughs Canyon on the south (pl. 1). The formation also occurs as fault-bounded blocks along the crest of the range from Goughs Canyon to Perforate Canyon. The formation is on a thrust plate that overrides strata of the Etchart limestone at the south end of the northern belt and along the west side of the southern blocks. The northern belt is bounded on the west by a thrust plate of the Harmony formation. A belt of conglomerate mapped as Battle formation occurs with some interruptions along the east side of the belt. The conglomerate is in contact with the Preble formation along a steep reverse fault which in places cuts out the conglomerate and at the north end flattens into a thrust that carries the Preble above the Adam Peak formation.

LITHOLOGY

The Adam Peak formation is chiefly shale and siltstone, dolomitic clastic rocks, and chert but includes some limestone and dolomite.

The shale and siltstone weather yellowish brown to olive gray and are light olive gray to dark gray on freshly broken surfaces. Some of the shale is slightly calcareous, much is slightly dolomitic, and some of the siltstone is dolomitic. The principal silt-size particles are quartz; in addition, a few percent of chert fragments, less than 1 percent of fresh plagioclase, and a small number of grains of subrounded zircon and, less commonly, green tourmaline have been seen. Most of the matrix is too fine for positive identification, but it contains shreddy, colorless mica in abundance. Some specimens of the shale and siltstone have flat, twisted impressions on bedding planes that suggest worm trails.

Distinctive sandstone and dolomitic sandstone make up more than 50 percent of the section where it is thickest. Unweathered surfaces are medium dark gray, medium gray to olive gray, and light olive gray; the weathered rock is brown, generally pale yellowish brown. Grain size ranges from very fine to coarse, but in most exposures the rock is fine grained. On the whole, the size sorting is poor; but thin-section studies show that the sorting in some specimens is good. The fragments are angular to subangular and a few are subrounded. Grains of quartz and quartzite are most plentiful, with minor amounts of detrital chert and shale; a few small, subrounded to rounded grains of zircon and tourmaline are visible in all thinsections. The phosphate, collophane, is a characteristic constituent, and though present in most specimens examined, it does not amount to more than 2 to 3 percent, commonly less. It occurs as small, brown, weakly birefringent to isotropic pellets and irregularly shaped bodies in the matrix. Many specimens contain fine-grained apatite that seems to be inversely proportionate in amount to the amount of collophane, which suggests that it is reconstituted collophane. The clastic fragments are loosely packed but firmly bonded by the matrix, which is composed of microcrystalline silica and dolomite in various proportions, and a few percent of sericite. In some specimens, dolomite is the chief cement; in others, it forms scattered, individual rhombs.

In the upper part of the section, chert is common. The chert is grayish black to brownish black and varies from massive to thin bedded; most commonly, it is massive to thick bedded. Some thin beds of dark-gray, brown-weathering dolomite, dark siliceous shale, and dolomitic sandstone are interbedded with the chert.

Limestone occurs as thin beds and lenses that are commonly more or less sandy and may locally be pebbly. One thin bed in the lower part of the section is a dark-gray bioclastic limestone composed predominantly of the broken remains of fusulinids and bryozoans. Light-gray, commonly brown-weathering, fine-grained dolomite forms beds several tens of feet thick interbedded with dolomitic sandstone.

STRATIGRAPHY AND THICKNESS

The Adam Peak formation rests with apparent conformity on coarse-grained conglomerate of the Battle formation on the crest and west side of the Osgood Mountains. West of Adam Peak, where the section is most complete, the formation is divisible into three members: a lower member of interbedded shale, limestone, and dolomitic sandstone; a middle member of chert, dolomite, and interbedded shale and chert; and an upper member of dolomite and dolomitic sandstone. The beds dip steeply west and are locally overturned to the west, and although there is probably some unrecognized tight folding, the beds are generally younger to the west.

The lower member is approximately 850 feet thick. Its lower part is predominantly shale and siltstone but includes thin interbeds and lenses of limestone; these rocks give way upward to predominantly dolomitic sandstone and shale. The middle member, about 950 feet thick, is composed of about 700 feet of interbedded shale and thin-bedded chert, but dark, thick-bedded to massive chert prevails in the top 200 feet or so of the member. The upper member consists of fine-grained dolomite and dolomitic sandstone having an exposed thickness of about 300 feet; however, its true thickness is unknown because of cutting out by the overriding Harmony formation. Thus, the exposed section is on the order of 2,100 feet thick.

AGE AND CORRELATION

Fossils indicate that the Adam Peak formation is Late Pennsylvanian to Early Permian in age and is contemporaneous in part with the Etchart limestone in the Osgood Mountains and the Antler Peak limestone of the Battle Mountain region.

A bed of bioclastic limestone approximately 250 feet above the base of the section west of Adam Peak contains fragments of fusulinids and bryozoans that have been identified as Late Pennsylvanian or Early Permian. Mr. Henbest reported (written communication, 1953) on the fusulinids as follows:

Collection F-9708 Nevada. Osgood Mountains quadrangle. On knob in center SW¼SW¼ sec. 13, T. 38 N., R. 41 E. (field No. 52-W-48).

Triticites sp.
Schwagerina sp. (early form)
Schwagerina sp.

The fusulinid fauna in this clastic limestone is represented by numerous more or less fragmentary specimens and by a smaller number of entire specimens. The fauna is more nearly like that of the lower or middle part of the Wolfcamp than that in any later part of the West Texas section. It definitely represents a younger faunal stage than that of the Virgil (Pennsylvanian) but older than that in the upper (Mill Canyon) and lower (Jory) quartzite members of the Havallah formation, Antler Peak area, recently reported to Roberts (that is, late Hueco or Leonard).

Whether it correlates with the Antler Peak formation or not is very uncertain. One nearly entire specimen of *Triticites* in this collection resembles the small ventricose species of *Triticites* that characterizes the Antler Peak collections f9653 and f9661, evidence of doubtful value for close correlation.

Of the bryozoans, Miss Duncan said (written communication, 1953):

The bryozoan genera I am able to recognize in collection No. USGS 19814-PC (field No. 52-W-48) include the following:

- *Meekoporella* sp.
- *Hexagonella*? sp.
- Other fistuliporoids, genera undet.
- *Stenodiscus* sp. (laminar form)
- *Rhombotrypella* sp.
- Stenoporoid bryozoan (incrusting (?) form)
- *Fenestella* sp.
- *Rhombopora* sp.
- *Ascopora* sp.

This assemblage is characteristically found in rocks of Pennsylvanian and Hueco age. It certainly suggests some part of the Antler Peak to me. I have not seen any bryozoans from rocks we are sure are Edna Mountain formation and very much doubt that, if they do occur, they would be sufficiently well preserved to identify. The assemblage listed above does not look like any Phosphoria (with which the Edna Mountain is correlated) bryozoan faunule I have seen, and it does not resemble that of the Kaibab. Neither does the faunule correspond to the one that I think is possibly of Middle Pennsylvanian age in this part of Nevada.

A collection of brachiopods and bryozoans was obtained from dolomitic sandstone in the upper part of the lower division of the Adam Peak formation.

USGS locality 19816-PC (field No. H-44-52). SW¼SW¼ sec. 13, T. 38 N., R. 41 E., on ridge, brachiopods, Bryozoa, *Conularia*?, from brown weathering gray dolomitic sandstone.

Mr. Williams wrote (written communication, 1953) concerning the brachiopods:

This collection contains several large specimens of one or more species of a large productid, possibly belonging to the genus *Juresania* or to a closely related genus. It also contains indeterminate remains of one or two species of an undetermined type of productid, a fragment of a productid that is probably a linoproductid, and incomplete remains of a rhynchonellid or a terebratuloid brachiopod and of a large chonetid brachipod. Other brachiopods are represented by specimens that cannot be classified even as to type. Some crinoid columnals are in the collection. No conularias were seen. If present they may have been inadvertently destroyed by preparing. The conularias are, in general, not of very much use stratigraphically.

Except for the specimens tentatively referred to *Juresania*, the shells have been so altered by replacement or broken or "skinned" in freeing them from the matrix that even those retaining a fairly large portion of the shells do not have enough of the ornamentation preserved to provide a basis for identification. The juresanias in this collection are very similar to ones I have seen in the Antler Peak formation (and the accompanying fragmentary specimens do not dispute an Antler Peak age for the collection), but closely related species of *Juresania* occur in both younger and older rocks. The collection could be Antler Peak, but it could also belong to a younger or older part of the late Paleozoic section.

The bryozoans from this collection were also inconclusive, according to Miss Duncan (written communication, 1953):

This lot contains a few indications of twiglike bryozoans. The specimens are crudely silicified and not identifiable. In form they suggest some type of rhomboporoid. Bryozoans of this general group are common in the Antler Peak but unknown in the type Edna Mountain. Bryozoan evidence on the age of the rocks at this locality is entirely inconclusive, for rhomboporoids have a considerable range in the Paleozoic.

The dolomitic sandstones and shales of the Adam Peak formation resemble the Edna Mountain formation of Permian age, which rests with an erosional unconformity on the Antler Peak limestone in the northern part of the Golconda quadrangle (Ferguson, Roberts, and Muller, 1952). The fossil evidence, however, precludes correlating the Adam Peak with the Edna Mountain formation. The Adam Peak is regarded as equivalent in age to part of the Etchart and Antler Peak limestones and possibly represents a locally developed facies.

CONDITIONS OF DEPOSITION OF THE ETCHART LIMESTONE AND ADAM PEAK FORMATION

The variability of the Etchart limestone, its abundant content of clastic material, and its relations to the Battle formation are indicative of deposition in a shallow-water marine environment. The lower beds probably are the offshore marine equivalent of part of the Battle formation, but higher beds overlap the westward-thinning conglomerates of the Battle and represent the landward encroachment of the sea. A marine environment is also indicated for the Adam Peak formation, but in contrast to that of the Etchart limestone its basin of deposition received more sand and silt. The wide variation of these sedimentary rocks of Pennsylvanian and Early Permian age indicate that they were laid down in environments that also differed widely from place to place, as might be expected in an archipelago.

ROCKS OF PENNSYLVANIAN(?) TO PERMIAN(?) AGE

FARREL CANYON FORMATION

A sequence of sandstone, shale, chert, metavolcanic rocks, and small amounts of limestone in the northwestern part of the Osgood Mountains (pl. 1) is here named the Farrel Canyon formation. Farrel Canyon is on the northwest side of the Osgood Mountains, approximately 1 mile south of Anderson Canyon, in the northeastern part of T. 38 N., R. 41 E., and the southeastern part of T. 39 N., R. 41 E. The formation is not well exposed, and its lithology and stratigraphic succession are poorly known.

DISTRIBUTION

The Farrel Canyon formation extends along the west side of the Osgood Mountains from the ridge south of Cave Canyon to the north end of the range. It continues around the north end of the range, about 2 miles north of the quadrangle boundary, and continues for at least 3 miles in an east-northeasterly direction from the northeast corner of the Osgood Mountains quadrangle. Rocks of somewhat similar lithology are also known at the north end of the Hot Springs Range in the Hot Springs Peak quadrangle. In the Osgood Mountains the formation is bounded on the west and north by alluvium and Tertiary volcanic rocks. Thrust faults and high-angle reverse faults limit it on the east and south.

LITHOLOGY

The Farrel Canyon formation is an interbedded sequence of sandstone, chert, shale, siltstone, and some volcanic flow rocks and pyroclastics. Individual units are lenticular and variable.

One of the most distinctive features of the formation is the sandstone, which may constitute 25 percent of the formation. It commonly is thick bedded and alternates with beds of shale and lenses of chert. Most of it is fine grained, some is medium grained, and some beds are pebbly in places. The sandstone is firmly lithified, tough, and hard to break with a hammer. Fresh specimens are medium gray to light brown, but weathered surfaces are pale yellowish brown to moderate brown and olive gray. Detrital grains chiefly of quartz and scattered fragments of gray to brown chert give the rock its characteristic appearance in hand specimens. Poorly sorted angular to subangular mineral and rock fragments are set in a matrix of mica, chlorite, clay, and silt-size quartz and feldspar, which ranges from 30 percent to 55 percent of the rock. The coarser grains are chiefly quartz, some chert, a few grains of fresh feldspar, and, uncommonly, a few pieces of shale. Most of the quartz grains are single individuals, commonly strained, though a few are combined grains, possibly quartzite; the feldspar includes plagioclase and untwinned K-feldspar. Reaction between the grains and the matrix has corroded the grain boundaries. "Heavy" minerals amount to a small fraction of a percent and are mostly well rounded zircon—some clear, some of the pinkish "hyacinth" variety—and grains of magnetite and tourmaline. The matrix is composed of shreds of colorless mica (sericite), a pale greenish-brown to yellowish clay mineral (possibly nontronite or hydrobiotite), and very pale chlorite, plus silt-size quartz and feldspar. Some specimens contain a little carbonate, presumably calcite, in the matrix. Brown iron oxide stains the rock along minute fractures and along grain boundaries. The pebble conglomerate has a similar composition; however, most of the pebbles are chert and some quartzite, in a sandy matrix. The sandstone is classified as a lithic graywacke (Pettijohn, 1954, p. 363–364) or quartz graywacke (Gilbert, in Williams, Turner, and Gilbert, 1954, p. 292, 294).

Shale and siltstone are interbedded with the sandstone and chert in units that are seldom more than a few tens of feet thick. Most of these fine-grained clastic rocks are silty shale and siltstone that weather olive gray and are dark greenish gray, light brownish gray, and dark gray on fresh surfaces. Most specimens have a hackly fracture rather than a well-developed shaly parting.

Chert may amount to 50 percent of the formation. Thick lenses are interbedded with the sandstone; some sections are mostly chert with some interbedded dark shale, and commonly chert is associated with altered volcanic flow rocks. In many exposures the chert is well bedded to thin bedded, individual strata ranging from 1 inch to 4 or 5 inches in thickness and having interbeds of sandy or silty shale 1 inch or less thick. In many exposures the chert is tightly and complexly folded. The generally thin-bedded nature of the chert distinguishes it from chert of the Goughs Canyon formation. It also has a lighter appearance than chert in most of the other formations, except the Valmy. Some of it is dark gray to black, but in many exposures it is a light brown, light brownish gray, light greenish gray, and greenish gray, and some is very light gray. Some of the chert was examined microscopically, and most specimens revealed nothing but cryptocrystalline silica with some faintly visible structure that suggested remnants of micro organisms, such as sponge spicules or spines of radiolaria, though one thin section showed an abundance of circular bodies that presumbly were radiolarians. A few specimens, however, contain material suggestive of volcanic origin, including angular to subangular quartz fragments, some fresh plagioclase and hazy, altered feldspar, plus many shreds of sericite and flakes of nontronite(?) in a matrix of cryptocrystalline silica.

Altered volcanic rocks are interbedded with the chert at a few places. The units of volcanic rock seem to be no more than a few tens of feet thick, discontinuous, and probably lenticular. Volcanic rocks probably make up somewhat less than 25 percent of the formation. They are fine-grained to aphanitic nonporphyritic dark-gray, medium-gray, or greenish-gray rocks. Some are obviously fragmental evenin hand specimens; the clastic nature of others is only apparent from thin-section study.

Typical specimens of the flow rock have a microscopic intersertal texture and are composed of albitic plagioclase and interstitial chlorite and a brown clay mineral, possibly nontronite. Additional secondary minerals are sericite, leucoxene, and calcite. Presumably, these originally were flow rocks of intermediate composition, possibly andesites. The pyroclastic rocks were crystal-vitric tuffs and crystal tuffs. In some, relicts of original glass shards are visible, now devitrified and crystallized to an untwinned potassium feldspar (probably originally sanidine), in a cryptocrystalline groundmass of silica. Also embedded in the groundmass are crystal fragments of quartz, plagioclase, untwinned mottled-appearing potassium feldspar that may have been sanidine originally, and minor amounts of sericite, some chlorite, and rare epidote. Other specimens contain crystal fragments and no relicts of shards. A few fragments of microcrystalline volcanic rock, quartzite, and chert were also recognized. The mineral composition suggests that originally these pyroclastics were of rhyolitic or dacitic composition.

From the microscopic studies it is concluded that these volcanic rocks have been affected by low-grade retrogressive metamorphism, which may have been caused partly by deuteric alteration at the time of extrusion and deposition, and party by burial and involvement in subsequent diastrophism.

In addition to the rocks that are obviously volcanic, some dark-gray, streaked, microcrystalline rocks that look like argillites in hand specimens probably are also of volcanic origin. They are fine grained and have a streaked or sheared appearance, and the individual components are difficult to resolve even with microscopic study. Mostly, they are composed of lenticular bodies of cryptocrystalline silica in a brownish mass of clay minerals, sericite, chlorite, and epidote, and may have been pyroclastic rocks that have been intensely sheared.

STRATIGRAPHY AND THICKNESS

Neither the top nor the bottom of the Farrel Canyon formation is exposed, and exposures are so poor that its structure and stratigraphy cannot be worked out. In the western belt, south of Anderson Canyon, there is a general succession from east to west of chert and interbedded metavolcanic rocks that grade westward and possibly upward — through interbedded chert, shale, and some sandstone—into a predominantly sandstone unit on the west. At the north end of the range, however, the formation seems to consist of interbedded chert, sandstone, and some shale, with no clearly recognizable succession of beds. The thickness of the formation is unknow, but it seems likely that it is at least several thousand feet thick. Along the west side of the range the exposed thickness must be on the order of 3,000 to 4,000 feet, and perhaps twice this amount is represented by the exposures across the north end of the range.

POSSIBLE AGE AND CORRELATION

No fossils have been found in the Farrel Canyon formation, and its stratigraphic position relative to the other sedimentary rocks is unknown. Rocks of somewhat similar lithology exposed in the north end of the Hot Springs Range have yielded some rather poorly preserved fossils which range from Mississippian to Permian; however, experience has shown that long-range lithologic correlations in the western part of the Great Basin are hazardous. The formation contains rock assemblages that resemble parts of both the Pumpernickel and Havallah formations of the Sonoma Range (Ferguson, Roberts, and Muller, 1952) and Antler Peak quadrangle (Roberts, 1951). The Havallah formation ranges in age from Middle Pennsylvanian (Atoka) to Permian. The Pumpernickel formation is Pennsylvanian(?) (R. J. Roberts oral communication). Without fossils and sound stratigraphic evidence, the age of the Farrel Canyon formation must remain unknown; however, it is tentatively assigned a Pennsylvanian(?) to Permian(?) age.

INTRUSIVE BASALT OF UNCERTAIN AGE

A small body of basaltic rock is exposed in the central part of sec. 20, T. 39 N., R. 42 E., in the northern part of the Osgood Mountains (pl. 1). Another body of similar size is also known beyond the quadrangle boundary in the southern part of sec. 18, T. 39 N., R. 42 E., in the Hot Springs Peak quadrangle.

The fresh rock is dark gray to black, very fine grained to microcrystalline, and in places is slightly porphyritic and contains small plagioclase phenocrysts in an aphanitic groundmass. Locally, the rock has small amygdules as much as 3 mm in diameter. Plagioclase occurs in randomly oriented laths that average 0.2 mm in length, and in a few small (0.7–0.9 mm) phenocrysts. The plagioclase composition is uncertain, but its optical properties as determined from thin-section study suggest sodic oligoclase or albite-oligoclase. Fibrous pale-green actinolitic hornblende is interstitial to the plagioclase and also forms small squarish phenocrysts. Probably it is an alteration product of pyroxene. In addition, minute grains of magnetite are plentiful, and a little chlorite is present, mostly as partial replacements of the central parts of plagioclase phenocrysts. In some specimens the interstitial material is so fine grained that it cannot be clearly resolved with the microscope, but in one thin section it appeared to be a mixture of pale-brown biotite and minute rods of colorless actinolitic amphibole. Amygdules in some specimens are filled with quartz and rimmed with pale-brown biotite; in others, small amygdular bodies are filled with chlorite. Petrographically, the rock appears to be a basalt which has been subjected to low-grade metamorphism, with resultant change of the plagioclase to a sodic variety, and replacement of original pyroxene to actinolitic amphibole. Chemical analysis of a sample of the rock (table 7) also suggests that it was a basalt. This is indicated by the low silica content, absence of normative quartz, and negative qz value. The low CaO:Na_2O value, relatively high H_2O, and moderate CO_2 reflect the low-grade metamorphism the rock has undergone.

TABLE 7.—*Analysis, norm, and Niggli values of intrusive basalt of uncertain age*

Northern Osgood Mountains; extreme SW¼SE¼ sec. 18, T. 39 N., R. 42 E., Hot Springs Peak quadrangle

[Analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036-C; P. L. D. Elmore, K. E. White, S. D. Botts, P. W. Scott, analysts]

Chemical analysis		C.I.P.W. norm		Niggli values	
		Q	0.5		
SiO_2	50.8	Or	0.3	al	21.9
Al_2O_3	14.2	Ab	38.0	fm	51.5
Fe_2O_3	2.4	An	18.4	c	15.0
FeO	11.0	Wo	1.0	alk	11.6
MgO	5.7	En	14.1	si	133
CaO	5.3	Fs	14.6	k	.986
Na_2O	4.5	Mt	3.5	mg	.431
K_2O	.06	Il	4.0	c/fm	.291
TiO_2	2.1	Ap	0.7	c/alk	1.29
P_2O_5	.29	Cc	1.3	qz	−13
MnO	.18	H_2O	3.0		
H_2O	3.0				
CO_2	.58		99.4		
Sum	100				
Sp gr (lump)	2.85				
(powder)	2.91				

Field relations indicate that the bodies of basalt are intrusive; they are essentially thick dikes that cut across the strike of the enclosing rocks. The body in this quadrangle intrudes the thrust plate of metamorphosed rocks of the Comus formation; the body in the adjacent quadrangle also intrudes these rocks and rocks of the Preble formation as well. The latter body apparently cuts the thrust fault between the Comus formation and the Preble formation; the body in section 20 of this quadrangle is terminated on the east by a high angle fault that is possibly of Tertiary age. The other body is cut by a dacite porphyry dike believed to be related to the Late Cretaceous episode of granitic intrusion. The available evidence suggests that these weakly metamorphosed basaltic bodies are hypabyssal intrusions of postthrusting pre-Late Cretaceous age and hence may be of Late Jurassic or Early Cretaceous age.

INTRUSIVE IGNEOUS ROCKS OF LATE CRETACEOUS AGE

GRANODIORITE AND RELATED ROCKS

DISTRIBUTION

The principal body of granodiorite is exposed on the eastern side of the Osgood Mountains in the northeastern part of the quadrangle (pl. 1). It extends from the ridge south of Granite Creek about to the Getchell mine; much of its eastern contact is near the range front, and its western contact is east of the crest of the range, except for a small area near the Burma Road where the contact extends several hundred yards west of the range crest. In general plan the body is a double stock consisting of two lobes joined by a narrow dikelike septum. The northern lobe, which is somewhat elongate, is about 3½ miles long and about 2 miles wide. The southern lobe has a nearly circular outline 2 miles in diameter. The total exposed area of the stock is 7.7 square miles. A few dikelike apophyses extend from the stock for short distances, and north of the stock there are a few small isolated bodies that may be connected with the stock at depth.

Some granodiorite is exposed east of the Getchell fault in the main gold pit (south pit), in the school yard north of the mill, and granodiorite was encountered in excavations for the mill foundations.

Three small bodies of granodiorite are exposed in the Hot Springs Range (pl. 1). The largest, which is in the little valley east of the Dutch Flat mine (NW¼ sec. 16, NE¼ sec. 17, T. 38 N., R. 40 E.), has a nearly circular outline and is 0.35 mile in diameter. To the north, two smaller bodies are found west of the range crest in NW¼ sec. 23 and SW¼ sec. 26, T. 39 N., R. 40 E. All three bodies are considerably altered and the one east of the Dutch Flat mine has metamorphosed the adjacent rocks for a distance of 200 to 1,000 feet from the contact.

STRUCTURAL RELATIONS OF THE STOCK IN THE OSGOOD MOUNTAINS

The Osgood Mountains granodiorite body has a well-developed joint pattern, which is illustrated by the point diagram, figure 3. Three major joint systems

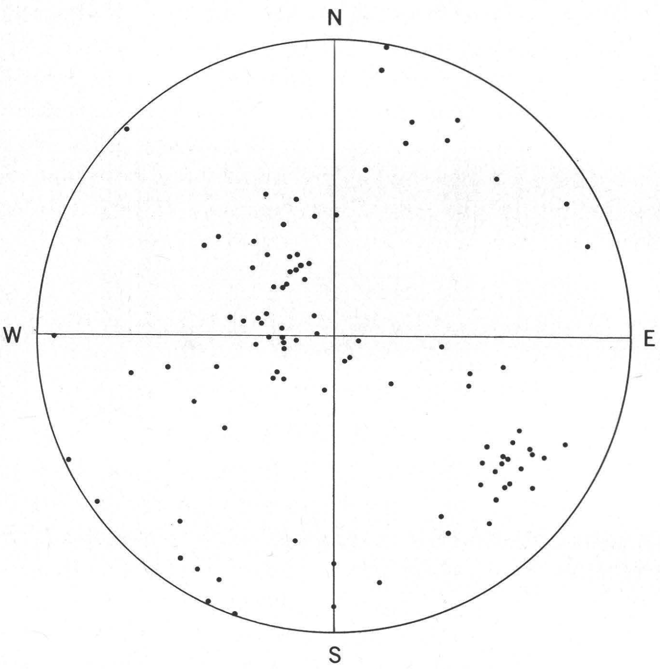

FIGURE 3.—Point diagram of poles of 95 joints measured in granodiorite stock of the Osgood Mountains. Plotted on upper hemisphere.

have been recognized: (1) one of the two most prominent systems strikes in a north to north-northeasterly direction and dips gently (average, 30°) west to northwest; (2) equally prominent is a northeast-striking system that dips steeply (average, 75°) southeast, nearly at right angles to the west-dipping joints; and (3) a less common system of west-northwest joints dips steeply north and south, or are vertical.

No planar and linear structural elements can be seen in the granitic rocks. Even adjacent to contacts with the country rocks, no preferred orientation is apparent in the granodiorite.

Both the eastern and western contacts of the stock dip eastward, but the eastern contact is less steeply inclined than the western, being parallel or nearly so to the dip of the sedimentary rocks, whereas the western contact is generally steeper than the bedding. The bedded rocks tend to wrap around the lobes of the stock and appear to have partly accommodated themselves to the boundaries of the granodiorite. In detail, however, the contact at many places sharply transects the bedding, and the north and south ends of the southern lobe and the south end of the northern lobe show markedly discordant relations.

The contact relations can be best observed in the mine workings and therefore more observations have been made along the eastern contact, where most of the tungsten deposits are located, than along the western boundary. Beginning at the north on the east side, at the Riley mine the general dip of the granodiorite contact is from 30° to 60° E., parallel to or slightly transecting the bedding in the limestone; the parallelism is not continuous, however, for the contact makes several sharp bends or jogs and cuts sharply across the bedding in some places (pl. 1). Farther south at the Kirby mine, which is on the south end of the northern lobe of the stock, the contact is irregular and in general cuts across the north-trending structure of the sedimentary rocks. The contact mostly dips steeply east or is vertical; but at the main deposit in the western part of the mine area, the contact dips about 30° E. beneath eastward-dipping tightly folded limestone beds. At the Pacific mine on the north-striking eastern boundary of the southern lobe, the granodiorite-limestone contact, like the bedding, dips steeply east (60°), with the usual local parallelisms and discordant relations. At the Granite Creek mine on the south end of the southern lobe of the stock the contact trends generally east and west, with many local irregularities and stubby projections, and transects the sedimentary rocks; the contact dips southeast rather steeply in the mine workings.

Direct observations cannot be made as easily on the west contact because of the absence of mine workings, except on the northwest side of the northern lobe. At the Richmond mine, the contact trends northeast across the general structure of the metamorphosed sedimentary rocks and is vertical. The contact at the Alpine mine in part cuts across the general sedimentary rock structure, but in the mine area it strikes parallel to the limestone and dips steeply eastward. Elsewhere on the western border the configuration of the contact and its relation to topography indicate that the contact dips steeply east.

The stock in the Osgood Mountains is a downward-enlarging asymmetric eastward-dipping body (pl. 2, section *B–B'*) as indicated from its contact relations. The steeply dipping, abruptly terminated southern contact and the absence of outlying bodies to the south suggest that the stock ends abruptly in this direction. The north end of the northern lobe extends out into the country rock in several dikelike apophyses, and north of the main mass some small plugs of granodiorite intrude the sedimentary rocks. In the low terrain north of Getchell the carbonate rocks of the Etchart limestone show widespread weak to moderate metamorphic effects. These data all suggest that the granodiorite has an underground extension to the north and may lie fairly close to the surface. The upward-converging contacts and the apparent gentle northward plunge of the crest of the stock suggest that the granodiorite body is the upper part of a larger subjacent mass. The rough parallelism between the east and west boundaries and the enclosing sedimentary rocks is an indication that the shape of the visible part of the body was partly controlled by the structure of the sedimentary rocks.

STRUCTURAL RELATIONS OF GRANDIORITE IN THE HOT SPRINGS RANGE

The largest granodiorite body in the Hot Springs Range east of the Dutch Flat mine has outward-dipping steep contacts, judging from the shape at the surface and its relation to topography, and from data obtained by mapping mines and prospects. Except very locally, the contacts cut across the structure of the sedimentary rocks. Extensive hydrothermal alteration of the granodiorite, a metamorphic aureole from 200 to 800 feet wide, and the many prospects and small mines within the body and around its periphery are an indication that the body is a cupola or a pluglike mass that may be of offshoot from a larger body at depth. The smaller bodies of identical composition in the northern part of the range may also be offshoots from the same body.

AGE OF THE GRANODIORITE

Relations to the country rocks indicate that the granodiorite is younger than Late Pennsylvanian or Early Permian and older than the Tertiary volcanic rocks. A lead-alpha age determination on zircon from the Osgood Mountains stock confirms this approximate dating and gives an age of 69 million years, corresponding to Late Cretaceous time (Jaffe and others, 1959, p. 82).[3]

PETROGRAPHY

The unaltered granodiorite has a rather uniform mineralogical composition. Some of the small pluglike bodies north of the stock in the Osgood Mountains differ slightly mineralogically, and in places along the contacts some different facies have been formed by reaction with wallrocks; but no important variations are found within the main bodies themselves. Except in a few places next to the contact, the granodiorite is free from inclusions.

The granodiorite is predominantly an equigranular medium-grained light-gray to very light gray rock composed mostly of visible intergrown white feldspar and clear inconspicuous quartz, with a uniform sprinkling of dark minerals which closer examination shows to be mostly biotite, less abundant hornblende, and a few metallic grains of magnetite. An occasional grain of honey-colored sphene can be seen in most specimens. Much of the biotite is in little columnar books with a hexagonal cross section commonly as much as 3 mm in diameter. The hornblende is in stubby subhedral prisms which are generally less than 1 mm long. In places the granodiorite is slightly porphyritic, where euhedral to subhedral crystals of plagioclase a few millimeters long occur in a contrastingly finer grained groundmass. At a few places the granodiorite is stained yellowish orange by iron oxides, but rock in these areas is no different in mineralogy from unstained rock; the staining possibly is related to an old surface of deep weathering, from which iron-bearing surface waters percolated downward into the underlying rock.

The microscope shows that the granodiorite is composed predominantly of plagioclase, orthoclase, and quartz, with biotite the most abundant mafic mineral and hornblende considerably less abundant but nevertheless consistently present. Magnetite, sphene, apatite, and, much more rarely, zircon are the remainder of the primary accessory minerals. Even the freshest rock contains a few secondary minerals, which include chlorite derived by alteration of biotite and, less commonly, hornblende, a little sericite and cloudy-looking clay minerals in some of the plagioclase, and a very few grains of epidote. Minute amounts of secondary sphene accompany some of the chloritized biotite.

The plagioclase is subhedral to euhedral, but its crystal faces are corroded where it is enclosed by orthoclase and quartz. The crystals show pronounced oscillatory zoning except for the rims, which are progressively zoned; the boundaries between zones are somewhat corroded. Except for the more sodic rims, the plagioclase composition ranges from An_{31} to An_{48}, and the average is approximately An_{40} (andesine); the rims range from An_{25} to An_{10} (oligoclase). Orthoclase is completely anhedral and interstitial, for it, along with quartz, was one of the last minerals to crystallize; commonly it encloses small corroded plagioclase crystals. It is almost entirely free from perthitic inclusions. A little myrmekite occurs along some boundaries between orthoclase and plagioclase.

Biotite, which is strongly pleochroic from reddish brown to golden yellow, commonly is partly altered to chlorite. Some of the mica plates poikilitically enclose small rounded grains of early-formed plagioclase. Many of the hornblende crystals are partly replaced by biotite and also show some alteration to chlorite. The hornblende is the common variety, pleochroic in yellowish green and greenish yellow.

Magnetite occurs in anhedral grains which tend to be distributed in groups, though it is also in isolated grains.

[3] Specimen 54-W-48, granodiorite, north side of Julian Creek, north center sec. 17, T. 38 N., R. 42 E., Osgood Mountains quadrangle, Nevada. α per milligram per hour, 378; lead (ppm), 10 and 11.

690-159 O—63——4

Apatite is in small euhedral crystals and tends to be more plentiful near magnetite. Sphene is in anhedral grains, though some typical diamond-shaped crystals are also seen. Zircon is extremely rare.

Modal analyses of 10 specimens of the granodiorite are in table 8. The mineral composition agrees with Johannsen's (1939, v. 1, p. 141–161; table 38) classification of granodiorite (227P). The positions of the specimens in the granodiorite field are shown in figure 4.

CHEMICAL COMPOSITION

The chemical compositions of four specimens of granodiorite are listed in table 9 with the computed norms and Niggli numbers. For comparison, Nockolds' (1954, p. 1014) average compositions of biotite granodiorite and granodiorite are included.

Spectrographic analyses for minor elements in the same samples are listed in table 10.

MARGINAL FACIES

A pyroxene-rich variant of this marginal facies occurs along the contact between gradiorite and marble, calc-silicate rock, or tactite. Both kinds may have been formed by endomorphic processes, but the evidence seems particularly good for the pyroxene-rich rock, which, therefore, will be described in a later section of the report on endomorphism of the granodiorite (p. 681).

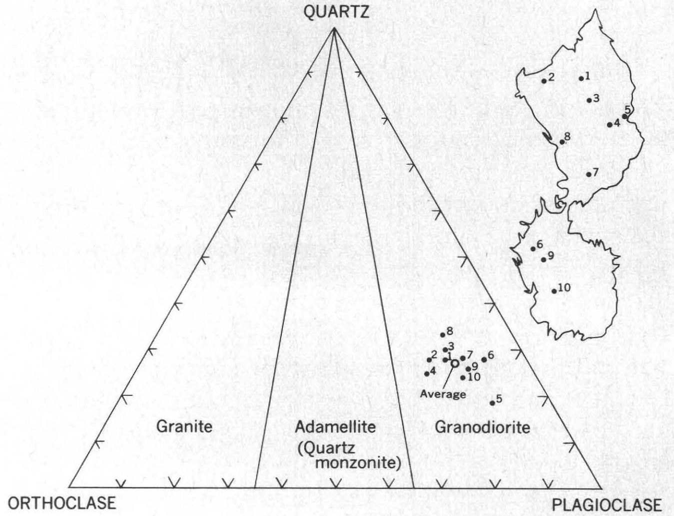

FIGURE 4.—Modal composition of granodiorite from the Osgood Mountains. Sketch shows specimen localities in stock.

TABLE 8.—*Modes of granodiorite from stock in the Osgood Mountains*

[In volume percent. Location of specimens shown on sketch map in fig. 4. Obtained from thin sections by the Chayes (1949) point-counter method and from stained slabs by the use of dot pattern (Jackson and Ross, 1956)]

Specimen	Location	Potassium feldspar (orthoclase)	Quartz	Plagioclase	Biotite	Hornblende	Accessories: Nonopaque	Accessories: Opaque
1 (A) [1]	Southeast side of Rocky Canyon, Osgood Mountains. NE¼NW¼ sec. 5, T. 38 N., R. 42 E. (75H53)	14	26	52	6	<0.5	1	1
2 (B) [1]	Point where Burma Road crosses crest of Osgood Mountains. Center NW¼NE¼ sec. 6, T. 38 N., R. 42 E. (53W81)	16	25	49	6	1	2	1
3 [2]	Near center, sec. 5, T. 38 N., R. 42E. (H-5-55)	12	27	49	13	Tr		
4 [2]	N. part NE¼ sec. 8, T. 39 N., R. 42 E. (H-2-55)	18	22	49	4	7		
5 [2]	West of Riley mine, c. a. 50 ft. from contact SW¼SW¼ sec. 4, T. 38 N., R. 42 E. (H-1-55)	9	16	59	15	<.5		
6	South side of ridge between Granite and Osgood Creeks, Osgood Mountains. Northeast of center, sec. 19, T. 38 N., R. 42 E. (54W12)	7	25	56	8	3	<.5	1
7	North side Julian Creek. N. center sec. 17, T. 38 N., R. 42 E. (54W48)	11	24	51	12	5	1	.5
8	South side of ridge between Julian and Summer Camp Creeks, Osgood Mountains. Near center, SE¼NE¼ sec. 7, T. 38 N., R. 42 E. (54W16)	12	29	48	8	1	1	.5
9	North side of ridge between Granite and Osgood Creeks, Osgood Mountains. Near NW¼SE¼NW¼ sec. 19, T. 38 N., R. 42 E. (54W13)	11	23	55	7	2	1	1
10 [2]	Granite Creek, N. Part NE¼ sec. 30, T. 38 N., R. 42 E. (H-34-55)	12	21	54	12	1		
Average		12	24	52	9	1	1	1

[1] Letters refer to chemical analyses, table 9.
[2] From stained slabs.

TABLE 9.—*Analyses, norms, and Niggli values of granodiorite*

[Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C; P. L. D. Elmore, K. E. White, S. D. Botts, P. W. Scott, analysts, U.S. Geol. Survey]

	A	B	C	D	E	F
Chemical analyses						
SiO_2	67.6	68.4	67.2	66.4	68.97	66.88
Al_2O_3	16.8	16.8	16.0	16.6	15.47	15.66
Fe_2O_3	1.3	1.6	.9	.9	1.12	1.33
FeO	1.6	1.4	1.7	1.4	2.05	2.59
MnO	0.10	.09	.9	.5	.06	.07
MgO	1.0	1.2	.92	.94	1.15	1.57
CaO	4.0	4.0	3.3	3.0	2.99	3.56
Na_2O	3.5	3.5	3.6	3.9	3.69	3.84
K_2O	2.8	2.8	3.2	3.2	3.16	3.07
TiO_2	.42	.40	.38	.43	.45	.57
P_2O_5	.22	.22	.20	.28	.19	.21
H_2O+	--------	--------	--------	1.3	.70	.65
H_2O-	--------	--------	--------	.36	--------	--------
Ig [1]	.52	.34	1.8	--------	--------	--------
CO_2	--------	--------	--------	.69	--------	--------
Sum	100	101	99	100	100	100
Total S as S	.04	.05	.14	.51	--------	--------
SO_4	--------	--------	.05	.2	--------	--------
CO_2	.05	.05	.90	--------	--------	--------
Sp gr (lump)	--------	--------	--------	2.62	--------	--------
(powder)	2.69	2.68	2.68	2.67	--------	--------

	A	B	C	D	E	F
C.I.P.W. norms						
Q	26.3	27.2	24.4	24.0	26.2	21.9
Or	16.7	16.7	18.9	18.9	18.9	18.3
Ab	29.3	29.3	30.4	33.0	31.4	32.5
An	18.9	18.9	15.6	13.1	14.2	16.4
En	2.5	3.0	2.3	2.3	2.9	3.9
Fs	1.4	.66	3.4	.9	2.2	2.9
C	1.1	1.1	.9	1.9	.7	--------
Mt	1.86	2.3	1.4	1.4	1.6	1.9
Il	.76	.8	.8	.8	.8	1.1
Ap	.34	.3	.3	.67	.4	.5
Pr	--------	--------	--------	2.3	--------	--------
Sum	99	100	98	99	99	99
Niggli values						
al	42.8	42	42.7	45.0	41.1	32.7
fm	16.6	17.9	16.3	16.3	19.5	23.2
c	18.4	18.1	16.0	14.4	14.3	15.8
alk	22.2	22	25.0	25.9	25.1	23.4
si	292	291	304	295	321	275
k	.349	.349	.370	.351	.37	.348
mg	.391	.428	.384	.377	.40	.391
qz	103	103	140	91	121	81

[1] Includes gain due to oxidation of FeO.

A. East side Osgood Mountains, SE side Rocky Canyon, NE¼NW¼ sec. 5, T. 38 N., R. 42 E. (75H53).
B. Osgood Mountains, where Burma Road crosses range. Center NW¼NE¼ sec. 6, T. 38 N., R. 42 E. (53W81).
C. Hot Springs Range, Dutch Flat district. Extreme west center NW¼NW¼ sec. 16, T. 38 N., R. 40 E. (53W94).
D. Hot Springs Range, Dutch Flat district. Bottom of gully, SW¼NW¼NW¼ sec. 16, T. 38 N., R. 40 E. (20H53).
E. Nockolds' average biotite granodiorite (Nockolds, 1954, p. 1014).
F. Nockolds' average granodiorite (Nockolds, 1954, p. 1014).

TABLE 10.—*Quantitative spectrographic analyses for minor elements in granodiorite*

[Harry Bastron, analyst, U.S. Geol. Survey]

	A	B	C
Cu	0.0003	0.002	0.002
Mn	.03	.03	.02
Ga	.001	.001	.001
Cr	.0006	.0008	.0006
V	.007	.008	.008
Y	.002	0	0
Yb	.0001	.0001	0
Ti	.2	.2	.2
Zr	.007	.007	.007
Sr	.08	.08	.09
Ba	.1	.2	.2
B	.001	.002	0

Looked for but not found: Ag, Mo, Co, Ni, P, Au, Hg, Ru, Rh, Pd, Ir, Pt, W, Re, Ge, Sm, Pb, As, Sb, Bi, Te, Zn, Cd, Tl, In, Sc, Th, Nb, Ta, U, Be, Li, Cs.
The above results have an overall accuracy of ±15 percent of the value shown.
A. East side Osgood Mountains, SE side Rocky Canyon, NE¼NW¼ sec. 5, T. 38 N., R. 42 E. (75H53).
B. Hot Springs Range, Dutch Flat district. Extreme west center NW¼NW¼ sec. 16, T. 38 N., R. 40 E. (53W94).
C. Hot Springs Range, Dutch Flat district. Bottom of gully, SW¼NW¼NW¼ sec. 16, T. 38 N., R. 40 E. (20H53).

At some places along the margin of the granodiorite stock in the Osgood Mountains a somewhat more mafic facies generally characterized by abundant hornblende has been recognized. The marginal zone is narrow; its width is seldom as much as 10 feet, more often 3 to 5 feet, and locally only a few inches.

The marginal facies is characterized by a slightly higher content of mafic minerals than in the normal granodiorite, less quartz, and less orthoclase, although the amount of potassium feldspar varies between rather wide limits. The grain size is about the same as in the granodiorite, and microscopic examination shows that texturally the rocks are the same. As in the granodiorite, the plagioclase is subhedral to euhedral and shows strong rhythmic zoning; the composition is from calcic andesine to oligoclase. Hornblende instead of biotite is the most abundant mafic mineral; in some specimens there is no biotite. Very locally, as in pit 2 of the Kirby mine, biotite is the chief mafic mineral of the border facies. The hornblende is the common green variety and apparently the same as that which occurs in lesser amount in the granodiorite away from the border facies. The hornblende is in subhedral to anhedral prisms. Actinolitic hornblende (Z, very pale green; X, colorless; $Z \wedge c$ 15°) is seen in some specimens where apparently it is an alteration product of the hornblende. Most biotite occurs in euhedral to subhedral crystals, but some partly replaces hornblende. Sphene is considerably more abundant in some of the marginal facies than in the normal granodiorite. Some of it is in small grains and trains of granules within chloritized biotite, but most of it is in relatively large irregular-shaped grains and euhedral crystals. Commonly it is a rather strongly colored variety and contains little areas within the larger grains that are markedly pleochroic from pink to colorless.

Mild alteration effects are common in rocks from the marginal facies. Biotite and hornblende usually show some alteration to chlorite; plagioclase in some specimens is rather severely altered to sericite and subtranslucent gray finely divided clay minerals; in others the majority of grains are essentially fresh. Some calcite as veinlets and small interstitial masses is fairly common, and a grain or two of epidote can be seen in some thin sections.

The modal composition of specimens of the marginal facies is given in table 11.

TABLE 11.—*Modes of marginal facies of granodiorite from stock in the Osgood Mountains*

[In volume percent]

Specimen	Potassium feldspar	Quartz	Plagioclase	Biotite	Hornblende	Accessories	
						Nonopaque	Opaque
1	22	13	48	1	12	4	tr
2	20	16	48	------	12	4	tr
3	5	13	66	5	8	2	1
4	12	15	59	4	11	------	------
5	10	18	60	6	5	1	------

1. Approximately 1 ft from contact with hornfels, west of Top Row Pit mine, SE¼ NW¼SW¼ sec. 9, T. 38 N., R. 42 E. (54W18).
2. Zone 2–3 ft wide between granodiorite and pyroxene-bearing granite rock adjacent to contact. Pacific mine. (H-50-54).
3. Darker granitic rock near tactite. Getchell extension of Riley mine (H-58-55).
4. Approximately 3 ft from contact with tactite. Kirby mine, north side Julian Creek (54-W-26).
5. Approximately 10 ft from tactite contact. Tonopah mine (To-8-54).

ALTERATION OF THE GRANODIORITE

Most of the granodiorite of the Osgood Mountain stock and its outliers in the north end of the range is essentially free from alteration, except for minor effects such as partial chloritization of biotite and hornblende and slight sericitization of plagioclase. Approximately in the center of the northern lobe (SW¼ sec. 5, T. 38 N., R. 42 E.), however, the granodiorite over an area of approximately one-fourth square mile has been modified by late magmatic solutions prior to final crystallization and altered by hydrothermal solutions after its consolidation. The small granodiorite stock at Dutch Flat and two smaller bodies farther north in the Hot Springs Range have also been rather extensively altered.

The granodiorite in the Osgood Mountains is altered adjacent to some of the tungsten deposits and where it has been affected by solutions related to mineralization along the Getchell fault; these effects are described elsewhere in the report. The discussion here is limited to a consideration of the "internal" alteration in the Osgood Mountains body and at Dutch Flat.

In the areas of alteration the granodiorite is stained reddish brown with iron oxide derived by weathering of pyrite. Much of the rock is finer grained than unaltered granodiorite and in places, particularly in the stock at Dutch Flat, has a porphyritic texture with feldspar and quartz in a finer groundmass of granular quartz and feldspar. In much of the rock, feldspar is dull gray and altered appearing, and white mica instead of biotite can be observed. Where alteration has been more severe, commonly along the periphery of the bodies, the granodiorite has been transformed to a soft, somewhat porous light-gray quartz-sericite rock with spots of brown iron oxide.

In the areas of alteration the granodiorite has been affected by both late magmatic and postmagmatic changes. The late magmatic episode resulted in a peculiar texture that is responsible for the porphyritic appearance. At Dutch Flat most of the granodiorite, instead of having the usual granitic texture, is porphyritic with relatively large euhedral crystals of plagioclase, some anhedral quartz, and biotite books enclosed in a fine-grained equigranular groundmass of quartz and orthoclase. The plagioclase phenocrysts are generally more or less cloudy with kaolinitic material and are partly sericitized, though parts may be quite fresh and in some specimens the plagioclase is unaltered or only slightly so. In the Osgood Mountains, little of the granodiorite in the zone of alteration is as clearly porphyritic as at Dutch Flat, but earlier crystallized plagioclase is partly replaced by orthoclase which, with small rounded bodies of quartz, forms a slightly finer grained interstitial groundmass.

The porphyritic texture was formed partly by replacement, which is indicated by corrosion of the plagioclase phenocrysts by the groundmass, but there is no evidence that the groundmass has formed by wholesale replacement of previously crystalline material. The plagioclase phenocrysts are all of a similar size and are like the plagioclase crystals in the normal hypidiomorphic granodiorite. Similar textures have been described at Ajo, Ariz., by Gilluly (1946, p. 29) who wrote that "* * * if the groundmass were entirely of replacement origin, the preservation of the major form of the plagioclase crystals would be inexplicable." There is, furthermore, no indication that quartz and orthoclase have been added in excess of the normal amount in granodiorite. An alternative explanation suggested by Gilluly (1946, p. 29) is that equilibrium conditions suddenly changed prior to complete consolidation, and the earlier crystals were attacked by the residual liquid, but the replacement was quickly halted by the final crystallization.

Postmagmatic changes in the granodiorite involved hydrothermal alteration effects similar to some of those that have been described by many geologists who have studied altered granitic rocks (Schwartz, 1939; 1947). In the rocks that have been hydrothermally altered, plagioclase has been albitized and strongly sericitized. The prominent zoning so characteristic of plagioclase in the unaltered granodiorite becomes faint or disappears, but in general the twinning is still visible; the composition becomes more sodic and approaches pure albite; the crystals are filled with sericite and minute blebs of kaolinite. Biotite is completely replaced by chlorite that contains blebs and chains of leucoxene and laminae of colorless to slightly yellowish strongly birefrigent mica. Orthoclase remains relatively fresh, though it is slightly clouded with submicroscopic inclusions. Where alteration has been more severe, the feldspar, including orthoclase, is completely replaced by sericite; and instead of chlorite, colorless mica [4] with interlaminated rutile and leucoxene marks the former place of biotite and also occurs as patches and sheaves in the groundmass, where in part it appears to fill small fractures and cavities. At a few places, commonly on the periphery of the alteration zones, the altered rock is heavily impregnated with late quartz that apparently has replaced the altered feldspar, and sericite [5] occurs as larger flakes and sheaves intergrown with the quartz. In the Osgood Moutains, near the southern boundary of the alteration zone but within the altered granodiorite, irregularly shaped masses of coarsely crystalline gray to white quartz cut the altered rock. Some of the quartz is in terminated crystals as much as 6 inches long and 4 inches in diameter, which include large crystals of pyrite an inch or more across. In one place—the Section 5 pit—the quartz contains some scheelite. In and adjacent to the Dutch Flat stock, gold-bearing quartz veins cut the altered granodiorite and country rock.

The altered granodiorite contains a few percent of pyrite as small scattered grains. Its paragenetic relations are uncertain, but it seems to have been introduced either during the late magmatic stage or very early in the episode of postmagmatic alteration, for it is present in rock that has been only slightly sericitized. Some pyrite also accompanied the late development of coarsely crystalline quartz in the Osgood Mountains alteration zone.

CHEMICAL CHANGES

Samples of altered granodiorite from Dutch Flat were chemically analyzed and are compared with unaltered rock in table 12. Gains and losses of chemical constituents in milligrams per cubic centimeter for the samples are graphically illustrated in figure 5.

[4] Examination under the microscope shows that this mica has a birefringence (estimated 0.038) and index of refraction (b=1.595±0.003) like muscovite, but a smaller optic angle (5°–10° in many). X-ray powder diagrams have characteristics of muscovite.

[5] Optical properties like those of muscovite (estimated birefringence 0.038; index of refraction, b=1.595±0.003; 2V approximately 45°); X-ray powder diagrams also have lines characteristic for muscovite.

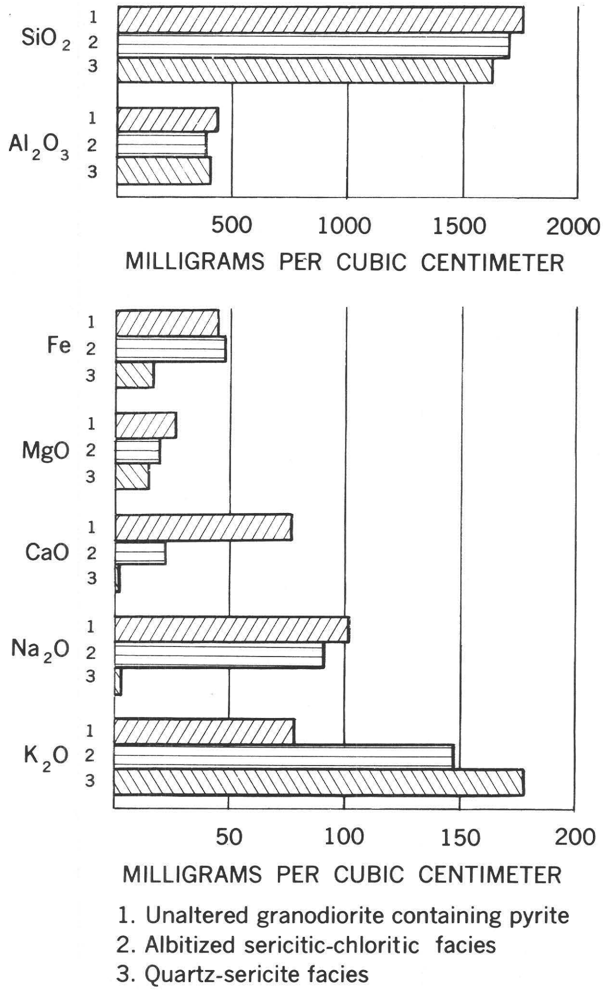

FIGURE 5.—Graph showing gain and loss of principal rock constituents by alteration of the granodiorite stock at Dutch Flat. 1, Unaltered granodiorite containing pyrite. 2, Albitized-sericitic-chloritic facies. 3, Quartz-sericite facies.

SiO_2, Al_2O_3, and Na_2O remain fairly constant in the formation of the albite-sericite-chlorite facies but there is a significant loss of CaO and an important gain in K_2O. Loss of CaO and constancy of Na_2O indicate that albitization of the plagioclase is due to the breakdown of the more calcic plagioclase and removal of CaO rather than an introduction of Na_2O. The increase in K_2O is reflected in the formation of sericite, the sericite taking up Al_2O_3 and SiO_2 released by alteration of the plagioclase while CaO set free by this replacement was removed. Total Fe and MgO do not change appreciably in alteration of granodiorite to the albite-sericite-chlorite facies, but the analyses show a marked increase in Fe_2O_3 and a concomitant decrease in FeO, reflected mineralogically by the presence of chlorite in place of biotite and by iron oxide pseudomorphs of pyrite, which probably was more abundant

TABLE 12.—*Analyses, norms, and Niggli values for unaltered and altered granodiorite*

[Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C; P. L. D. Elmore, H. F. Phillips, K. E. White, analysts, U.S. Geol. Survey]

	1	2	3
Chemical analyses			
SiO_2	66.4	68.5	70.5
Al_2O_3	16.6	16.5	18.6
Fe_2O_3	.9	2.0	.85
FeO	1.4	.75	.12
MnO	.5	.08	.02
MgO	.94	.75	.58
CaO	3.0	.90	.15
Na_2O	3.9	3.7	.16
K_2O	3.2	4.0	5.6
TiO_2	.43	.54	.48
P_2O_5	.28	.33	.08
H_2O+	1.3	1.4	2.4
H_2O-	.36	.06	.06
CO_2	.69	.05	.05
Sum	100	100	100
Total Fe	1.72	1.98	0.69
Total S as S	----	.02	.04
S	.51	----	----
SO_4	.2	----	----
Sp gr (lump)	2.62	2.47	2.29
(powder)	2.67	2.69	2.73

	1	2	3
C.I.P.W. norms			
Q	24.0	29.2	47.1
Or	18.9	23.91	33.36
Ab	33.0	31.44	1.36
An	13.1	2.50	----
C	1.9	5.1	12.1
En	2.3	1.9	1.4
Fs	.9	----	----
Mt	1.4	.93	----
Il	.8	1.06	.30
Pr	2.3	----	----
Ap	.67	.67	.34
Hm	----	1.44	.80
Ru	----	----	.03
H_2O	1.3	1.4	2.4
Sum	100.57	99.55	99.19
Niggli values			
al	45.0	50.0	67.6
fm	16.3	13.3	7.8
c	14.4	4.9	1.1
alk	25.9	31.8	23.4
si	295	346	437
k	.351	.416	.954
mg	.377	.422	.667
qz	171	118	243

1. Hot Springs Range, Dutch Flat. Bottom of gully, SW¼NW¼NW¼ sec. 16, T. 38 N., R. 40 E. Unaltered granodiorite containing pyrite.
2. Hot Springs Range, Dutch Flat. Prospect in NW¼NW¼ sec. 16, T. 38 N., R. 40 E. Albite-sericite-chlorite facies.
3. Hot Springs Range, Dutch Flat. Adit, NW¼NW¼ sec. 16, T. 38 N., R. 40 E. Quartz-sericite facies.

in the sample of altered rock selected for analysis than it was in the unaltered sample. In the most intensely altered granodiorite, the quartz-sericite facies, there is a notable loss of Fe, MgO, CaO, and Na_2O, but a further increase of K_2O. SiO_2 decreases slightly, but Al_2O_3 remains constant.

Throughout the alteration, titania shows no appreciable change, reflected mineralogically by the formation of stable leucoxene and rutile from the breakdown of sphene and biotite in the original rock. P_2O_5 is constant in the albite-sericite-chlorite facies, but is markedly leached away in the more severely altered quartz-sericite facies.

COMPARISON WITH OTHER EXAMPLES OF HYDROTHERMAL ALTERATION

The alteration of granodiorite at Dutch Flat and in the Osgood Mountains is remarkably similar to the hydrothermal alteration of quartz monzonite at the principal porphyry copper deposits in the southwestern United States, which has been summarized and compared graphically by Anderson (1950, p. 613–615, 617–625), Anderson, Scholz, and Strobell (1955, p. 52–59), and Creasey (1959). Marked addition of K_2O accompanied by loss of CaO and Na_2O, with SiO_2 and Al_2O_3 remaining constant, are features common to this kind of alteration. MgO is generally leached, but in some examples it increases. Fe generally does not increase, and commonly is leached in the more altered facies. The main difference in the alteration in the Osgood Mountains quadrangle is that it was not accompanied by the introduction of chalcopyrite, whereas in the porphyry copper deposits introduction of chalcopyrite accompanied the stage of sericitic alteration (Schwartz, 1947, p. 351; 1955, p. 302). Spectrographic analysis (table 13) shows 0.002 percent copper in the unaltered granodiorite at Dutch Flat, and 0.00003 percent in a sample from the northern lobe of the stock in the Osgood Mountains. The albite-sericite facies at Dutch Flat contains 0.005 percent, 2.5 times as much as the unaltered rock, and the quartz-sericite facies contains 0.002 percent copper. In contrast, Anderson (1950, p. 618; Anderson, Scholz, and Stobell, 1955, p. 55) reported 1.0± percent copper and 2.4 percent chalcopyrite in the albitic facies of quartz monzonite at Bagdad. No chalcopyrite has been observed in the granodiorite at Dutch Flat or in the altered zone at the Osgood Mountains, and there are no indications that it was present and subsequently removed by supergene leaching. Sulfide metallization seems to have been restricted to the deposition of iron sulfide at an early stage in the alteration, possibly during or shortly following the late magmatic episode. Chalcopyrite is known only in some of the contact metamorphic scheelite deposits, where it may be later than the episode of alteration in the granodiorite. Possibly it is significant that the alteration zones in the granodiorite here are not zones of extensive fracturing. McKinstry (1955, p. 192–193) has emphasized that "porphyry copper" deposits "in all cases * * * are in thoroughly fractured and shattered rock * * *."

TABLE 13.—*Quantitative spectrographic analyses for minor elements in unaltered and altered granodiorite*

[Harry Bastron, analyst]

	1	2	3		1	2	3
Cu	0. 002	0. 005	0. 002	Y	------	0. 002	------
Ag	------	. 0005	------	Yb	------	. 0001	------
Mo	------	. 0005	------	La	------	. 006	------
Mn	. 02	. 02	. 001	Ti	0. 2	. 2	0. 2
Ni	------	. 0003	------	Zr	. 02	. 01	. 01
Ga	. 001	. 001	. 001	Sr	. 09	. 05	------
Cr	. 0006	. 0008	. 0007	Ba	. 2	. 2	. 1
V	. 008	. 008	. 008	P	------	. 1	------
B	------	. 006	. 04				

Looked for but not found: Au, Hg, Rn, Rh, Pb, Ir, Pt, W, Re, Ge, Co, Sn, Pb, As, Sb, Bi, Te, Zn, Cd, Tl, In, Sc, Th, Nb, Ta, Nb, Ta, U, Be, Li, Cs.
The above results have an overall accuracy of ±15 percent of value shown.
1. Unaltered granodiorite containing pyrite.
2. Albite-sericite-chlorite facies.
3. Quartz-sericite facies.

SIMILAR ALTERATION OF SEDIMENTARY ROCKS OF THE HARMONY FORMATION

Sandstone and shale of the Harmony formation have been altered locally for as much as half a mile from the granodiorite stock at Dutch Flat, at the Last Chance prospect in south-central sec. 33, T. 38 N., R. 40 E., and in the vicinity of the K and K quicksilver prospect in the central part sec. 5, T. 37 N., R. 40 E., approximately 4 and 5 miles, respectively, from the stock. The alteration is very similar to the hydrothermal alteration that affected the granodiorite. In the zones of alteration, sandstone and shale have been more or less softened and converted to light-gray or white rocks composed principally of quartz, sericite, and kaolinite. Presumably the sites of alteration were localized by fractures that made the rocks accessible to fluids that caused the alteration, for the rocks are not affected everywhere adjacent to the granodiorite stock, but wherever quartz veins cut them alteration effects are apparent. Where the rock is altered and there are no quartz veins, there is abundant evidence of fracturing, and rocks at the Last Chance and K and K quicksilver prospects are altered where there has been fracturing and cinnabar mineralization.

Where the rocks are altered they are noticeably softer than the unaltered shale and sandstone, but their sedimentary structures are preserved except where the rocks are crushed in fault zones. As a result of the alteration the rocks are bleached light gray to white, but in many places downward-percolating solutions have precipitated iron oxides which stain them shades of brown and yellow.

Where feldspathic sandstone so typical of the Harmony formation has been strongly altered, the product is a rock composed of original quartz grains in a matrix of sericite, kaolinite,[6] and silt-sized and smaller quartz grains. Former feldspar grains are barely discernible under the microscope, for they are composed of sericite and kaolinite like the matrix; plagioclase and potassium feldspar are equally affected by the alteration process. Siltstone and shale also have been converted to fine-grained quartz-sericite-kaolinite rocks.

Where the rocks are less severely altered, as indicated megascopically by their medium-gray color in contrast to the white or very light gray of the strongly altered rocks, the feldspar grains are cloudy with kaolinite and contain small amounts of sericite, the plagioclase more so than the potassium feldspar. The matrix between the grains in sandstone and the fine material in siltstone and shale is not much different than in unaltered rocks; some material of low birefringence that may be kaolinite can be seen, but biotite, where present, is unchanged except for slight bleaching.

MINOR INTRUSIVE BODIES

QUARTZ DIORITE

Some small intrusive bodies mapped as altered granodiorite at the northern end of the Osgood Mountains in sec. 24, T. 39 N., R. 41 E., and secs. 19 and 30, T. 39 N., R. 42 E., proved to be altered quartz diorite when petrographic studies were made of them in the laboratory. They are, however, probably related closely to the granodiorite, which they resemble except for the absence of potassium feldspar. Their alteration is like the alteration of the granodiorite in the northern lobe of the main stock and at Dutch Flat. Plagioclase is albitized and largely replaced by sericite, original biotite is replaced by chlorite, and some pyrite has been introduced, though in most specimens it has been replaced by iron oxide due to weathering. Unlike the altered

[6] Confirmed by X-ray powder patterns of −2 micron fraction.

granodiorite, this rock contains a few percent of calcite, which is interstitial and seems at least partly associated with the sericite. The quartz diorite is finer grained than the granodiorite, and some is slightly porphyritic. The bodies are small, some of them are dikes, and some are irregular in outline and have dikelike apophyses. The mode of occurrence as well as the generally finer grain size suggests that these small bodies may be hypabyssal intrusives.

APLITE AND PEGMATITE

Dikes of aplite cut the granodiorite, but are not abundant. They range in width from an inch or so to as much as 10 feet, but are more commonly on the order of 3 to 5 feet in width. These dikes usually are several tens of feet long, and some are more than 100 feet in length. The aplites are white even-textured medium-grained rocks composed principally of quartz and feldspar with virtually no accessory mafic minerals. Microscopic study of several typical specimens shows that although some of the plagioclase crystals are subhedral, the texture is allotriomorphic, an intergrowth of anhedral plagioclase, orthoclase, and quartz. Orthoclase makes up 50 to 60 percent of the rock; quartz, 25 to 30 percent; and plagioclase (sodic andesine to sodic oligoclase), 15 to 25 percent. The remaining few percent is biotite, chlorite, and sphene, with traces of zircon.

Pegmatite dikes and veinlets composed of coarsely crystalline intergrowths of quartz, feldspar, and a little mica are much less common than aplite.

DACITE PORPHYRY

Thin tabular bodies of intrusive dacite porphyry are widespread in the Paleozoic sedimentary rocks of the Osgood Mountains and also intrude the granodiorite. They range from approximately 5 feet to 20 feet in thickness, but most of them are about 10 feet thick. Although relatively thin, the bodies are very persistent laterally, being continuous for many hundreds or even thousands of feet; one body, approximately 1 mile south of Adam Peak, is continuous for more than 2 miles (pl. 1). Some of the bodies are dikelike in the sense that they cut across the preexisting structure, but many are nearly flat to gently dipping sheets which have been intruded along the contact between Osgood Mountain quartzite and the Etchart limestone, or they lie within the Etchart a short distance above the contact. Similarly, a sheet of dacite porphyry has been intruded partly along the low-angle thrust fault between the Preble formation and the plate of metamorphosed rocks of the Comus formation at the north end of the Osgood Mountains.

Typically, the intrusive dacite porphyry is an altered-appearing medium light greenish-gray porphyritic rock with dull white euhedral phenocrysts of felspar 2 mm to 6 mm in greatest dimension in a fine-grained dull groundmass containing chloritic remnants of original mafic minerals. On microscopic examination it is seen that the original plagioclase phenocrysts of most specimens are completely altered to sericite and sodic plagioclase containing unidentified clay minerals, carbonate, and some epidote, although a few contain unaltered cores of andesine to sodic labradorite that show twinning and strong oscillatory zoning. Other original hornblende and biotite phenocrysts, smaller than the plagioclase, are now chlorite pseudomorphs. A few small phenocrysts of quartz may be present. The groundmass is holocrystalline and composed of cloudy feldspar most of which apparently is plagioclase near sodic oligoclase or albite in composition, some orthoclase, subordinate interstitial quartz, and an abundance of sericite, chlorite commonly containing secondary sphene, carbonate, some epidote, apatite, and very little magnetite. Some specimens contain small miarolitic cavities 1 or 2 mm in diameter filled with quartz and less commonly, calcite.

A sample of typical dacite porphyry from the long sheet that crops out on the northeast side of Goughs Canyon was analyzed (table 14). Its composition is very similar to the granodiorite (table 9) of the main stock except that it is slightly lower in SiO_2 and contains slightly more FeO. The relatively high H_2O+ and CO_2 content reflect the altered condition of the rock. The analysis strengthens the conclusion arrived at from the field relations and the petrography that these widespread thin tabular intrusive bodies are probably genetically related to the granodiorite intrusives.

TABLE 14.—*Analysis, norm, and Niggli values of typical altered intrusive dacite porphyry, Goughs Canyon: SE¼SE¼ sec. 22, T. 38 N., R. 41 E.*

[Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C; P. L. D. Elmore, K. E. White, analysts, U.S. Geol. Survey]

Chemical analysis		C.I.P.W. norm		Niggli values	
SiO_2	63.3	Q	24.4	al	39.3
Al_2O_3	16.5	C	4.9	fm	22.8
Fe_2O_3	1.6	Or	17.1	c	16.0
FeO	2.2	Ab	31.3	alk	22.9
MgO	1.7	An	6.6	si	256
CaO	3.7	En	4.2	k	.340
Na_2O	3.7	Fs	1.9	mg	.449
K_2O	2.9	Mt	2.3	qz	+68
TiO_2	.55	Il	1.0		
P_2O_5	.26	Ap	.6		
MnO	.08	Cc	3.6		
H_2O+	2.0	H_2O	2.0		
CO_2	1.6				
			99.9		
	100				

ANDESITE PORPHYRY

Some thin dikes of andesite porphyry were observed in the northern part of the Osgood Mountains but are not shown on the geologic map. They are much less plentiful than the dacite sheets. One of the most prominent bodies cuts granodiorite at the Riley mine and is exposed underground and at the surface on the slope west of the open cuts. Some dikes, which have been altered, are known to cut sedimentary rocks at the Getchell gold mine.

The unaltered rock is dark gray to medium gray, fine grained, holocrystalline, and porphyritic, with prominent white phenocrysts of plagioclase, smaller randomly oriented prismatic crystals of hornblende, and, in some specimens, small biotite phenocrysts. Even in the freshest rock the plagioclase phenocrysts are mostly altered to sericite, but clear remnants show strong normal progressive zoning with cores of labradorite. The groundmass is fine grained and composed chiefly of plagioclase, hornblende, and biotite. The plagioclase is fresh and euhedral, and the squarish crystals have strong progressive zoning from cores of labradorite (An_{60}) to andesine (An_{43}) in the outer parts of the crystals, and very narrow rims of oligoclase (An_{20}). Hornblende and biotite in the groundmass as well as the phenocrysts are partly chloritized. The specimens that were examined microscopically contain 1.5 to 2 percent interstitial quartz in the groundmass.

ALTERATION OF THE MINOR INTRUSIVES

A characteristic feature of the hypabyssal intrusives, especially the dacite porphyry, is their widespread alteration. For the most part the alteration has been a mild saussuritization, accompanied by the chloritization of hornblende and biotite. These alteration effects cannot be attributed to the effect of heated solutions given off by the granodiorite, which is only locally altered. Furthermore, many of the intrusives are in unaltered sedimentary country rocks, whereas others cut rocks that have been metamorphosed by the granodiorite intrusive body. The alteration seems to have been the result of reactions peculiar to the hypabyssal intrusives themselves—in other words, an endomorphic process.

In a study of alteration of sills on Loveland Mountain, Colorado, Singewald (1932, p. 27) concluded that because the sedimentary rocks show no sign of alteration comparable to that in the intrusive rocks, the agents of alteration were probably contained in each intrusive magma or followed the same paths, and the alteration was an end phase of the igneous intrusion. There is no indication of fracturing that would have provided channels for the altering solutions other than those taken by the magma. The presence of miarolitic cavities in some specimens of dacite porphyry is an indication that after emplacement and prior to final crystallization the magma or the crystal mush contained hydrothermal solutions that presumably would have been capable of reacting with and altering earlier crystallized constituents. Perhaps, as Singewald (1932, p. 27) postulated, late-stage solutions within the sills themselves were augmented by hyperfusibles from more deeply seated magma before complete consolidation in the conduits themselves. In the Ajo district, Arizona, Gilluly (1946, p. 83) discussed the possibility that similar alteration in andesite dikes may have resulted from the action of water in the sediments that was heated by the intrusives and actively penetrated them; however, such an explanation would not hold for dikes that cut the granodiorite, as at the Kirby mine.

AGE OF THE MINOR INTRUSIVES

Relations between the dacite porphyry and the andesite porphyry are not known, for neither has been seen cutting the other. The composition of the dacite porphyry suggests strongly that it is the hypabyssal intrusive equivalent of the granodiorite. An andesite dike cuts granodiorite at the Riley mine; therefore, the andesite is later than the crystallization of the granodiorite. Hobbs (1948,[7] p. 43) regarded the intrusive andesite as "* * * related to a much later phase of the same igneous activity * * *" that was responsible for the emplacement of the granodiorite, and Joralemon (1949,[8] p. 25; 1951, p. 269) considered the andesite to be an early phase of volcanism of Tertiary age. The mineralogy of the unaltered andesite is quite different, however, from the andesite flows of Tertiary age elsewhere in the quadrangle, which contain olivine and pyroxene; the intrusive andesite contains hornblende biotite and zoned plagioclase, which suggests its consanguinity with the granodiorite.

ROCKS OF TERTIARY AGE

Remnants of a formerly more extensive cover of volcanic rocks are scattered over the quadrangle in the Hot Springs Range and in the lower parts of the Osgood Mountains (pl. 1). These are mostly andesitic flows, but in a few places the flows are underlain by stratified water-laid tuffs and rhyolitic welded tuffs. Some remnants of conglomerates that are presumably of Tertiary age are also found in the southern end of the Osgood Mountains and in the Hot Springs Range.

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[7] See footnote 1, page 6.
[8] See footnote 2, page 6.

CONGLOMERATE

Pebble conglomerate is poorly exposed at a few places in the lowland east of the southern part of the Osgood Mountains in secs. 5, 6, and 7, T. 36 N., R. 41 E., and secs. 1 and 12, T. 36 N., R. 40 E. Because of the smallness and scarcity of exposures, the conglomerate is not shown on the map; and the alluvium derived from it, which is characteristically an unconsolidated sand with pebbles of chert scattered through it, was not mapped separately from the Quaternary alluvium. The sandy aluvium with chert pebbles differs from alluvium along the stream courses, which contains abundant coarse debris of volcanic rock and quartzite.

Pebbles of the conglomerate are mostly black, brown, tan, and green chert, and smaller amounts of gray quartzite. In addition, some fragments of gray and tan porphyritic silicic volcanic rock were also observed. The pebbles are mostly subangular, but they range from angular to subrounded, and from 2 to 3 mm to as much as 6 cm in maximum dimension. A matrix of sand and opal cement binds the pebbles together fairly firmly, yet loosely enough so that pebbles can be broken free from the matrix without fracturing.

Alluvial sand containing chert pebbles is also common on the west side of the Osgood Mountains at the south end of the range. In a ravine in SW¼SW¼ sec. 23, T. 37 N., R. 40 E., poorly consolidated gravel with pebbles of chert and some light-colored volcanic rock is exposed in a shallow shaft. The hill to the south and west is capped by andesitic flow rocks, which apparently overlie the gravel.

Four small remnant patches of conglomerate that rest on the Harmony formation were discovered in the Hot Springs Range at widely separated localities: in the southern part of the range in center SE¼ sec. 27, T. 38 N., R. 40 E., on the northeast side of a ridge slightly below exposures of lava; on the west side at two places on a ridge south of Dutch Flat in center NE¼ and north of center SE¼ sec. 20, T. 38 N., R. 40 E., where it apparently underlies flow remnants; and north of Mill Creek, also on the west side of the range, on the ridge in SW¼ sec. 3, T. 38 N., R. 40 E., where it occurs east of a down-faulted flow remnant. The fragments are subrounded to well-rounded pebbles, cobbles, and boulders. Many are obviously feldspathic sandstones and grits derived from the Harmony formation. In addition are plentiful pebbles of dark chert and clean fine-grained white quartzite. Mostly, the conglomerate is not exposed but weathers down to a loose sandy soil with many pebbles and boulders. At the first locality mentioned, in the southern part of the range, however, an estimated 25 to 30 feet of moderately well lithified conglomerate is visible in an old prospect shaft.

RHYOLITE TUFFS

DISTRIBUTION

Tuffaceous rocks underlie the flow rocks at places in the south end of the Osgood Mountains and are exposed in some low ridges and knobs in the lowland east of the southern part of the range. Generally, the tuffaceous unit is covered by talus and soil wash from the steeper slopes above because it is thin and softer than the overlying flows. The tuffaceous rocks are probably discontinuous, and many of them are remnants of a unit that was formerly much more continuous and was largely removed by erosion before the flows were laid down.

Bedded pumiceous tuffs are poorly exposed in some small ravines in SW¼SW¼ sec. 35, T. 37 N., R. 40 E., and NW¼ sec. 2, T. 36 N., R. 40 E. A small exposure of tuff is found on the southeast side of the Dry Hills in the northern part of sec. 28, T. 38 N., R. 41 E. The only tuff known in the Hot Springs Range is in a small isolated outcrop in a gully below some andesite in NE¼ SW¼ sec. 21, T. 39 N., R. 40 E., in the extreme northwest corner of the quadrangle; elsewhere Tertiary andesite rests directly on the Harmony formation.

Light-gray to greenish-gray fine-grained thinly stratified to massive welded rhyolitic tuffs make up parts of some small low ridges and knobs on the southeast side of the Osgood Mountains in sec. 6, T. 36 N., R. 41 E., and sec. 12, T. 36 N., R. 40 E. Characteristically, they have a very fine grained dense porcellaneous appearance, but they contain a few brilliant 0.5 to 1.0 mm crystal fragments of quartz and feldspar. Many pinprick-like open cavities are visible with a hand lens.

PETROGRAPHY

The pumiceous tuffs are commonly buff or light gray to white, and some are dark gray. They are more or less stratified and obviously fragmental. Microscopic examination shows that some are almost almost wholly composed of pumice and fragments of very fine grained vesicular to scoriaceous volcanic rock composed of glass and tiny laths of feldspar. Some contain discrete mineral grains in addition to volcanic rock fragments and pumice; a typical specimen contains fragments of orthoclase and sodic plagioclase, a little quartz, hypersthene and augite, hornblende, and small amounts of magnetite and zircon.

The welded tuff is composed mostly of devitrified glass in which the original shard outlines are plainly visible. Welding has been only slight, for the shards are mostly undeformed and most of the vesicles show no flattening; yet there was some compaction while the glass was still in a hot, plastic condition, for the shards show slight tendencies to wrap around crystal fragments. The crystal fragments, many of which show resorption by the groundmass, are quartz and orthoclase. The vesicular cavities are lined with finely crystalline tridymite and sanidine, and some contain a little brownish opal.

BODY OF TUFFACEOUS ROCK NEAR GETCHELL MINE

The north shaft at Getchell mine, approximately half a mile north-northwest of the mill site, was sunk in tuffaceous rock through a vertical distance of nearly 300 feet. In the 600-level crosscut from the shaft, the tuff is in contact with hornfels 180 feet west of the shaft, and the contact seems to be mostly vertical, though lagging of the walls and roof make observations difficult. Tuff is also exposed in the walls of a small room, excavated in the hillside for storage of fuse and dynamite caps, about 1,000 feet south of the shaft. We have not shown the tuff on the geologic map (pl. 1) because alluvium and slope wash cover make it impossible to determine its true extent. Hobbs (1948,[9] pl. 2) showed it as an elliptical area nearly 2,500 feet long and 1,250 feet wide elongate from north to south, and Joralemon (1951, fig. 3) showed a body of similar size but somewhat different shape.

Bedding in the tuff could not be seen. Underground near the north shaft it is a structureless light-buff soft porous fine-grained rock containing small recognizable pumice fragments, a few flakes of biotite, and angular grains of quartz and feldspar. Some angular fragments of quartzite and shale as much as 1.5 inches across were also identified. Some blocks of granodiorite that range from a few inches to several feet in diameter are enclosed in the tuff in the small underground excavation south of the shaft. The blocks are subangular to subrounded and are considerably weathered.

Hobbs (1948,[10] p. 44–45) and Joralemon (1949,[11] p. 34; 1951, p. 269), in order to explain the relatively thick body of tuff with limited areal extent, regarded it as a roughly conical pipe formed by explosive volcanic action. Without more substantial evidence as to the form of the tuff body, however, we do not believe that this is its only explanation. Elsewhere in the region the tuff varies widely in thickness over short distances. Situated as it is near the front of the range, where undoubtedly more faults exist than are recognized at the surface, it seems likely that a downfaulted block of an originally more extensive tuff unit might have been preserved from erosion. We do not believe that the blocks of granodiorite in the tuff near the surface are necessarily fragments torn from the granodiorite by volcanic explosion. Their form and rather strongly weathered condition suggest that the deposit may be a reworked tuff, possibly a mudflow that incorporated foreign fragments from the surface over which it moved.

[9] See footnote 1, page 6.
[10] See footnote 1, page 6.
[11] See footnote 2, page 6.

FLOWS

Andesitic flows cover the southern one-fifth of the Osgood Mountains and extend about 2 miles southward into the Golconda quadrangle (Ferguson, Roberts, and Muller, 1952). Remnants of the same kind of rock are scattered north of this main area of flows for about 2 miles. Flows of andesite also make up the Dry Hills on the west side of the Osgood Mountains, and remnants of similar flows lie along the east and west flanks of the Hot Springs Range and occur in small patches within the range. Basaltic lavas make up Soldier Cap on the west side of the Osgood Mountains. The general distribution of these flows suggests that the region was once more or less covered by volcanic rocks.

Classification of these flow rocks is based on chemical analyses of typical examples. We have used a modification of the scheme used by Williams (1950, p. 234–235; Williams, Turner, and Gilbert, 1954, p. 43) to distinguish between andesite and basalt: lavas containing less than 54 percent SiO_2, with negative qz values and without normative quartz are classed as basalts; and rocks containing more than 55 percent SiO_2 and having high positive qz values are classed as andesites.

ANDESITE FLOWS OF THE SOUTHERN OSGOOD MOUNTAINS, DRY HILLS, AND HOT SPRINGS RANGE

The andesite flows are typically dark gray to black, but in some places where they are scoriaceous they are brick red; however, black scoriaceous rocks are also common. Many of the flows have a platy fracture, but more commonly they break with smooth conchoidal surfaces. Some are massive, many contain a few vesicles, and some are highly vesicular or scoriaceous. Flow structures are not prominent, for the most part, but are shown here and there by the subparallel orientation of feldspar laths and elongation of vesicles. Commonly where the flow is highly vesicular it also has a twisted bulbous appearance with slightly ropy surfaces. In a few places cavernous forms have been seen. In some areas the soil above the flows contain many small flat chips of light-colored chalcedony, apparently derived from chalcedony-filled fractures. Some surfaces bear mammilary coatings of clear glassy, rarely somewhat iridescent, opal.

The flow rocks are uniformly very fine grained to aphanitic. Some contain visible lathlike phenocrysts of plagioclase seldom more than 1 or 2 mm long and less than 1 mm wide, and less abundant millimeter-sized grains of olivine; aphanitic varieties generally have no prominent phenocrysts. The rocks also commonly contain glass. Texturally, therefore, most of them are hyalopilitic, although some are pilotaxitic and contain no glassy residue. Most are porphyritic, but the phenocrysts are commonly of microscopic size.

Plagioclase, olivine, and orthopyroxene constitute phenocrysts and microphenocrysts within a groundmass of plagioclase laths, intergranular olivine, and both orthopyroxene and clinopyroxene, accessory magnetite, and, in some specimens, apatite, and glass.

The plagioclase phenocrysts are commonly zoned and are somewhat more calcic (An_{57-62}) than the groundmass laths (An_{55} on the average).

Many specimens contain minor amounts of olivine, mostly as phenocrysts but also in the groundmass as intergranular crystals between plagioclase laths; some varieties of andesite, however, contain no olivine. Measurements of optic angles suggest an olivine composition range from Fo_{59} to Fo_{72}.

Faintly pleochroic hypersthene is common in most specimens that were examined microscopically but is absent from a few; it occurs both as microphenocrysts and as minute stubby grains in the groundmass. Compositions, as determined from measurements of the optic angle, range from En_{80} to En_{85}. The interstitial clinopyroxene is generally too fine grained for reliable identification, but it appears to be an augite or diopsidic augite.

Minute acicular crystals of apatite are plentiful in the groundmass, and a percent or so of magnetite is found in most specimens as small scattered grains and as dust in the glass. The interstitial glass is dark brown and translucent to nearly opaque. Trace amounts of calcite have been observed in some specimens, which seems to have crystallized in an interstitial position in the groundmass and not as cavity fillings.

Analyses and model compositions of typical andesites from the area are given in table 15, Nos. 2, 3, and 4.

Perhaps some of the flow rocks are more properly to be classed as basaltic andesites, though none that have been chemically analyzed fall into this category. Although texturally the same as the andesites, some that were examined microscopically contain slightly more olivine and pyroxene, the plagioclase may be somewhat more calcic, and the hypersthene and clinopyroxene tend to be somewhat coarser than in the andesites. Stubby crystals of the clinopyroxene occupy an interstitial position and in some specimens may have a subophitic relation to the groundmass plagioclase. Optic angles ranging from approximately 44° to 47° and extinction angles ($z \wedge c$) from 40° to 43° suggest common augite. Clinopyroxene also occurs in some specimens as light-brown groups of subparallel needles, blades, and a few well-formed long prisms in the interstitial glassy material. The needlelike crystals are filled with opaque dusty inclusions of magnetite, which are less abundant in the stubby better developed crystals and appear to have been interrupted in the process of crystallizing from the interstitial liquid by chilling of the lava. Their optical properties are difficult to measure, but the optic angle is large (approximately 60°), though variable, and the extinction angle $z \wedge c$ is also large and variable. Magnetite (probably magnetiteilmenite) occurs as dust and minute grains and, in some specimens, as chainlike and elongate skeletal crystals. In contrast to the andesite, acicular apatite crystals are very uncommon. A little calcite occurs interstitially in the groundmass of some specimens. The residual glass is brown and commonly dusty with opaque inclusions. In some specimens it is faintly anisoptropic due to slight devitrification.

OLIVINE BASALT ON SOLDIER CAP

A pile of many relatively thin sheets of olivine basalt forms the prominent layered mesalike feature known as Soldier Cap north of Cave Canyon on the northwest side of the Osgood Mountains. The flows also cap a similar unnamed feature south of Cave Canyon and are present on two ridges north of Soldier Cap. The rock of these flows is petrographically quite distinct from the andesite in the Dry Hills to the south. The lower valley of the East Fork of Eden Creek separates the olivine basalt flows from the andesite, and the relative ages of the two volcanic units cannot be established.

The contact of the lava pile with the underlying rock dips from 20° to 30° NW., whereas the flows themselves are nearly horizontal or inclined slightly westward, indicating the flows were extruded into an area of considerable relief. It is a puzzling fact that flows of similar composition have not been recognized elsewhere in the quadrangle, and it is especially puzzling that they apparently do not cross the East Fork of Eden Creek into the Dry Hills and that the andesite in the Dry Hills is not found north of the creek.

In the upper part of the olivine basalt pile, individual flows are easily recognized for they form prominent steplike cliffs and benches. In the lower part of the section, possibly because individual flows are thicker, the steplike topography is not developed and the flow members cannot be seen. The flows range from approximately 10 feet to 30 feet in thickness; the majority probably are around 15 to 20 feet thick. The total thickness of basalt is on the order of 1,250 feet.

The upper parts of individual flows are highly vesicular; the central parts are dense and considerably less vesicular; and the lower parts of the thickest flows contain a few small cavities. Freshly broken surfaces of vesicular rock are medium gray, mottled with areas of light brownish gray; specimens of the more massive rock are medium dark gray. Weathered surfaces are brownish gray to moderate brown. In general, the color of the fresh rock is lighter than the andesite and balsaltic andesite elsewhere in the quadrangle. In hand specimens the rock is porphyritic with millimeter-sized phenocrysts of olivine and, very rarely, plagioclase laths in an aphanitic groundmass.

The phenocrysts of olivine and plagioclase lie in a microcrystalline groundmass of plagioclase and intergranular pyroxene and olivine. All specimens contain magnetite and acicular apatite as minor accessories. A little interstitial brown glass occurs in some specimens. All specimens contain a little biotite interstitial to the essential constituents and in places seemingly occupying microscopic cavities. Montmorillonitic clay occupies a similar position and commonly is associated

TABLE 15.—*Analyses, norms, modes, and Niggli values of Tertiary flows in the Osgood Mountains quadrangle*

[Analysts (1, 3), J. M. Dowd and Katrine White; (2, 4) P. L. D. Elmore and Katrine White, U.S. Geol. Survey]

	1	2	3	4
Chemical analyses				
SiO_2	50.6	57.6	58.0	59.8
Al_2O_3	16.3	15.0	17.0	16.0
Fe_2O_3	6.9	2.0	4.6	2.4
FeO	2.4	6.5	2.6	3.8
MnO	.16	.18	.12	.15
MgO	5.6	2.5	3.1	2.8
CaO	7.1	5.5	5.5	5.0
Na_2O	4.0	3.5	3.9	3.4
K_2O	2.1	2.7	2.8	3.2
TiO_2	1.9	1.9	1.2	1.0
P_2O_5	.76	.82	.66	.54
Ig	2.0	1.3	.88	1.1
H_2O+	-----	-----	-----	-----
CO_2	-----	-----	-----	-----
Sum	99.8	99.5	100.4	99.2
Sp gr (powder)	2.82	2.73	2.72	2.69
C.I.P.W. norms				
Q	-----	11.5	10.1	13.3
Or	12.2	16.1	16.7	18.9
Al	34.1	29.3	33.0	28.8
An	20.3	17.2	20.6	18.9
Wo	4.3	1.9	1.6	0.9
En	13.6	6.2	7.7	6.9
Fs	-----	7.4	-----	3.6
Fo	.5	-----	-----	-----
Mt	2.6	3.0	5.1	3.5
Il	3.6	3.6	2.3	2.0
Hm	5.1	-----	1.1	-----
Ap	1.7	2.0	1.3	1.3
Sum	98.0	98.2	99.5	98.1
Niggli values				
al	25.2	28.8	31.5	32.1
fm	41.2	35.4	32.4	31.5
c	20.0	19.2	18.5	18.2
alk	13.7	16.6	17.6	18.2
si	133	188	182	206
k	.253	.342	.323	.382
mg	.534	.343	.448	.455
qz	−22	22	11.6	33

	1	2	3	4
p-values and Rittman classification				
p-value (Rittman, 1953)	49.5 (medium alkaline)	55.2 (weak calc-alkaline)	56.6 (weak calc-alkaline)	58.0 (weak calc-alkaline)
Classification (Rittman, 1952)	B5 (olivine andesine trachybasalt)	B5 (trachy andesite)	B5 (trachy andesite)	A5, FM 13 (trachy andesite)
Modes (volume percent)				
Phenocrysts:				
Olivine	[1]1	-----	-----	-----
Pyroxene	-----	1	-----	1
Plagioclase	-----	3	-----	-----
Groundmass:				
Plagioclase	59	27	72	61
Pyroxene	17	2	13	11
Olivine	[1]10	3	3	2
Magnetite	7	2	7	3
Glass	-----	62	-----	21
Cristobalite	3	-----	-----	-----
Apatite	1	Tr	Tr	Tr
Biotite	1	-----	-----	-----
Iron oxide	1	-----	-----	-----
Indeterminate groundmass	-----	-----	5	1
Sum	100	100	100	100

1. Olivine basalt from Soldier Cap, Osgood Mountains, sec. 2, T. 38 N., R. 41 E. (H5052.)
2. Andesite, Belmont Hill, Hot Springs Range, SE¼ sec. 8, T. 38 N., R. 40 E. (53W95.)
3. Andesite, west side southern end of Osgood Mountains, SE¼ sec. 21, T. 37 N., R 40 E. (OM251.)
4. Andesite, Dry Hills, SE¼ sec. 20, T. 38 N., R 41 E. (76H53.)

[1] Partly to wholly altered to iddingsite.

with the biotite. Some small cavities are partly to wholly filled with opal; others contain, in addition to opal, a zeolite, feldspar, and commonly biotite.

Olivine shows all stages of alteration to reddish-brown iddingsite and hematite. Phenocrysts are seldom completely altered, whereas the smaller interstitial grains in the groundmass are totally replaced by reddish-brown alteration products in most specimens. A less common type of alteration seen in the phenocrysts is the replacement of the olivine by serpentine and bowlingite along fractures. Some phenocrysts are crowded with grains of secondary magnetite. The phenocrysts are euhedral to subhedral and show virtually no resorption. Optic angle measurements indicate that the composition of the olivine ranges from Fo_{85} to Fo_{95}.

Two kinds of plagioclase can be seen in the groundmass of many specimens. First, and most abundant, are euhedral to subhedral twinned and slightly zoned laths of labradorite (An_{62-68}). Second, interstitial to the labradorite laths, are anhedral crystals that show wavy extinction, are commonly untwinned, and contain many inclusions of apatite needles. The refractive index (β) of the interstitial feldspar is distinctly less than that of the labradorite laths—it ranges about 1.540 and averages approximately 1.539—and has an optic angle of moderate size and positive sign. MacDonald (1942) described similar interstitial feldspar in Hawaiian lavas characterized by an abnormally small positive optic angle and with indices of oligoclase or andesine that he concluded was a potassic oligoclase or potassic-andesine. Some feldspar that may be anorthoclase, as indicated by its abnormally small optic angle of negative sign and refractive index higher than that of orthoclase, is also associated with the interstitial plagioclase.

The pyroxene, occurring exclusively as granules and stubby prisms in the groundmass, is mostly so small that its optical properties are very difficult to determine. It is a colorless to very faintly brown common augite. Some of the larger grains have the following characteristics: $\beta=1.695$, $2V(+)=49°$ to $51°$.

The biotite, which is a common minor constituent in all specimens, is strongly pleochroic from reddish brown to very pale yellow, with observed optic angles near 29° on the average, refractive indices $\beta=1.633$ and $\gamma=1.634$, and strong dispersion $r>v$. Its position in interstitial cavities in the groundmass and in amygdules indicates that it is of late magmatic or hydrothermal origin.

An analysis and the modal composition of the lava from Soldier Cap are given in table 15.

CHEMICAL COMPOSITION OF THE TERTIARY FLOWS

Table 15 lists the chemical compositions of four samples of flows together with modes, computed norms, and Niggli values; the results of spectrographic analysis for minor elements in three of the samples are given in table 16. On the basis of these few analyses, the alkali-lime index—that is, the SiO_2 value at which Na_2O+K_2O equals CaO—appears to be approximately 59, and the lavas belong to the calc-alkalic igneous rock series of Peacock's (1931) classification. Rittman (1953) has devised a system of calculation by means of which a number which he designates as the *p*-value can be used to indicate the character of either a series or single rocks. Where only a few analyses are available, Rittman's method has an advantage over Peacock's classification because no variation diagram is needed to obtain the index. The *p*-values for the analyzed volcanic rocks of the Osgood Mountains quadrangle are given in table 15 and shown on figure 6 (Nos. 1 to 4). Three of the analyzed samples (Nos. 2 to 4) have similar *p*-values which average 56.6 and according to Rittman's scheme are weakly calc-alkaline; these we regard as essentially contemporaneous and representative of the flows in the quadrangle with the exception of those at Soldier Cap. The basaltic rocks at Soldier Cap probably are a different series; for them $p=49.5$, medium alkaline.

TABLE 16.—*Quantitative spectrographic analyses of minor elements in volcanic rocks*

[Harry Bastron, analyst]

	1	2	3		1	2	3
Cu	0.006	0.001	0.005	Yb	0.0002	0.0004	0.0003
Mn	.07	.07	.06	La	.007	.004	.01
Co	.003	.001	.001	Ti	1.0	1.0	.6
Ni	.01	------	.003	Zn	.01	.02	.01
Ga	.001	.001	.001	Sr	.1	.05	.07
Cr	.002	------	.002	Ba	.2	.2	.1
V	.02	.01	.01	P	.2	.2	.2
Y	.003	.005	.004	B	.001	.001	.001

1. Olivine basalt from Soldier Cap, Osgood Mounatins, sec 2, T. 38 N., R. 41 E. (H–50–52).
2. Andesite, Belmont Hill, Hot Springs Range, SE¼ sec. 8, T. 38 N., R. 40 E. (53–W–95).
3. Andesite, Dry Hills, SE¼ sec. 20, T. 38 N., R. 41 E. (76–H–53).

Looked for but not found: Au, Ag, Mo, Hg, Ru, Rh, Pd, Ir, Pt, W, Re, Ge, Sn, Pb, As, Sb, Bi, Te, Zn, Cd, Tl, In, Sc, Th, Nb, Ta, U, Be, Li, Cs
The above results have an overall accuracy of ±15 percent the value shown.

It is apparent from the analyses and the amount of normative orthoclase that these lavas have a relatively high K_2O content compared to SiO_2. The high K_2O is also reflected by the naming of the analyzed rocks according to Rittman's (1952) proposed nomenclature of volcanic rocks, for they are classified as trachyandesite, trachybasalt, and rhyodacite (table 15). Although these rocks do not contain any recognizable potassium feldspar, the microcrystalline and glassy interstitial residue has potentially the composition of orthoclase plus quartz; the potassic character of the flows on Soldier Cap probably accounts in part for the late interstitial potassic-oligoclase in those rocks. If, as was done by Merriam and Anderson (1942, p. 1724), the *k*-ratio is plotted against SiO_2 (fig. 7), it is readily apparent that the Tertiary volcanic rocks (spec. 1 to 4) from the Osgood Mountains and Hot Springs Range, together with others from western Nevada, are intermediate between the potash-rich province of central and eastern Nevada and Utah, and the relatively low-potash province of the Sierra Nevada and Cascade Range in California.

The analyses bring out another feature worth noting; the andesitic lavas have a relatively high content of TiO_2. Presumably it is mostly in the form of titaniferous magnetite, for the pyroxenes do not have the purplish or brownish cast that typifies titaniferous varieties. It is also noteworthy that these rocks are exceptionally rich in barium (table 16).

AGE AND CORRELATION

The age of the volcanic rocks and the conglomerate can only be determined as post-Paleozoic and pre-Quaternary from their relations with other rocks in this quadrangle, but the Tertiary age of similar rocks in adjoining areas has been firmly established.

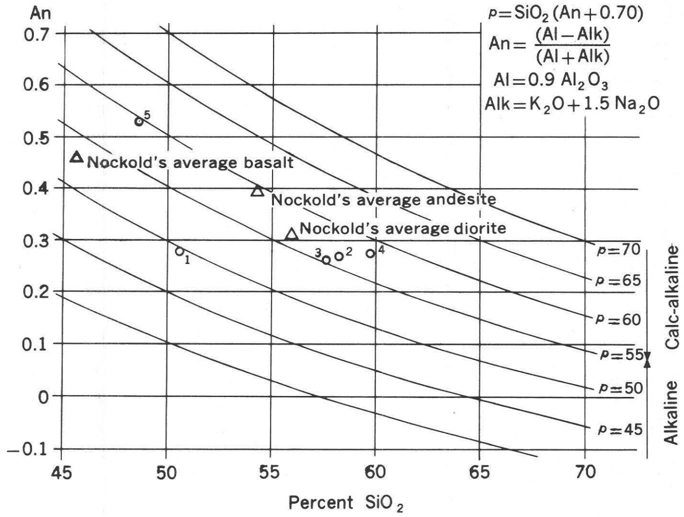

FIGURE 6.—Rittman (1953) diagram showing *p*-values of volcanic rocks in the Osgood Mountains quadrangle. Nos. 1 to 4 are Tertiary flow rocks, corresponding to same-numbered analyses in table 15; No. 5 is basalt near Comus station, table 17.

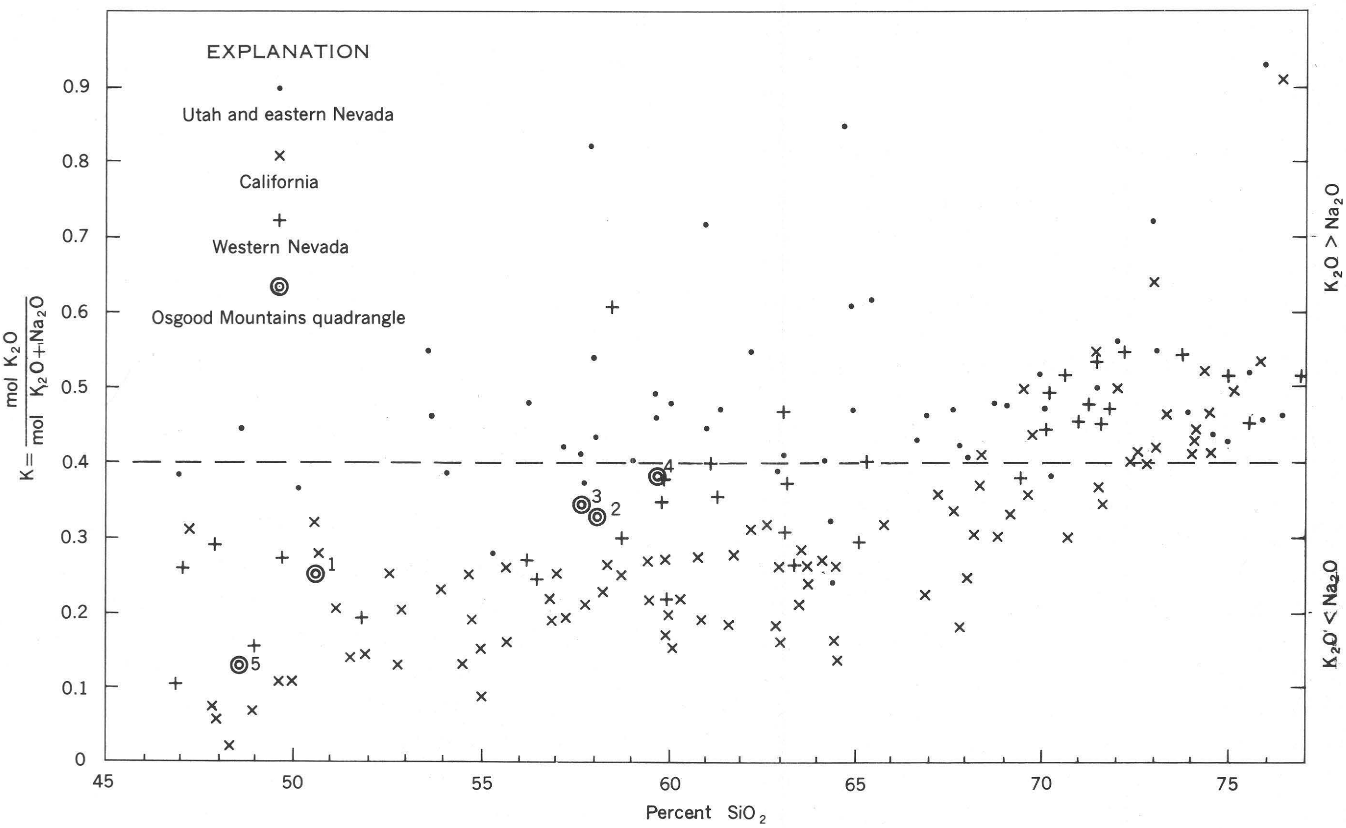

FIGURE 7.—Diagram of the *k*-ratio of volcanic rocks plotted against their SiO_2 content. (Taken from Merriam and Anderson, 1942, p. 1724, with some additional data.) Nos. 1 to 4 are Tertiary flow rocks, corresponding to same-numbered analyses in table 15; No. 5 is basalt near Comus station, table 17.

The tuffaceous rocks that underlie the andesitic flows are tentatively correlated with vitric tuffs and welded tuffs that constitute a widespread unit of late Miocene to middle Pliocene age in Nevada (Van Houten, 1956, p. 2814–2819). Similar beds at the north end of the Hot Springs Range yielded a transitional Miocene and Pliocene flora (Van Houten, 1956, p. 2805, fig. 2). The andesitic flow rocks that overlie the rhyolite tuffs are probably of early to middle Pliocene age (Nolan, 1943, p. 168–169; Van Houten, 1956, p. 2816, fig. 6). The pebble conglomerate underlies the flows in at least one place, but its relation to the rhyolite tuffs is uncertain. Van Houten (1956, p. 2814–2819) reported that Paleozoic pebble conglomerates occur sporadically through the vitric tuff unit.

Ferguson and others (1952) mapped gravels at the south end of the Osgood Mountains that might be equivalent with the pebble conglomerates in this quadrangle, but he considered them to be of Quaternary age.

ROCKS OF QUATERNARY AGE

Surficial deposits of Quaternary age in the Osgood Mountains quadrangle have been subdivided into older fan gravels, older and younger alluvium, and talus. A flow of basalt that extends into the southern part of the quadrangle, west of Comus station, is also believed to be of Quaternary age.

BASALT FLOW NEAR COMUS STATION

The northern end of a basalt flow that was mapped in the Golconda quadrangle (Ferguson, Roberts, and Muller, 1952) extends about half a mile into the south end of the Osgood Mountains quadrangle, west of Comus station on the Southern Pacific Railroad (pl. 1). Its physical relation to the other volcanic rocks in the quadrangle cannot be determined, but petrographically and chemically it is considerably different. It is tentatively believed to be of Quaternary age because it is considerably less dissected than the other flows and because it appears to have obstructed the course of the Humboldt River and caused it to flow around the end of the tongue of lava. The flow is tilted westward to northwestward at a small angle, possibly by normal faulting along its eastern edge.

The rock is dark gray, very fine grained to aphanitic with a vesicular and scoriaceous structure. It is texturally different from the flows in the Osgood Mountains and Hot Springs Range as it is typically basaltic with an intergranular to subophitic groundmass, a small amount of interstitial glass, and a few olivine and plagioclase phenocrysts. The chemical composition (table 17) is also much different from that of the other flows.

Olivine and plagioclase phenocrysts make up 6 percent of the total volume. The olivine (3 percent) is approximately Fo_{82} according to its optic angle ($2V=87°$). The plagioclase phenocrysts (3 percent) are labradorite to bytownite (An_{53} to An_{75}). The groundmass, about 94 percent of the total volume, includes about 12 percent olivine and 45 percent plagioclase (An_{47} to An_{68}). Intersertal to subophitic clinopyroxene (27 percent) in the groundmass is pale greenish-brown and nonpleochroic augite ($\beta=1.703$, $2V(+)=48°$, $r>v$ distinct). The rock also contains magnetite (3 percent) in small grains and rods, and interstitial brown glass (7 percent).

TABLE 17.—*Analysis, norm, mode, and Niggli values of basalt near Comus station. Railroad cut west of Comus, SE¼SE¼ sec. 10, T. 36, N., R. 41 E. (H–53–54)*

[Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C; P. L. D. Elmore, K. E. White, analysts, U.S. Geol. Survey]

Chemical analysis		Norm		Mode		Niggli values	
SiO_2	48.6	Q	------	Plagioclase	48	al	20.1
Al_2O_3	15.4	Or	3.6	Pyroxene	25	fm	48.6
Fe_2O_3	2.9	Al	21.1	Olivine	17	c	25.1
FeO	7.1	An	29.0	Magnetite	3	alk	6.2
NnO	.16	Wo	7.2	Glass	7	si	108
MgO	9.2	En	17.3			k	.137
CaO	10.6	Fs	6.4			mg	.624
Na_2O	2.5	Fo	3.9			qz	−17
K_2O	.60	Fa	1.6				
TiO_2	1.4	Mt	4.2			p-value (Rittman. (1953)=59.4 (weak calc-alkaline)	
P_2O_5	.24	Il	2.7				
H_2O+	.54	Ap	.6				
CO_2	.74	Cc	1.7				
		H_2O	.5				
Sum	100		99.8		100	Rittman (1952) classification: B8, FM<6ca″, FM=29.3 Basalt.	

The chemical composition of a sample of the flow is included in table 17. According to Rittman's (1952) classification, the rock is a labradorite basalt; its *p*-value of 59.4 is slightly greater than that of the andesitic flows in the quadrangle, but like them its character is weak calc-alkaline in Rittman's (1953) system (fig. 6). It is the most mafic volcanic rock in the quadrangle.

OLDER FAN GRAVELS

Remnants of older fan gravels are scattered along the east side of the Osgood Mountains from Felix Canyon on the north to a point on the south about 2 miles southwest of Lone Butte, and they are shown on the geologic map (pl. 1). The deposits are fanglomerates like the more recent alluvial deposits along the front of the range. They are as much as 25 to 30 feet above the surrounding alluvium in most places, and their upper surface may be as much as 50 feet above an adjoining drainage channel. Unlike the conglomerate that is believed to be of Tertiary age (p. 52), the younger deposits are generally coarser and poorly consolidated, but the chief difference is in the character of the fragments. The older fan gravels have only locally derived fragments, while the deposits of Tertiary age contain much exotic debris.

These older fan gravels apparently represent an extensive fanglomerate that once mantled bedrock along this part of the range. Accelerated erosion, perhaps caused by the slight uplift of the range along buried range-front faults, or lowering of the general base level resulted in deep incision of the gravels and their large scale removal by streams that issued from the range.

TALUS

Many slopes in the Osgood Mountains and Hot Springs Range are mantled with talus. No consistent criteria for mapping talus bodies have been followed, and only a few talus accumulations in the Osgood Mountains are shown on the geologic map (pl. 1). Generally, talus was mapped where it concealed critical structural elements such as faults and contacts between formations. In one or two places, however, it was mapped because it formed a large conspicous deposit. Most of the talus shown on the map is derived from the Osgood Mountain quartzite, although one prominent mass composed of volcanic rock is shown in the Dry Hills. Many other slopes underlain by volcanic rocks in both ranges are covered by talus, which forms dark streaks that are most conspicuous, even from a long distance.

ALLUVIUM

Alluvium covers about 45 percent of the quadrangle. It includes alluvial fans on the gentle slopes at the foot of the ranges; the veneer of gravels that cover pediment surfaces, gravels, and sands along the streams; and finer sands, silts, and clays that are found farther out in the main valleys and along the flood plain of the Humboldt River.

The thickness of the alluvium in the main valleys is not precisely known, but probably it is not less than several hundred feet in the axial parts. A well in Paradise Valley west of the northwest corner of the Osgood Mountains quadrangle is reported to have penetrated 800 feet of valley fill (Loeltz, Phoenix, and Robinson, 1949, p. 27). A well 400 feet deep was drilled 6 miles east of Getchell mine to supply water for the mill; presumably it did not reach bedrock. A veneer of alluvial gravel covers pediment slopes on the northwest and along much of the east side of the Osgood Mountains. The alluvial cover is less than 10 feet thick at many places but may be as much as 25 feet thick.

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Alluvium of two ages can be recognized. An older alluvium occurs on the upper parts of the fans and much of the veneer on the pediments; it can also be seen where it forms terraces along some of the streams. The older alluvium has been incised by the present streams, and a younger alluvium occupies the stream channels. In the lower parts of the fans and out in the main intermontane basins the younger alluvium overlaps the older alluvial deposits.

Near the mountains the alluvium is composed of coarse, bouldery to cobbly gravels intermixed with sand and silt; the material rapidly becomes finer toward the valleys. In the southeast corner of the quadrangle, the alluvium in the nearly flat part of the valley floor is silty, with only small local pebbly patches. Probably some of this fine material was deposited at times of flood by the Humboldt River. On the southwest side of the Osgood Mountains, south of Stone Corral, much of the alluvium, even fairly close to the mountain front, is composed at the surface of fine loose sand; rounded chert pebbles are locally abundant in the sand. Some of the sand obviously has been blown by the wind. It is possibly derived in part from fine sediments deposited to the west and southwest in Lake Lahontan. The eastern limit of Lake Lahontan was 2 miles or so east of Golconda (Russell, 1885, p. 31, pl. 46) and a small part of the shore would have lain in the southwest corner of the quadrangle, assuming that the original altitude of the Lahontan beach was 4,378 feet (Russell, 1885, p. 101). Part of the material, as attested by the pebbles of chert, probably is reworked sand from the prevolcanic Tertiary conglomerate.

METAMORPHISM

CONTACT METAMORPHISM AND METASOMATISM

A prominent metamorphic aureole surrounds the granodiorite stock in the northern part of the Osgood Mountains, and a small but well defined halo of metamorphism was formed around the small stock at Dutch Flat. A distinctive pattern has been used on the geologic map to show the distribution of metamorphic rocks around the stocks (pl. 1). Emplacement of the granodiorite bodies raised the temperature of the country rocks sufficiently to cause reorganization of the original constituents of the sedimentary rocks and crystallization of new minerals. For the most part these changes were attained without any addition of material from the granitic rock and are the result of reactions that took place between the original minerals of the rock through the medium of aqueous pore solutions, the water of which was supplied by the rock itself. Close to the contact, however, the metamorphosed rocks contain mineralogical, textural, and structural evidence that material has been added from the igneous body. The formation of the tungsten deposits in the Osgood Mountains was an integral part of the metamorphic process, for they were formed by solutions of the same origin as those that caused the formation of tactites from limestones adjacent to the granodiorite. For purposes of description, however, the tungsten deposits are considered separately in the chapter on ore deposits.

CONTACT METAMORPHISM AT DUTCH FLAT

A narrow belt of metamorphosed rock surrounds the small stock of granodiorite at Dutch Flat, where the rocks affected by the metamorphism are shale and feldspathic sandstone of the Harmony formation. On the north and east sides of the stock the rocks have been metamorphosed for only about 200 feet from the contact; on the south side metamorphic effects are recognizable for as much as 1,000 feet from the contact. The variation in width of the belt of metamorphism has no apparent relation to the structure of the sedimentary rocks or exposed shape of the stock.

The first sign of metamorphism is a darkening and hardening of the shale and sandy shale of the Harmony formation. The microscope shows that there has been some recrystallization of the finer material in the rock and the formation of some chlorite and much fine-grained muscovite. About 100 feet from the contact, spongy porphyroblasts of andalusite appear, as well as small flakes of biotite. Chlorite persists to within 50 feet of the contact. Introduction of orthoclase is a fairly common phenomenon, not recognized in the metamorphosed shaly rock but observed in some of the sandstones near the granodiorite. The introduced feldspar occurs as irregularly shaped bodies in the matrix of the sandstone. Some clinozoisite accompanies the orthoclase, and some specimens contain carbonate veinlets with intergrown prehnite.

CONTACT METAMORPHISM IN THE OSGOOD MOUNTAINS

The granodiorite stock in the Osgood Mountains is surrounded by a metamorphic aureole that varies considerably in width, a variation that for the most part seems to be a reflection of the shape and form of the intrusive body. At the south end of the stock, where the contact presumably is steeply dipping to vertical, the zone of contact metamorphism is narrowest, measuring from less than 1,000 feet to 1,500 feet in width, though in a few places metamorphism extends somewhat farther, parallel to the structure of the sedimentary rocks which strike into the contact. North of the stock, however, metamorphic effects are noticeable as far as the quadrangle boundary, or nearly 10,000 feet from the outcrop of the main intrusive body, which has a decided prolongation in this direction and, as indicated by the presence of small outliers, probably is continuous beneath a relatively thin cover of country rock for perhaps 5,000 feet north of the main exposure. The metamorphosed belt is from 2,000 feet to nearly 8,000 feet wide along the west side of the stock. On the east side, the width of the belt is uncertain because the rocks pass beneath alluvium and only a narrow strip of metamorphic rock is exposed.

Most of the sedimentary rocks metamorphosed by the granodiorite in the Osgood Mountains are the shales and carbonate rocks of the Preble formation. The Comus formation, which is near the east side of the stock and on a thrust plate at the north end of the range, is also affected. Some of the other formations come within the metamorphic aureole but are so distant from the intrusive body that they have been only slightly affected. As would be expected, the metamorphism has caused the formation of a considerable variety of rock types, containing several metamorphic minerals whose variations in composition and distribution depend partly on the composition of the original rock, partly on the distance from the intrusive body, and partly on the addition of substances by solutions derived from the granodiorite. In a general way, the metamorphic aureole is zoned with respect to the stock, for the products of the most intense metamorphism are at the contact between the stock and the country rock, and with increasing distance from the stock they grade rapidly into a less strongly metamorphosed zone of variable width that fades off gradually into unmetamorphosed country rock. Excepting the tactites formed from carbonate rocks at the contacts of the granodiorite, most of the metamorphism has been accomplished by reactions between the original constituents of the rocks with no addition of material and hence is designated isochemical metamorphism. Where the metamorphism has been accomplished by the addition of substances to form garnet-pyroxene tactites from the carbonate rocks, the process is designated allochemical silicate metamorphism (Holser, 1950, p. 1072).

METAMORPHISM OF THE PELITIC ROCKS

Argillite and hornfels formed by the metamorphism of shale and siltstone belonging to the Preble and Comus formations are the most common metamorphic rocks that surround the granodiorite stock. In the outermost zone, the most obvious metamorphic effect in these fine-grained sedimentary rocks is a hardening and loss of fissility commonly accompanied by a darkening of the rock. A faint foliation made visible by slight color and textural differences may be apparent in some specimens. These rocks are so fine grained that their mineral components are difficult to resolve even by microscopic study, but even at this early stage of metamorphism biotite is visible as microscopically small flakes. This early biotite characteristically is pale and its pleochroic colors vary from light brown to very pale yellow. Sericite and quartz can also be recognized.

The appearance of the minerals cordierite ($2MgO \cdot 2Al_2O_3 \cdot 5SiO_2$) and andalusite ($Al_2O_3 \cdot SiO_2$) is indicative of a further increase in the degree of metamorphism of the pelitic rocks. Cordierite is more common than andalusite in the metamorphosed shale and siltstone of the Preble formation, but in the fine-grained sedimentary rocks of the Comus formation andalusite seems to have formed more readily than cordierite. Exceptions to this general rule are to be found, of course, and some hornfels contains both minerals. Undoubtedly, however, the tendency for one or the other to be formed reflects differences in the original sedimentary rock. Tilley (1924, p. 31) has pointed out that andalusite normally arises from the metamorphism of kaolin-bearing rocks, and if magnesia is abundant, generally in the form of chlorite, cordierite is formed.

Cordierite appears in the pelitic rocks where the intensity of metamorphism is apparently only slightly greater than that which caused the formation of biotite. In fact, in many places cordierite and biotite seem to make their appearance simultaneously; however, biotite is found in rocks farther from the granodiorite than cordierite. In the Osgood Mountains cordierite is found in hornfels as much as 5,000 feet horizontally from the granodiorite, and in rocks of the appropriate composition it persists up to the contact. Its first appearance is commonly signalled megascopically in otherwise microcrystalline rock by minute ovoid spots that are somewhat lighter than the rest of the rock. Under the microscope these spots appear as relatively large anhedral cordierite crystals crowded with fine flakes of biotite and indeterminate fine groundmass material. The biotite-cordierite hornfels persists without much change over most of the width of the contact aureole, though as the granodiorite is approached the cordierite porphyroblasts contain somewhat less included biotite and the biotite becomes perceptibly darker.

In some of the rocks andalusite forms instead of cordierite. The shales of the Comus formation in several places east of the granodiorite body, for example, have been metamorphosed to hard microcrystalline dark-gray to black rocks in which small prismatic porphyroblasts of andalusite (commonly the variety chiastolite) can be recognized with the unaided eye. The andalusite is accompanied by muscovite instead of the biotite which is common in the cordierite hornfels. Some of the hornfels recrystallized from shales of the Prebel formation contain andalusite, with or without cordierite; in these rocks the andalusite forms anhedral spongy porphyroblasts that enclose minerals of the ground mass and is partly altered to matted aggregates of sericite.

Within a few feet of the granodiorite contact—generally less than 25 feet, most commonly 10 feet or less—the grain size of the hornfels increases and the metamorphosed rock has a granoblastic texture. Most commonly these hornfelses are biotite-cordierite-plagioclase-quartz assemblages, though some contain muscovite in place of the cordierite. Locally, hornfels close to the granodiorite contains orthoclase. In these rocks muscovite is not primary, though it may be present as a retrograde alteration product of andalusite or cordierite. Common assemblages of this type are cordierite(less commonly andalusite)-biotite-orthoclase-minor plagioclase-quartz. The biotite in those rocks that have been more strongly metamorphosed is typically dark and strongly pleochroic from moderate reddish brown to colorless. Plagioclase, which is rarely seen in the lower grade rocks, is in well-defined anhedral grains, mostly untwinned, and approximates andesine in composition. Cordierite in some specimens close to the granodiorite contact is partly to almost wholly altered to a low birefringent or isotropic substance (pinite?) and pale phlogopitic mica. The biotite in some specimens is partly altered to chlorite. Probably these mineral alterations are due to a lowering of the temperature following metamorphism and are to be regarded as retrograde metamorphic effects. Chemical analysis of a biotite-cordierite-plagioclase-quartz hornfels is given in table 18.

Microscopic study of the hornfels shows that the following minerals are commonly present in minor amounts: tourmaline, sphene, zircon, magnetite, and graphite. Tourmaline is perhaps the most widespread minor accessory, for nearly every thin section contains several grains. Most of the grains are subhedral prisms that are pleochroic in shades of yellow and yellowish brown, indicating a member of the dravite–schorlite series (magnesium–iron). Tourmaline has been reported elesewhere as an introduced mineral in contact zones; however, here it is a common detrital mineral in the unmetamorphosed sedimentary rocks, and as it is no more common in hornfels near the granodiorite than in the outer parts of the metamorphic aureole, it probably is an original constituent. Zircon, too, probably is carried over from the unmetamorphosed sedimentary rocks.

TABLE 18.—*Analyses of cordierite hornfels and schistose biotite hornfels*

[Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C]

	1	2		1	2
Chemical analysis [P. L. D. Elmore, K. E. White, S. D. Botts, and P. W. Scott, analysts; U.S. Geol. Survey]			**Spectrographic analysis** [Harry Bastron, analyst, U.S. Geol. Survey]		
SiO_2	63.4	62.9	Cu	0.003	0.004
Al_2O_3	17.8	15.8	Pb	.002	[2] nd
Fe_2O_3	1.2	1.1	Mn	.008	.02
FeO	3.4	5.2	Co	.002	.001
MgO	3.6	4.9	Ni	.004	.004
CaO	1.1	2.1	Ga	.001	[2] nd
Na_2O	1.5	1.5	Cr	.006	.007
K_2O	4.5	3.9	V	.008	.007
TiO_2	.65	.60	Sc	.005	.005
P_2O_5	.24	.06	La	.02	.02
MnO	.02	.04	Ti	.4	.4
H_2O+	1.3	1.4	Zr	.01	.01
CO_2	.06	.05	Be	.0001	.0001
			Sr	.002	.003
Sum	99	100	Ba	.1	.08
Sp gr (bulk)	2.57	2.70	B	.02	.002
Sp gr (powder)	2.69	2.77			
Loss on ignition	[1] 2.4	--------			

Looked for but not found: Ag, Au, Hg, Ru, Rh, Pd, Ir, Pt, Mo, Re, Ge, Sn, As, Sb, Bi, Zn, Cd, Tl, In, Y, Yb, Th, Nb, U, P.

[1] Sample contained organic matter.
[2] nd, no data.

1. Cordierite hornfels from Top Row pit, near center SW¼ sec. 9, T. 38 N., R. 42 E.; 5 feet from granodiorite contact.
2. Schistose biotite hornfels from Riley mine.

An unusual type of metamorphic rock for this area, a fine-grained biotite-rich commonly schistose rock, occurs along the granodiorite contact at the Riley and the Riley Extension mines where it forms a continuous stratum ranging from 1 foot to 10 or 20 feet in thickness between the granodiorite and metamorphosed carbonate rocks. In hand specimens it is very fine grained, light brown with commonly a purplish hue, and although hard at most places where it is exposed underground—on surface exposures it commonly is soft and crumbly. Much of it has a visible foliation, but some lacks any apparent directional fabric. Under the microscope, textures are fine grained and range from granoblastic to schistose; the principal minerals are strongly colored red-brown biotite, muscovite, plagioclase, and quartz.

Chemical analysis (table 18) shows that this rock is of about the same composition as the nonschistose biotite-cordierite-plagioclase-quartz hornfelses occurring next to the granodiorite contact elsewhere. Possibly it was formed by metamorphism of a shale bed under conditions of stress that promoted the development of a directional fabric and inhibited the formation or caused the destruction of cordierite or andalusite. Because of the persistence of the unit along the granodiorite contact, even where the contact cuts across the bedding, Hobbs (1948,[12] p. 22; Hobbs and Clabaugh, 1946, p. 75) concluded that this unit represented a metamorphosed sheared zone in country rock adjacent to the granodiorite.

METAMORPHISM OF THE CARBONATE ROCKS

Most of the carbonate rocks within the metamorphic aureole surrounding the granodiorite stock belong, like the peltic rocks, to the Preble formation. Carbonate rocks of the Comus formation and the Etchart limestone are generally farther from the stock and therefore, have been affected by the metamorphism to a lesser extent.

The products of metamorphism of the carbonate rocks vary widely depending partly on their original composition and distances from the granodiorite, and partly on the extent to which materials have been added. At many places at the contact with the granodiorite, the carbonate rocks have been transformed to rocks which, from their mineralogical composition and structural relations, have surely resulted from the addition of some chemical compounds and the removal of others—principally CO_2. Away from the granodiorite contact, however, the relative importance of allochemical and isochemical metamorphism of the carbonate beds are not always easy to assess. Metamorphism has caused the formation of rocks which, given the proper amounts and kinds of impurities—such as SiO_2 and MgO—in the original sedimentary rock, could result from recrystallization without any addition of material, but with the loss of CO_2 and H_2O. Addition of materials in the correct proportions, however, could equally well have caused the formation of the same kind of rocks. There is no certain way to tell whether or not there has been an introduction of material without detailed tracing of strata from their unaltered condition to their metamorphosed equivalents, accompanied by chemical analyses and careful measurements to determine changes in thickness of beds.

Two well-defined metamorphic zones are commonly present where the carbonate rocks are in contact with the granodiorite; the tactite zone (Hess, 1919; Hess and Larsen, 1921, p. 251–253), composed of dark silicates that replace the carbonate rock; beyond the tactite, the zone of light-colored silicate rocks, commonly called calc-silicate rocks. The calc-silicate rocks in turn change outward into marble and silicated carbonate rocks. At some places a zone of marble separates the

[12] See footnote 1, page 6.

calc-silicate rocks from the tactite; in some places the tactite zone is lacking, and marble and calc-silicate rocks are in contact with the granodiorite. The marble and silicated carbonate rocks change outward into unmetamorphosed rocks through a zone of variable width around the stock. Small areas of more intensely metamorphosed rocks are found at a few places completely surrounded by weakly metamorphosed rocks.

MARBLE AND SILICATED MARBLE

The best example of isochemical metamorphism of carbonate rocks is limestone that has recrystallized to marble. Some of the limestone beds that contained small deposits of impurities were recrystallized to marble containing various amounts of silicate minerals.

Beds of pure limestone are recrystallized to medium-grained, locally coarse-grained, marble around the borders of the granodiorite stock for as much as 5,000 feet from the contact. Limestone of the Preble formation is ordinarily dark gray, and though recrystallized it maintains its color until near the granodiorite contact. The gray marble is distinctly granular, its grain size averaging about 0.2 to 0.4 mm. Near the stock, over distances ranging from a few tens of feet to several hundred feet, it is bleached to light gray or pure white, probably because of the removal of bituminous or carbonaceous matter which is responsible for the original dark color of the rock. In the outer part of the zone of bleaching, where removal of carbonaceous material is incomplete, the marble is streaked and mottled in gray and dark gray; locally it is very friable and has a dark-gray or black sooty appearance, possibly representing local redeposition of carbonaceous material carried away from the bleached marble. Bleaching of the marble is accompanied by an increase in grain size to an average of 1 mm, in places coarser. Holser (1950, p. 1069) noted a similar relation in the Philipsburg region, Montana, and concluded that "* * * bleaching is closely related to macrocrystallization * * *".

Silicate minerals of metamorphic origin have been formed in some of the impure carbonate rocks as much as 5,000 feet from the nearest surface exposure of granodiorite. Most commonly, mild metamorphism of slightly impure limestone results in a recrystallized matrix of calcite containing a few percent of tremolite as small prismatic blades and fibers or larger megascopically visible sheaves or bundles of fibers, and with or without scattered grains of detrital quartz. Commonly the original bedding, if any, is well preserved and made visible by parallel layers of different grain size and mineral content.

Shaly and silty limestones are recrystallized to finely layered rocks composed of alternating light-colored bands of calcite and tremolite, and darker, commonly greenish, laminae of diopside-quartz, diopside-quartz-calcite, and less commonly diopside-calcite-plagioclase-quartz assemblages.

Diopside and wollastonite instead of tremolite are observed in the more intensely metamorphosed impure calcareous rocks. Some of the purer slightly siliceous marbles are composed of calcite and wollastonite, but commonly wollastonite-bearing marble is found close to the granodiorite contact, where there is a strong possibility that the wollastonite has formed by the metasomatic introduction of silica. Light-brown isotropic grossularite [13] garnet accompanies the wollastonite at some places.

CALC-SILICATE ROCKS

Close to the granodiorite the carbonate rocks commonly are metamorphosed to light-colored calc-silicate rocks, which are mostly found closely associated with dark tactite. Calc-silicate rocks unaccompanied by tactite have been seen at a few places where limestones are in contact with granodiorite, but generally the two occur together. Similar rocks are mentioned in many published descriptions of contact metamorphic tungsten deposits, where they are commonly referred to as light silicate rocks or calc-silicate hornfels; these are the rocks of Hess' and Larsen's (1921, p. 251, 253) " zone of light-colored silicates." In addition, light-colored calc-silicate rocks are known at a few other places in the Osgood Mountains where no granodiorite is exposed.

The calc-silicate rocks near the granodiorite constitute a clearly defined zone that generally is parallel to the contact of the intrusive body and closely parallel to the bedding of the adjoining sedimentary rocks. The calc-silicate zone is situated in several different ways with respect to the marble and tactite, but mining developments have shown that the relations at any given locality tend to be the same throughout the deposit. At some places it lies between granodiorite and marble, in others it separates marble from the tactite zone, and in still other localities a belt of marble lies between the calc-silicate zone and tactite. No evidence of gradational relations between the calc-silicate zone and tactite has been seen; the contact between them is sharp.

The calc-silicate rocks are typically well layered, commonly thinly layered. Many are composed of alternating light and dark layers than range from less than ½ inch to 4 inches in thickness; some of the layers may, in turn, be finely laminated within them-

[13] Garnet associated with wollastonite in marble at the Riley Extension mine has a refractive index of 1.759 and a unit cell size of 11.86 A which indicate a high ratio of grossularite to andradite (Frietsch, 1957). Refractive index was measured by immersion methods; unit cell was determined by X-ray powder methods using Cu radiation, geiger counter diffractometer, and recording chart.

selves. The layering is compositional and probably reflects bedding in the original sedimentary rock. Most of the rocks are so fine grained that microscopic ex-examination is necessary to determine their mineral composition.

Many of the calc-silicate rocks are composed of alternating light-gray to white layers interbedded with generally thinner greenish-gray layers. The light layers commonly are the softer, and on weathered surfaces they are depressed, whereas the greenish-gray layers form hard ribs. The light layers may be mostly marble, although usually they are composed of mixtures of calcite–wollastonite, wollastonite-diopsidic pyroxene, wollastonite-diopsidic pyroxene-grossularite, and rarely, wollastonite-diopsidic pyroxene-idocrase. The greenish layers are composed of granoblastic assemblages of diopsidic pyroxene–plagioclase–quartz or diopsidic pyroxene–plagioclase–grossularite, less commonly diopsidic pyroxene–grossularite.

Some calc-silicate rocks are hard microcrystalline hornfels with a light-gray porcellaneous appearance. These rocks have layers of more or less uniform hardness, but they obviously are thinly layered to finely laminated. They are composed of microcrystalline granoblastic intergrowths of diopsidic pyroxene and plagioclase; some have wollastonite in addition. Megascopic layering is due to slight variations in the proportions of minerals, which may not be readily apparent under the microscope; dark-gray laminations in some specimens are caused by variations in amount of minute opaque grains, possibly carbonaceous material.

Another variety of calc-silicate rock is light gray with brown laminations composed of idocrase and grossularite garnet. The light-colored layers are composed of interlayered calcite and wollastonite; subsidiary amounts of diopsidic pyroxene are associated with some of the wollastonite in some specimens.

Calc-silicate rock at some places contains some interstratified dark biotite and cordierite hornfels in thin laminae or in layers an inch or so in thickness.

The composition of clinopyroxene in some of the calc-silicate rocks was determined by measurement of the refractive index β and the optic angle, used in combination with Hess' (1949, pl. 6, p. 641) composition diagram for the diopside-hedenbergite series. In the light layers where wollastonite is the predominant mineral, the intergrown pyroxene is diopside, whose refractive index is $\beta=1.676$ and $2V=57°$; in the greenish-gray layers it is salite—refractive index $\beta=1.699$ to 1.703. Two generations of pyroxene can be recognized in some specimens. For example, in one dominantly pyroxene rock, early fine-grained clear colorless diopside is replaced by coarser dusty slightly greenish salite.

Where garnet occurs in the calc-silicate rocks it is light brown, like that in silicated marble, and isotropic. Typical garnet from calc-silicate rock at the Pacific mine has a refractive index of 1.757, and unit cell of 11.85 A,[14] which indicate grossularite (Frietsch, 1957).

The light-colored calc-silicate rocks near the granodiorite contact, which are commonly associated with tactites and their accompanying scheelite deposits, were formed from well-bedded carbonate rocks. It is difficult, however, to say how much of the transformation was accomplished by heating rocks of the appropriate composition without addition of material, and how much of the metamorphism was due to metasomatic addition of some materials and loss of others. The interbedding of wollastonite and wollastonite-bearing carbonate layers with layers containing diopsidic pyroxene strongly suggests the metamorphism of interbedded impure limestone and dolomitic limestone. The thin layers of grossularite and idocrase in some calc-silicate rocks may originally have been aluminous shaly partings that reacted with adjoining calcareous layers. Field data suggesting that formation of the calc-silicate rocks was aided by addition of material, mostly silicia, are the occurrence of irregular-shaped replacement bodies of wollastonite in otherwise pure marble found at several places near calc-silicate rocks and the common close association of a zone of light-colored calc-silicate rocks with dark tactities, which have been formed by addition of material. There is some indication that the calc-silicate rocks were formed prior to the formation of tactite, which is certainly the result of large-scale addition of material, because at some places replacement veins of tactite cut calc-silicate rock (fig. 8).

Some light-colored calc-silicate rocks have been formed elsewhere in the Osgood Mountains away from the main granodiorite contact. For some there is evidence of a granitic intrusive nearby or at shallow depth; for others there is no direct evidence for the cause of the metamorphism, but it is assumed that the heat and solutions responsible came from a buried intrusive body nearby, or they traveled outward from granitic bodies along a favorable structure.

In the extreme northeast corner of the quadrangle, part of the Etchart limestone has been metamorphosed to a calc-silicate hornfels (pl. 1). In two irregular-shaped areas, light-gray to white microcrystalline commonly porcelaneous-appearing rock apparently overlies limestone that has not been much recrystallized. Because of its fineness of grain little can be seen in the rock, even with a hand lens, except for fibers of tremolite in some specimens and a faint layering in others.

[14] See footnote, p. 63.

FIGURE 8.—Calc-silicate rock cut by tactite at the Pacific mine pit. gd, granodiorite; T, tactite; csr, calc-silicate rock.

Under the microscope, however, the rock is seen to be a fine-grained (0.05 mm. average) crystalloblastic diopside-quartz-plagioclase rock, with a few percent of orthoclase, sphene, and some minute scattered grains of zircon. Some specimens contain a little tremolite and chlorite on the borders of small cavities containing calcite.

Presumably the calc-silicate hornfels is metamorphosed calcareous or dolomitic siltstone and shale, which is interbedded with essentially unmetamorphosed limestone. The area of hornfels is a mile or more from the main granodiorite stock, but one or two very small granitic bodies, not shown on the map, are exposed in the hornfels area. These and the rather extensive area of hornsfels suggest that an intrusive body underlies the area at no great depth.

In the NE¼ sec. 25, T. 38 N., R. 41 E., a sandy dolomitic member of the Etchart limestone has been metamorphosed to a fine-grained diopside-quartz-plagioclase hornfels. Nearby, some of the carbonate rock is recrystallized to tremolite marble. Here the hornfels can be related to the dikes and apophyses of granodiorite which intrude the rock.

In the NE¼ sec. 1, T. 38 N., R. 41 E., again a short distance below a thrust fault, in a "hot spot" a few yards square, limestone of the Preble formation has been metamorphosed to diopside-wollastonite-grossularite hornfels. Here the nearest outcrop of the main granodiorite stock is nearly half a mile to the east.

TACTITE

At many places where carbonate rocks have been intruded by granodiorite, a zone of dark silicate rock has been formed at the contact. This zone, commonly lying between the granodiorite and the zone of light-colored calc-silicate rocks or marble, is the innermost zone of contact metamorphism in the carbonate rocks. The general term, tactite (Hess, 1919, p. 378), is commonly used to designate the dark silicate rock, though the term skarn has about the same meaning. In the Osgood Mountains, as in many other contact metamorphic tungsten deposits, most of the scheelite is contained in tactite, though not all the tactite carries scheelite.

Unlike the zone of calc-silicate rock and marble, the limits of the tactite zone are clearly defined. Although it may be irregular, the contact between tactite and calc-silicate rock or between tactite and marble is abrupt and easily recognized. The inner limit between tactite and granodiorite is sharp and clean at many places, but it may be less abrupt where reaction and late stage alteration effects have resulted in a hybrid zone a few inches wide.

The tactites characteristically are dark heavy rocks of relatively simple mineral composition. Mostly they range from fine to medium in grain, and commonly they are massive; but in many places a compositional layering can be seen that is parallel to the bedding in the adjoining carbonate and calc-silicate rocks. The predominant constituents are calc-silicate minerals, which may include garnet, clinopyroxene, actinolitic amphibole, epidote, and idocrase; but typically only one or two mineral species compose the bulk of the rock. Minor quantities of quartz and calcite are found in nearly every specimen. Perhaps the most common rock types are garnet-pyroxene tactites; almost monomineralic garnetites have been formed in many places; and garnet-actinolite tactites are commonly encountered. Less abundant types are nearly pure actinolite tactites, pyroxene-quartz rocks, and varieties in which epidote is a plentiful constituent. An uncommon variety of tactite of only local occurrence is composed predominantly of dark grayish olive-green pyroxene without garnet.

Garnet of the tactites is dark reddish brown to dusky red, in contrast to the pale brown or honey-colored garnets of the silicated marble and calc-silicate rocks. The color also reflects the contrast in composition, for the garnet of the tactites contains substantial amounts of ferric iron—that is, the ratio of andradite ($3CaO \cdot Fe_2O_3 \cdot 3SiO_2$) to grossularite ($3CaO \cdot Al_2O_3 \cdot 3SiO_2$) is high. Garnets in contact-metamorphosed carbonate rocks commonly contain very small amounts of FeO, MgO, and MnO (Kennedy, 1953, p. 15). The measurements of refractive index and size of the unit cell as determined by X-ray measurements show that the ratio of andradite to grossularite is high (Frietsch, 1957).

	Refractive index	*Unit cell (A)*
1	1.810–1.820	11.92
2	1.811	11.97
3	1.815	11.89
4	1.824	11.95
5	1.861	11.95

Almost without exception zoning is visible in the garnet, in some specimens both under the microscope and megascopically. The zoning is made apparent in part by color differences; the central parts of zoned crystals are commonly nearly colorless, indicating a lower iron content than the outer parts, which are colored in shades of of pale brown to yellow. Under crossed polarized light the garnet in most specimens is anisotropic, and the zoned structure is made strikingly apparent by narrow concentric zones of varying birefringence, ranging from 0 to about 0.008. No clearly defined association of isotropic or anisotropic garnet with particular kinds of tactite can be established, but there is some evidence that anisotropic garnet is a late-stage modification of the isotropic variety. Isotropic garnet occurs with unaltered pyroxene, and tactites in which it occurs lack epidote, quartz, and sulfide minerals, which are known to be late-stage minerals elsewhere in the tactite zone. Some confirmation of this apparent relationship was observed in a specimen of an isotropic garnet-pyroxene tactite where the garnet adjacent to a quartz-calcite-epidote veinlet is slightly sheared and anisotropic.

Clinopyroxene of the diopside-hedenbergite series occurs as dark grayish-green layers in garnet-pyroxene tactite and as small subhedral granules poikilitically enclosed in the garnet. From the mutual boundary relations it is not clear that pyroxene is earlier than garnet, but in a few specimens garnet veinlets cut across the pyroxene layers. Viewed under the microscope, the clinopyroxene is pale green to almost colorless. Approximate compositions were determined from the refractive index β (Hess, 1949, p. 641) and measurement of the optic angle. From one specimen to another the clinopyroxenes range from salite to hedenbergite, but the composition in an individual sample is uniform, as evidenced by the constant optical properties. The refractive index β of the samples studied ranges from 1.694 to 1.727, and 2V ranges from 59° to 62°, indicating a variation in the diopside: hedenbergite ratio of from approximately 30:70 to 15:85 (Hess, 1949, p. 641).

In much of the tactite, clinopyroxene is partly or entirely uralitized. The alteration product is a fibrous pale-green pleochroic amphibole, probably near actinolite in composition. A little chlorite may accompany this secondary amphibole. An actinolite tactite composed of greenish-gray to dark greenish-gray matted fibrous actinolite has been formed in some places, presumably by the alteration of a pyroxene tactite. In some specimens the garnet is partly replaced by actinolitic amphibole. At a few places, principally at the Granite Creek mine, the pyroxene is replaced by a dark hornblende instead of actinolitic amphibole. The following optical properties were observed in the hornblende: $2V=35°$ approx.; extinction $Z \wedge c=18°$ approx.; pleochroism, X=yellow, Z=dark blue green; dispersion, $r<v$ strong.

Coarsely crystalline calcite in small amounts is found in nearly all the tactite. It fills interstices between garnet crystals, locally corrodes the crystal faces, and occupies zones in skeletal garnet crystals. Some calcite is intergrown with and partly replaces actinolitic amphibole, and, along with quartz, occupies veinlets that traverse the tactite. Its relations indicate that it was among the last minerals to form.

The occurrence of epidote is sporadic. It is not found in most of the tactite; however, it is abundant at some localities and minor amounts are found in veinlets in some specimens. Textural relations indicate the epidote crystallized late, essentially simultaneously with calcite.

Quartz, like calcite, is almost ubiquitous in the tactite; it fills veinlets and interstices, and replaces earlier formed minerals. It is clear and glassy, and much of it is fairly coarse grained. Rather commonly, the tactite next to the granodiorite contact is abundantly veined and extensively replaced by quartz over a zone a few feet wide. In these places the granodiorite, too, may be heavily impregnated with quartz. Relations with respect to calcite are not altogether clear, but the evidence indicates that, though some may be simultaneous with the introduction of calcite, the bulk of the quartz is later and, except for the sulfide minerals, is the last mineral to crystallize.

The tungsten ore mineral, scheelite, is an accessory constituent of low concentration in the tactite. Most commonly, it occurs as small discrete grains that range from 0.5 mm or less to a maximum of 1 or 2 mm across in the greatest dimension. Exceptionally, grains as large as 2 mm have been noticed in local concentrations. The relations of scheelite to other minerals of the tactite are discussed more fully in the section on mineral deposits (p. 84). The paragenesis affords evidence that scheelite was deposited during most of the episode in which tactite was being formed, and continued into the stage of late quartz deposition.

Traces of idocrase are found in some of the tactite. Its relations to the other minerals is uncertain, but it occupies late veinlets in some specimens.

The origin of tactite by replacement of marble and calc-silicate rocks is proved at many places by the field relations. Where it is found, it is always on or near the contact between granodiorite and the carbonate rocks; tongues and cross-cutting replacement bodies of tactite occur in the marble and calc-silicate rocks. It is difficult to assess the relative extent to which tactite has been formed by the replacement of marble rather than by replacement of light silicate rock. A some places the tactite and calc-silicate rocks are in contact, but original marble may have been completely replaced by tactite. At some places tactite obviously replaces marble. And examples of crosscutting replacement bodies of tactite in calc-silicate rock have been seen. Recrystallized but unsilicated marble may have been more susceptible to replacement by tactite because it was more permeable to the transforming solutions, whereas fine-grained calc-silicate rock would be less easily replaced, though one might expect that the platy, well-bedded structure in most of the calc-silicate rocks would probably afford suitable avenues for the introduction of the metasomatizing fluids. On the other hand, unsilicated marble would be more reactive with respect to the fluids than the calc-silicate rocks.

The consistent occurrence of tactite at or close to the granodiorite contact strongly suggests that temperature was an important controlling factor in the formation of tactite. It is to be expected that a steep temperature gradient outward from the granodiorite contact existed at the time of intrusion, and tactite formed where the temperature was highest—next to the granodiorite. The abrupt outer boundary of the tactite may also have been governed by the temperature distribution. Structural relations at the time of metamorphism were also important; that is, the rocks that were metamorphosed to tactite were situated so as to be available for reaction with the magmatic fluids when they were given off. These structural relations are discussed more fully in the section on tungsten deposits (p. 85).

The chemical changes involved in the conversion of limestone to tactite are exemplified in table 19, in which the compositions of marble and garnet-pyroxene tactite are compared, and in figure 9, which affords a visual comparison of the amounts of constituents gained and lost in the transformation. As might be inferred from the prevalence of garnet and clinopyroxene in the tactite, silica, iron, magnesia, and alumina have been added, and large amounts of carbon dioxide and considerable calcium oxide have been subtracted.

TABLE 19.—*Analyses of limestone and garnet-pyroxene tactite*

[Samples were analyzed by methods similar to those described in U.S. Geol. Survey Bull. 1036–C; P. L. D. Elmore, K. E. White, S. D. Botts, P. W. Scott, analysts, U.S. Geol. Survey]

	Limestone		Tactite
	1	2	3
SiO_2	5.8	2.4	42.2
Al_2O_3	.80	.34	4.1
Fe_2O_3	.43	.10	12.7
FeO	.01	.02	5.7
MgO	.26	.42	5.6
CaO	52.7	54.9	26.6
Na_2O	.07	.06	.17
K_2O	.12	.02	.06
TiO_2	.04	.02	.10
P_2O_5	.09	.10	.21
MnO	.00	.01	1.4
H_2O	.01	.02	.95
CO_2	40.1	42.6	.47
Sum	100	101	100
Sp gr (lump	2.58	2.34	3.36
Sp gr (powder)	2.71	2.70	3.52
Porosity (calc) percent	9.71	9.63	4.54

1. Limestone of Preble formation; NE¼ sec. 1, T. 37N., R. 41 E.
2. Limestone of Preble formation; NE¼ sec. 1, T. 37 N., R. 41 E.
3. Dark, layered garnet-pyroxene tactite, Pacific mine.

METAMORPHISM OF PALEOZOIC VOLCANIC ROCKS

Most of the volcanic rocks of pre-Tertiary age in the Osgood Mountains are beyond the metamorphic aureole that surrounds the granodiorite stock. In the northern part of the range, however, some of the altered volcanic rocks have been metamorphosed by intrusion of the granodiorite. The metamorphism has not been very intense: recrystallization has been slight or negligible and original textures and structures are well preserved; but new minerals have been formed which are indicative of a higher metamorphic grade than that attained by the rest of the altered volcanic rocks.

Volcanic rocks most affected by the contact metamorphism are part of the Farrel Canyon formation found in the low hills north of the quadrangle boundary. Some are exposed inside the quadrangle boundary in sec. 20, T. 39 N., R. 42 E. Megascopically, they look much the same as the other altered volcanic rocks—that is, dark greenish gray, hard, rough surfaced, and fine grained to microcrystalline. Original porphyritic textures with dull gray feldspar crystals can be recognized in some, and fragmental and amygdular structures are preserved in others. Seen under the microscope, however, they differ from the ordinary low-grade altered volcanic rocks in the absence or scarcity of chlorite and the presence of metamorphic pyroxene, actinolitic amphibole, and biotite. For example, in a fragmental volcanic rock of intermediate composition, porphyroblasts of colorless clinopyroxene have been formed in a microcrystalline granoblastic groundmass of plagioclase and pyroxene; a flow rock, probably also of intermediate composition, has amygdules of crystalline calcite bordered by pale-green clinopyroxene set in a fine-grained groundmass of small cloudy plagioclase laths, inter-

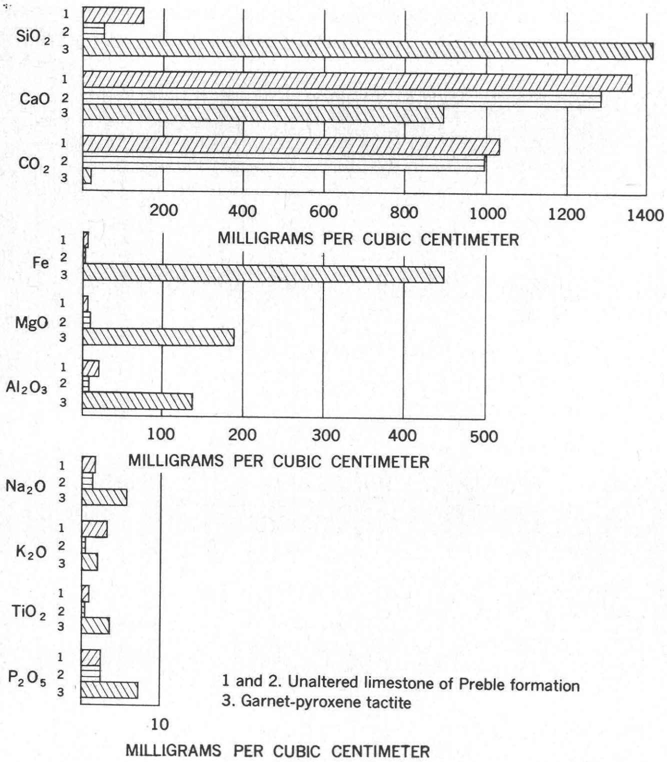


FIGURE 9.—Graph showing gain and loss of rock constituents in conversion of limestone to garnet-pyroxene tactite.

stitial fine-grained pyroxene and minor sphene, actinolitic amphibole, and clinozoisite. An example of a metamorphosed more silicic volcanic rock has a relict porphyritic structure consisting of phenocrysts of granular quartz and subhedral to euhedral phenocrysts of sodic feldspar in a very fine grained groundmass of abundant light-brown biotite and nearly unresolvable granular feldspar and quartz.

ENDOMORPHISM OF THE GRANODIORITE

The marginal facies of the granodiorite (p. 44) exhibits endomorphic effects where the granodiorite is in contact with marble, calc-silicate rocks, or tactite. The reaction zone is narrow, most commonly 6 inches to 2 feet in width, but may be as much as 10 feet or as little as a fraction of 1 inch. West of the Granite Creek mine, dikelike offshoots into limestone from the main body of granodiorite are entirely composed of granitic rock similar to that in reaction zones elsewhere.

Megascopically, the rock of the reaction zone is greenish and has a granitic texture. In some places an abrupt gradation from granitic rock into the wallrock is visible, but elsewhere the contact is sharp, though microscopic examination of specimens taken across a megascopically sharp contact shows gradation. The rock is characterized by diopsidic pyroxene as the principal mafic mineral instead of hornblende or biotite. Most specimens that were examined microscopically also contain some hornblende and rarely biotite intergrown with pyroxene in such a way as to show that some of the pyroxene was formed by replacement of these earlier crystallized minerals. The clinopyroxene has two main variations of habit: on the granodiorite "side" the crystals are equant to subequant and intergrown with plagioclase and quartz; toward the contact it commonly is in anhedral, more or less rounded granules which are poikilitically enclosed by feldspar and quartz. The amount of pyroxene is variable but equals and commonly is more abundant than hornblende and biotite in the normal granodiorite. Quartz, plagioclase, and orthoclase make up the balance and bulk of the rock. Plagioclase is in subhedral to euhedral crystals, shows prominent normal zoning, alteration to gray translucent kaolinitic(?) material, some sericite, and in some specimens contains small granules of clinozoisite. The unaltered crystals range from An_{40} to An_{50} (andesine); in some specimens the plagioclase has rims of oligoclase (An_{22}–An_{27}) that have corrosive boundaries against the more calcic cores, which in some specimens are altered in contrast to fresh rims. Orthoclase and quartz are anhedral and interstitial; orthoclase encloses, corrodes, and partly replaces plagioclase in some specimens. The amount of orthoclase is generally the same as in the normal granodiorite—approximately 10 to 12 percent—but in some specimens it is more plentiful; in the rock that forms dikelike projections west of the Granite Creek mine it amounts to about 35 percent. Sphene, commonly in diamond-shaped crystals, is the most plentiful minor accessory; apatite is ubiquitous but not plentiful; and only trace amounts of magnetite are visible.

Formation of the pyroxene-bearing granitic rock is attributed to enrichment of CaO in the granodiorite by mutual reaction of the solidifying magma with limestone at the contact, aided, perhaps, by limited physical incorporation of the wallrocks by the magma. The modification of the granodiorite in border zones adjacent to limestone is interpreted to be the result of reactions that took place as a consequence of the increase of CaO in the magma, which suppressed the formation of biotite and hornblende and caused diopsidic pyroxene to crystallize instead. Concomitantly, there was a loss of Al_2O_3 (Mg, Fe)O, and SiO_2 to the wallrocks, where garnet and clinopyroxene crystallized. It seems unlikely, however, that the great volume of tactite and light calc-silicate rocks in the contact zones were formed by transfer of material from only the very narrow selvage on the border of the granitic body. The increased amount of orthoclase in some parts of the reaction zone is possibly due to release of K_2O that normally would have been contained in biotite; TiO_2, freed by noncrystallization of biotite and hornblende, combined with the added lime to form sphene.

Similar reaction products have been formed on the border of the Boulder batholith in Montana, according to Knopf (1957, p. 97–98). Others, including Eskola (1914), Nockolds (1934), Joplin (1935) and Tilley (1949) have described pyroxene-bearing granitic rocks in selvages of normally biotitic and (or) hornblendic granitic rocks against carbonate rocks. All attribute the mineralogical changes in the granitic rock to the increase of lime in the granitic magma. Eskola (1914, p. 61) ascribed the crystallization of diopside "* * * to the fact that the amount of alumina was insufficient to form anorthite with the lime present * * *" and the CaO combined with FeO and MgO that normally would have formed hornblende, to form pyroxene. Joplin (1935, p. 110–112) regarded diopside as forming from the (Fe, Mg)O that normally would form biotite and hornblende, the released K_2O being free to form orthoclase in the contaminated rock, and the TiO_2 combining with CaO to form sphene. Tilley (1949, p. 88–92) showed that alumina was transferred from the granite at Skye, which led to the precipitation of an "* * * effectively non-aluminous member of the reaction series (pyroxene) * * *" in place of the aluminous types, biotite and hornblende.

STRUCTURAL GEOLOGY

The pages that follow include rather detailed descriptions of the folds and faults that have deformed and displaced the rocks of the Osgood Mountains and the Hot Springs Range. The chapter concludes with an interpretation of the structure and geologic history and their relation to the regional geology. The internal structures and structural relations of the granodiorite are excluded from the discussion, for they have been described in a preceding section, and minor structural features related to ore deposition will be set forth in the chapter on mineral deposits.

A glance at the geologic map (pl. 1) shows that the main structural trends in the Osgood Mountains and the Hot Springs Range are closely parallel to the trend of the ranges. In the central and south parts of the Osgood Mountains the bedding of the sedimentary rocks, axes of folds, and the major thrust and high-angle reverse faults have a predominant strike near N. 50° E. The strikes of the main structural elements in the north half of the range gradually become more northerly and are virtually north-south in the north end of the range. In the Hot Springs Range, where the main structural features are folds in the Harmony formation, the trends of the fold axes range from about north-south in the southern part of the range, to slightly east of north in the northwestern part. The granodiorite stock in the northeastern part of the Osgood Mountains is elongate in a north-northeasterly direction, the predominant structural trend in this part of the range. Within both ranges steeply dipping, mostly vertical, northwest-trending cross faults cut the rocks nearly perpendicular to the northeast structures. Locally, in both ranges, vertical or steeply dipping frontal faults parallel the ranges and are among the youngest structural features of the area.

FOLDS IN THE OSGOOD MOUNTAINS

Erosion of younger rocks has revealed a rather broad anticline in the Osgood Mountain quartzite and the overlying Preble formation in the central and south-central part of the range (pl. 1, sections *B-B′*, *C-C′*). The Preble formation is exposed in normal position on the west limb of the fold for a short distance near the mouth of Goughs Canyon; elsewhere the anticlinal structure is inferred from bedding attitudes in the Osgood Mountain quartzite. North of the northwest-striking fault that crosses the range between Goughs Canyon and Hogshead Canyon, the fold is concealed beneath younger thrust plates and unconformably overlying strata of Pennsylvanian age. Tertiary volcanic rocks conceal the south end of the anticline. The surface trace of the fold axis is sinuous and tends to be west of the range crest in the south; in its northern part it is straighter and lies more or less along the crest of the range. The axial plane is vertical or dips steeply westward.

On the limbs of the anticline, beds in the Osgood Mountain quartzite are markedly drag folded. The axial planes of the drag folds are steeply inclined east and west on the corresponding limbs of the major fold. The amplitude of most of the subsidiary folds is from a few feet to several tens of feet.

Throughout most of the Osgood Mountains, the Paleozoic sedimentary rocks younger than the Osgood Mountain quartzite are steeply tilted and tightly folded. Mostly, however, it is impossible to recognize and trace any well-defined persistent anticlines and synclines, and much of this folding is related to the thrust faults, which cut and displace most of the younger strata.

AGE OF THE FOLDING

The deformation responsible for the development of the major anticline took place prior to deposition of the Lower Pennsylvanian beds. The Osgood Mountain quartzite, the Preble formation, and possibly some younger strata, were folded; the Preble formation, and any younger beds that may have been present, were eroded from the uparched fold; and then strata of Pennsylvanian age were deposited on the Osgood Mountain quartzite. Evidence for a second episode of deformation in post-Pennsylvanian time that caused minor in-folding of the quartzite and Pennsylvanian limestone on the limbs of the anticline is found at a few places.

THRUST FAULTS IN THE OSGOOD MOUNTAINS

A thrust zone along the west side of the range is the dominant structural feature of the northern two-thirds of the Osgood Mountains. Here, rocks of Cambrian, Mississippian, Pennsylvanian, and Permian age have been thrust over one another in a pattern resembling a large imbricate thrust, although the rather poor exposures and inability to trace the thrusts into other areas make it difficult to substantiate such an interpretation. Thrust faults are also exposed along the crest and east side of the Osgood Mountains, but here they are less numerous and their origin is more easily interpreted. Thrust faults have not been recognized in the Hot Springs Range within the Osgood Mountains quadrangle, although one was mapped in the northern part of the range by Willden (in press) and the Valmy formation on the east side of the range is believed to have been brought into the area on a major thrust fault.

VALMY THRUST PLATE

The Valmy formation is very different lithologically from its partially temporal equivalent, the Comus formation, and must have been deposited under altogether different conditions. The Comus formation is believed to be autochthonous or parautochthonous, and we have concluded that the Valmy formation has been brought into the area in the upper plate of a thrust. This thrust fault is concealed in the Osgood Mountains quadrangle because of downfaulting, but probably it is equivalent to the Roberts Mountains thrust fault (Merriam and Anderson, 1942, p. 1693–1706), which is one of the important structural elements of north-central Nevada (Gilluly, 1954; Roberts and others, 1958, 2817–2820, 2850–2854). The Roberts Mountains fault is pre-Middle Pennsylvanian, probably Late Devonian or Early Mississippian (Roberts and others, 1958, p. 2852–2854).

TWIN CANYON FAULT

The Twin Canyon thrust is named for Twin Canyon on the east side of the Osgood Mountains, where it is well exposed. It crops out for about 4 miles between Hogshead Canyon at the north and Garden Spring (sec. 15, T. 37 N., R. 42 E.) at the south with an average strike of approximately N. 40° E. The fault dips southeastward and the angle becomes increasingly steep from south to north: north of Garden Spring the sinuous trace denotes a gentle dip; one-half mile south of Twin Canyon it dips approximately 35°; at Twin Canyon a dip of 65° was measured; and toward the north end of its outcrop it is nearly vertical.

The Twin Canyon thrust fault is properly designated a thrust only in the area south of Twin Canyon; north of Twin Canyon it is a high-angle reverse fault. The upper plate or hanging wall is Osgood Mountain quartzite, which is thrust over Preble formation for approximately 1½ miles north of Garden Spring and which overrides the Twin Canyon member of the Osgood Mountain quartzite for the rest of the distance to Hogshead Canyon (pl. 1, section *C–C′*). In the vicinity of Twin Canyon the rocks in the lower plate are folded into an asymmetric syncline whose axial plane dips southeast; beds of the southeast limb of the syncline, which are adjacent to the fault, are overturned to the northwest in places.

The displacement on the Twin Canyon is not accurately known, but a few thousand feet of movement would easily account for repetition of the Osgood Mountain quartzite over the Preble formation and the Twin Canyon member. The fault appears to have originated by fracturing along the attenuated common limb of an asymmetric syncline and anticline, with thrusting of the southwest limb of the anticline over the syncline caused by continued compressional deformation. The thrusting is older than Middle Pennsylvanian because the Battle formation has been deposited across the trace of the thrust in the southeastern part of sec. 3, T. 37 N., R. 41 E.

ADAM PEAK THRUST

The Adam Peak formation has been thrust over the Etchart limestone along the crest of the Osgood Mountains from the head of Perforate Canyon northward to the north side of the divide between Goughs Canyon and Hogshead Canyon. This thrust is named the Adam Peak thrust because the principal upper plate unit is the Adam Peak formation. The trace of the thrust can be only approximately delineated because it is concealed by thick overburden in most places. East of Goughs Canyon the trace indicates that the fault is nearly flat lying or dips gently northeast. The western part of the contact in the SW¼ sec. 23, T. 38 N., R. 41 E., seems to be about vertical and probably is a high-angle fault that has intersected the thrust. North of the divide between Goughs and Hogshead Canyons the Adam Peak thrust plate is overridden on the east by the Granite Creek thrust, and on the west by a thrust slice of Harmony formation (pl. 2, section *B–B′*). South of the divide, the Adam Peak thrust plate is bounded on the west by the trace of the Adam Peak thrust and on the east by a high-angle (probably normal) fault. This high-angle fault either marks the boundary between the Battle and Osgood Mountain formations or separates steeply dipping from gently dipping Battle.

The structure in the Adam Peak plate has not been worked out in detail, but generally the beds in the Adam Peak formation are steeply inclined, whereas the Etchart limestone, below the thrust, is nearly horizontal or gently inclined. Some small tight folds, whose axial planes dip steeply west, have been recognized at a few places, and overturned beds dip steeply east at several places along the east boundary of the plate, where it is overridden by the Preble formation on the Granite Creek thrust.

The likelihood of considerable horizontal displacement on the thrust is indicated by the approximately equivalent age of quite dissimilar units in the Adam Peak and Etchart formations. The Battle formation, however, underlies both formations and is to all appearances the same regardless of the overlying rock.

GRANITE CREEK THRUST

The Granite Creek thrust fault, so-named because it is well exposed in the head of Granite Creek (NE¼ sec. 25, T. 38 N., R. 41 E.), is a prominent feature on the geologic map and can be plainly seen in the field where it crosses the high part of the range north of Hogshead Canyon. In the northern part of sec. 25, T. 38 N., R. 41 E., contorted and generally steeply dipping beds of the Cambrian Preble formation have been thrust over gently northward-dipping strata of the Battle formation and Etchart limestone (fig. 10). Southward the thrust carries the Preble formation over the Osgood Mountain quartzite; northward it overrides the Adam Peak formation (pl. 2, section *B–B′*) and in sec. 12, T. 38 N., R. 41 E., it is overridden by the Harmony thrust plate.

The abrupt changes in direction of the trace of the thrust fault north of Hogshead Canyon (pl. 2) are caused by intersection of a warped fault surface with the uneven surface of the ground. The fault surface dips approximately 20° N. where it crosses the spur that projects south into sec. 25 from the main ridge. In the head of Granite Creek the fault surface has an anticlinal warp that plunges northward, carrying the fault trace around through 180°; south from here the fault dips moderately to steeply eastward. Where the fault rises over the Adam Peak formation and crosses the main ridge it becomes steeper and, going through a right-angle turn, strikes northward and dips eastward approximately 35°. In the head of East Fork of Eden Creek another warp gives the fault surface a gentler dip, and carries the Preble across the Adam Peak formation.

The edge of the fault plane is exposed in the head of Granite Creek and at a few places on the west side of the Adam Peak ridge. Wherever the contact can be seen it is sharp and tight, but with careful examination a thin seam of crushed rock can usually be found, and at some places the bedding in the rocks above the fault is considerably contorted. No marked brecciation of rocks above or below the fault has been recognized.

PEAK 6837 THRUST

Along the crest of the Osgood Mountains near the north boundary of the quadrangle, a sequence of strongly deformed phyllite, hornfels, and thin-bedded recrystallized limestone of the Comus formation is thrust over the Preble formation. From its surface trace, it is estimated that the thrust dips approximately 10° to 15° NE. Near the northern boundary of the quadrangle the plate is overridden by the Farrel Canyon formation and terminated by normal faults.

In detail, the rocks of the Comus formation on the upper plate are rather strongly deformed, particularly the thin-bedded limestones which commonly are intricately crumpled; exposures are not abundant enough to permit the structure to be fully worked out, but a synclinal fold plunging northeast is indicated by the plotted observations of bedding attitudes. The syncline seems to continue southwestward in the Preble formation. Fine-grained intrusive rocks are more plentiful in the sequence of rocks above the contact, and in places sheetlike intrusive bodies occupy the contact or occur in the upper plate close to the contact, indicating that the contact was a surface along which magma found relatively easy access.

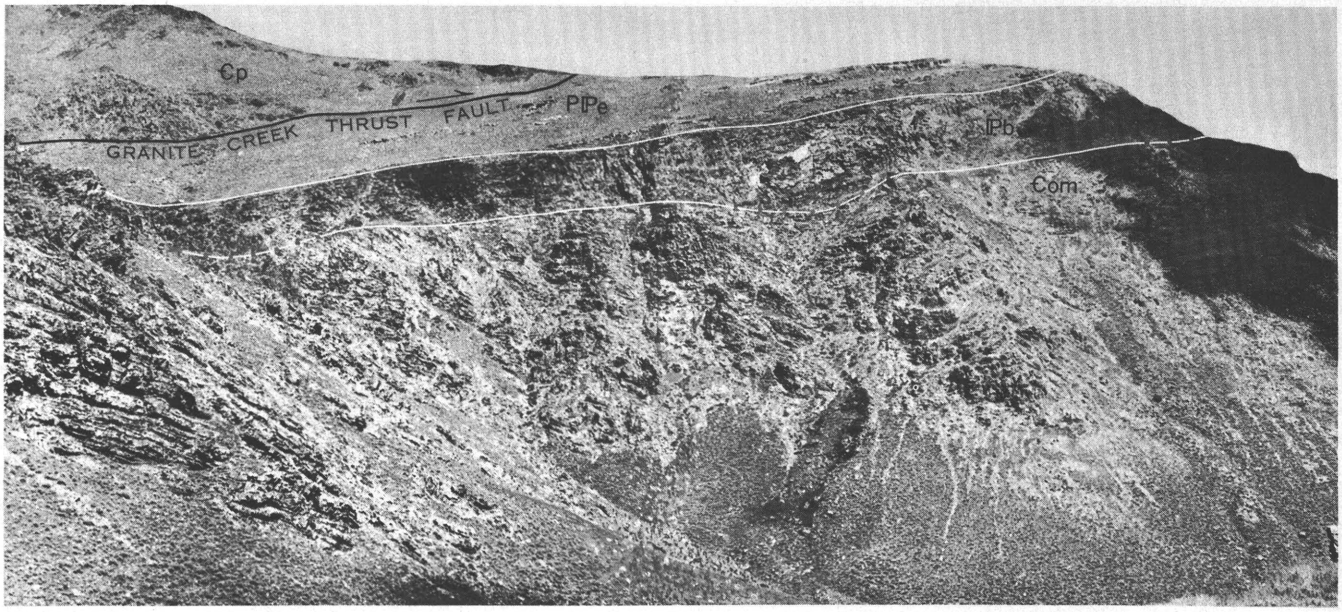
IMBRICATE THRUST ZONE IN OSGOOD MOUNTAINS

A complex pattern of overlapping thrust plates is exposed along the west side of the Osgood Mountains. Individual plates are continuous north of Goughs Canyon, but to the south they are represented by generally isolated klippen of various sizes. The thrusts are approximately parallel: their traces trend northeast to north, and they dip gently to moderately west. The general parallelism of the faults and their pattern of distribution suggests that they may have moved essentially simultaneously.

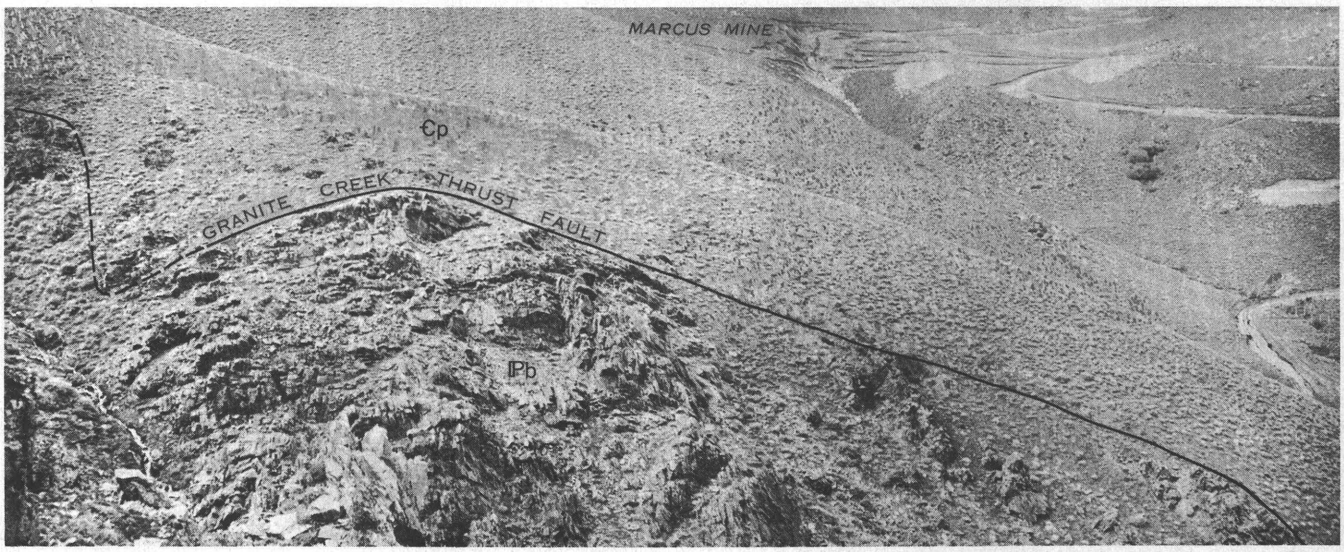
These faults in their approximate order of superposition from bottom to top are: the Preble, Etchart, Harmony, Goughs Canyon, and Farrel Canyon thrusts. All these faults are named after the dominant rock unit on their upper plate.

PREBLE THRUST PLATE

The Preble thrust is the lowermost of these thrusts. Its upper plate consists of exceedingly contorted phyllitic shale provisionally classified as Preble formation of Cambrian age. The thrust has overridden the Etchart limestone and, at a few places, the Osgood Mountain quartzite. The thrust plane locally is somewhat contorted, but in general it strikes northeast and dips gently northwest at an estimated 15° to 20°. The relatively thin thrust plate is estimated to be between 100 and 300 feet thick where it intersects the surface. The Preble thrust plate has formed a sole for some of the higher thrusts in Goughs Canyon.



A. North side Hogshead Canyon. Steeply inclined rocks of the Preble formation thrust over gently dipping strata of the Etchart limestone, which are underlain by conglomerate of the Battle formation and the Osgood Mountain quartzite. (Ꞓom, Osgood Mountain quartzite; Ꞓp, Preble formation; IPb, Battle formation; PIPe, Etchart limestone.)



B. Granite Creek. Rocks of the Preble formation thrust over conglomerate of the Battle formation. (Ꞓp, Preble formation; IPb. Battle formation.)

FIGURE 10.—VIEWS OF THE GRANITE CREEK THRUST FAULT.

ETCHART CANYON THRUST PLATE

The Etchart limestone has been thrust over several rock units in different parts of the Osgood Mountains. For convenience of description, all these overthrust blocks of Etchart, and some klippen of Battle formation, are collectively referred to as the Etchart thrust plate; although there is evidence that suggests this thrusting has occurred at different times.

Several klippen of Battle formation and Etchart limestone are exposed in the upper part of Goughs Canyon. Most of these rest on the Etchart limestone, but there is a thin sole of Preble formation intervening between some of the klippen and the underlying Etchart. These relations are particularly well exposed on a small hill in NE¼ sec. 27, T. 38 N., R. 41 E., where the Etchart limestone overrode a thin plate of Preble formation. The edge of the thrust plane is clearly exposed dipping northwest about 15°. It gradually cuts across strata of the upper plate, which are inclined somewhat more steeply in the same direction as the thrust fault. Crumpling in the underlying shale next to the contact gives evidence that the thrust moved from northwest toward the southeast. A klippe of Etchart limestone rests on the Adam Peak formation on the ridge west of Adam Peak and two klippen rest partly on the Adam Peak across the trace of the Adam Peak thrust on the north side of Goughs Canyon.

The Etchart limestone has been thrust on the Farrel Canyon formation on the west side of the range north of Anderson Canyon, and on the Valmy formation on the east side of the range northeast of the Getchell mine. On the west side of the range the Etchart thrust plate rests on the Farrel Canyon formation; whereas, on the east side of the range it has been overridden by the Farrel Canyon, presumably on the Farrel Canyon thrust. This relation can be used as evidence that the Etchart limestone has been involved in more than one period of thrusting, but we believe the Etchart limestone was first thrust into the area, coming to rest on the several different formations, and then the Farrel Canyon thrust broke across the earlier Etchart thrust plate, carrying part of the Etchart plate with it.

HARMONY THRUST PLATE

Next highest in this succession is the Harmony thrust plate, which has carried the Harmony formation of Late Cambrian age over the Adam Peak formation, autochthonous Etchart limestone, and thrust remnants of the Etchart limestone and the Battle formation. The fault strikes N. 40° E., dips northwest, and crops out for more than 3 miles northeast of Goughs Canyon. The dip is about 25° NW. where the fault trace crosses the ridge north of Goughs Canyon in SW¼ sec. 23, T. 38 N., R. 41 E.; northward, the relation of the fault trace to topography indicates that the dip is in the same direction but somewhat steeper. The strata, lithologically like the type Harmony—feldspathic sandstone and grit, shale, and limestone—are highly deformed throughout all of the thrust sheet, and no pattern of folds can be resolved. The strike of beds is northeast and rather constant, but dips are highly variable. Where exposures are good the exceedingly deformed condition of the rocks can clearly be seen. The shale has been so squeezed that it appears crumpled and sheared and weathers out into small polygonal fragments which in places are pencil like. Beds of sandstone and grit, and less commonly limestone, commonly a foot to several feet thick, are interbedded with the shale, in which they have been rolled and twisted as if in a plastic mass. Many of the sandstone beds show a kind of structure that resembles "boudinage" (Cloos, 1947). The beds are broken across the bedding, and the relatively softer and more plastic shale has flowed into the fractures, separating the original beds into an alined series of disconnected blocks. Many of the sandstone "boudins" are separated from their original position and are seen as hard boulders embedded in a shale matrix.

GOUGHS CANYON THRUST PLATE

The plate of Harmony formation rocks is overridden by a still higher thrust sheet that carries the Goughs Canyon formation of Mississippian age. At its south end, in and near Goughs Canyon, the thrust rests directly on thin thrust sheets of the Etchart limestone and the Preble formation, and on autochthonous beds of the Etchart limestone. A small klippe of the Goughs Canyon formation laps over the Preble sheet on to the Osgood Mountain quartzite just north of center sec. 33, T. 38 N., R. 41 E., in the head of Perforate Canyon. North of Goughs Canyon, the thrust gradually overlaps the plate of Harmony formation rocks and at its north end brings the Goughs Canyon formation into contact with the Preble formation. Westward the plate is covered by volcanic rocks of Tertiary age, and northward it disappears beneath a thrust plate of Farrel Canyon formation. The trace of the thrust fault trends northeast in its southern two-thirds, and very slightly east of north at its northern end. On the south slopes of Goughs Canyon the pattern of the fault trace is evidence that the fault plane dips very gently north or northwest; north of Goughs Canyon the trace is relatively straight, indicating a steeper dip estimated by graphic methods as about 25°. The fault is fairly well exposed on the north side of the East Fork of Eden Creek in north center sec. 14, T. 38 N., R. 41 E., where it dips approximately 40° NW.

The beds on the upper plate persistently strike northeast and dip steeply west in the southern part of the thrust plate; north of the East Fork of Eden Creek the strike is nearly north-south and dips are variable, but generally steep. In the northern part of the thrust plate some tight folds were seen, but they were not mapped in detail; in the southern part of the plate poor exposures limit the observation of folding. The structures strike north and beds near the thrust fault are overturned to the west, but graphic reconstruction of folds suggests that farther from the thrust fault the anticlines and synclines have axial planes that dip steeply west.

FARREL CANYON THRUST

The Farrel Canyon thrust is the uppermost thrust in the zone of imbrication on the west side of the Osgood Mountains. The thrust is exposed from the head of Cherry Canyon to the north side of Farrel Canyon, where it overrides the Goughs Canyon formation and a small part of the Preble formation. The thrust is also exposed in the extreme northeast corner of the quadrangle where it overrides the Comus formation and the Etchart limestone. On the west side of the range the Farrel Canyon thrust is cut off by the high-angle Anderson Canyon fault; on the northeast side of the range the thrust terminates against the Village fault, also a high-angle fault. North of Farrel Canyon the thrust is nearly flat; on the south side of the canyon it dips approximately 75° W., and farther south the pattern of the trace of the thrust suggests that the dip ranges from steep to gentle. The rather sinuous trace of the thrust in the northeast corner of the quadrangle indicates that the thrust dips gently north.

Strata in the thrust plate on the west side of the range are closely folded into a series of anticlines and synclines that strike mostly north; elsewhere, the Farrel Canyon formation is too poorly exposed to give any indication of the structure, but in most outcrops the beds are steeply inclined.

HIGH-ANGLE FAULTS IN THE OSGOOD MOUNTAINS

The rocks of the Osgood Mountains are broken by many steeply dipping or vertical faults which have two principal orientations, one northeast to north and the other northwest. The northeast faults are, for the most part, closely parallel to the trend of the formations and strike of the rocks and may be classed as longitudinal or strike faults. These are more continuous and more important structurally than the other set. The northwest-striking faults are classed as cross faults because their orientation is generally normal to the predominant structural trend. There is direct evidence that many of the faults are later than the thrusts; there is no evidence that any preceded the thrusting. Age relations between the high-angle faults, however, are not completely clear. At least one on the east side of the range near the Getchell mine is later than the granodiorite stock; there is no clear evidence that the others preceded or followed emplacement of the granodiorite. Some on the east side of the range probably should be regarded as range-front faults related to the basin-and-range system. The more prominent high-angle faults will be discussed in the following pages.

ANDERSON CANYON FAULT

The Anderson Canyon fault, named for a prominent canyon in the northwestern part of the Osgood Mountains, can be traced from Cave Canyon to the north boundary of the quadrangle. Phyllites of the Preble formation on the east side of the fault have been brought against the Etchart limestone at the north end of the range, the Farrel Canyon formation in Anderson Canyon, and limestone of the Preble formation in Farrel and Cave Canyons. The fault surface is not exposed, but the dip is known to be very steep or vertical from the straight trace of the fault. The relative displacement was down on the west side. The total displacement is not known, but a minimum displacement of 1000 feet can be postulated on the supposition that the Etchart thrust plate once extended across the range.

VILLAGE FAULT

The Village fault is on the east side of the Osgood Mountains in the northeast corner of the quadrangle; it is so named because it passes through the northeastern part of the village at the Getchell mine. In some ways it seems to be the eastern counterpart of the Anderson Canyon fault; it has a northerly trend on the east flank of the range and brings older rocks on the range side into contact with younger formations on the basin side.

The Village fault can be followed southward from the north boundary of the quadrangle for nearly two miles; south of the village it is covered by alluvium. The fault strikes approximately N. 20° W. with only one minor offset within the quadrangle, and presumably, from the way it maintains its straightness regardless of topography, the fault is vertical.

The apparent displacement is up on the west or range side of the fault and down on the east side. Thus, the older Preble formation and the metamorphosed Comus formation above it on the Peak 6837 thrust plate west of the fault are brought into juxtaposition with younger rocks east of the fault, including the Etchart limestone, the underlying Valmy formation and the overthrust Farrel Canyon formation. Here again, the total displacement is not known, but a minimum displacement of 1,000 feet can be postulated on the assumed offset of the Etchart thrust plate.

GETCHELL FAULT

A more or less continuous zone of faulting extends along the east side of the Osgood Mountains near their base from about 1 mile north of Getchell mine to the mouth of Granite Creek, a distance of about 6 miles. The fault takes its name from the Getchell mine, which is on the fault zone. The Ogee and Pinson fault south of Granite Creek possibly is a southwestward continuation of the Getchell fault zone.

Although the fault is mainly along the east front of the range, at most places it is not the boundary between bedrock and the alluvium-filled basin to the east. Bedrock is exposed in low hills for more than a mile east of the fault in some places. Throughout its length the fault, or fault zone, is generally parallel to the strike of the sedimentary rocks, though in detail at many places it transects the bedding at a small angle. Furthermore, where the sedimentary rocks tend to strike parallel to the eastern contact of the granodiorite stock, the fault, too, clearly bends eastward and outward around the bulging northern lobe of granodiorite, and less so east of the southern lobe. At several places where mining operations have exposed the fault (fig. 11), it dips east at moderate to steep angles; at the Getchell mine the fault dips 30° to 40° E. in its northern exposures, and 70° to 80° E. in the southern mine pits; farther south at the Riley mine the footwall break dips 49° E.; a parallel fault segment at the Pacific mine dips 45° E. Mining operations have also

FIGURE 11.—View of the Getchell fault, Getchell mine. Looking south along ore zone in south pit. Footwall of Getchell fault, dipping east, on right.

shown that a zone of fracturing and brecciation as much as 500 feet wide lies above the generally well defined footwall break, and many subsidiary faults split off the main zone into the hanging wall and footwall.

The Getchell fault is a zone of overlapping fractures. North of the Riley mine the rocks on both sides of the fault are shown on the map as Preble formation except near the mouth of Rocky Canyon, where granodiorite is on the footwall, and at the Getchell mine, where granodiorite is on the hanging wall. South of the Riley mine a fault separates rocks of the Preble formation from those of the Comus formation. Absence of exposures obscures the relations between the fault at the Riley mine and the fault to the south between the Preble and Comus formations; but it appears that, though the general zone of faulting may be continuous, the fault contact between Preble and Comus splits off from the main fault and possibly joins the Village fault somewhere between the Riley mine and the low hills of Comus formation to the east of the quadrangle. East of the Valley View mine the fault between Comus and Preble formations disappears southeastward beneath alluvium; but a strong parallel fault with a wide breccia zone lies one-fourth of a mile to the west within the Preble formation and continues southward for a mile until it, too, is covered by alluvium.

Near the mouth of Rocky Canyon, a mile south of the Getchell mine, the Getchell fault intersects the granodiorite contact, and for almost half a mile northward granodiorite forms the footwall. In the main mine pit south of the mill (South Pit) at the Getchell mine, granodiorite is exposed in the hanging wall next to the fault zone. It is also exposed east of the fault in an excavation for the primary crusher, and in a bank and on the playground of the school north of the mill.

In the South Pit at the Getchell mine the footwall of the fault is well exposed. The footwall surface, highly polished in places, dips moderately to steeply east, and on it are prominent grooves or mullions that are horizontal or plunge northward at angles of a few degrees. At set of striae and slickensides caused by late movement cut across the nearly horizontal grooves at about 90°, and thus pitch down the dip of the footwall plane. Here, then, is evidence of strike-slip movement along the fault, succeeded by chiefly dip-slip movement. No stratigraphic data are available for determining the amount of movement along the fault, but indirectly the evidence suggests that the granodiorite in the hanging wall is a segment of the main body of grandiorite cut by the fault and displaced northward about 3,500 feet by a predominantly strike-slip left-lateral movement. The occurrence of bedrock east of the fault and relations of formations on opposite sides of the fault, the presence of grandiorite east of the fault, and the absence of topographic expressions such as well-developed scarps and triangular facets along the range front—all indicate that the amount of late dip-slip movement was relatively small.

Clearly, the Getchell fault zone is later than the granodiorite, and earlier than the episode of mineral deposition that formed the gold-arsenic deposits at the Getchell mine, which may have been in Tertiary time. Evidence of late normal displacement on the fault—in Quaternary or perhaps Recent time—is shown by the alluvium, which is as much as 25 feet thick a few hundred feet east of the fault and only a few inches to 1 or 2 feet thick on the slope west of the fault. In a cut where Burma Road crosses the footwall fault, alluvium 5 to 10 feet thick on the east side ends abruptly at the fault, and west of the fault the overburden is but a foot or so thick.

OGEE AND PINSON FAULT

The Ogee and Pinson fault takes its name from a small open-pit gold mine near the top of a ridge half a mile southwest of the mouth of Granite Creek. The fault is shown on the geologic map (pl. 1) as the contact between the Preble and Comus formations south of Granite Creek. Actually, clear evidence of a fault is seen only at the Ogee and Pinson mine, where a wide zone of brecciated hornfels and chert is exposed. To the southwest, from Felix Canyon to southwest of the mouth of Hogshead Canyon, the contact between the Preble and Comus is concealed by soil and alluvium. A fault contact is postulated in the concealed area because of evidence of faulting to the north and the absence of higher Cambrian strata betwen the Preble and Comus formations; however, the original distribution of the Paradise Valley chert and the Harmony formation is poorly known, and they may never have been deposited here or may have been removed by erosion prior to deposition of the Comus. In the Edna Mountains, 12 miles to the south in the Golconda quadrangle, the Preble-Comus contact is a high-angle fault (p. 11).

At least the northern segment of the Ogee and Pinson fault may be an extension or offshoot of the Getchell fault zone. It is on the general trend of the Getchell fault zone, has a wide brecciated zone at the Ogee and Pinson mine, and contains a low-grade gold deposit.

OTHER HIGH-ANGLE FAULTS

Many other steep faults have been mapped in the Osgood Mountains. These faults and their effects are obvious from the geologic map (pl. 1), so individual faults are not described in detail.

Northwest-striking faults in the central part of the Osgood Mountains are most conspicuous on the geologic map, for they trend chiefly at right angles to the northeast-trending structures and have caused many interruptions and offsets of the strata. Some of the faults can be followed for as much as 1½ miles, but most are less than 500 feet long. Although the dip of the faults cannot be measured in most places because of inadequate exposures, the relation of their traces to the topography is evidence that they are vertical or very steeply inclined. Normal dip-slip movement accounts for the apparent displacement of units relative to one another on opposite sides of the faults—a displacement indicating total movement of the faults of only a few hundred to, in one place, a thousand feet.

Faults that strike slightly east of north, mostly diagonal to the regional structure, terminate some of the northwest faults. These, too, are steep to vertical faults, and the amount of displacement along them is similar to that on the northwest faults.

A few other steep minor faults have been mapped elsewhere in the Osgood Mountains, and most of them have either a northwest or north strike. Most of the steep minor faults obviously were formed later than deposition of the Pennsylvanian rocks, after the thrust faulting and later than the folding.

FOLDS IN THE HOT SPRINGS RANGE

Deformation chiefly by folding characterizes the comparatively simple structural unit of the Hot Springs Range in contrast to the complexly faulted Osgood Mountains. Inspection of the geologic map (pl. 1) shows many recorded measurements of bedding attitudes in the Harmony formation, which range widely in direction of strike and dip and are an indication of many intricate folds. Characteristically, the Harmony formation lacks persistent, distinctive lithologic units that can be easily followed and mapped, and by which the folds could be clearly delineated; therefore, the true pattern of folds in the Hot Springs Range is not known. However, some folds can be recognized in the field from bedding attitudes and outcrop patterns, and by mapping some of the interbedded limestone members. In general, the structure is a series of north-plunging asymmetrical anticlines and synclines whose axial planes dip moderately east; some of the folds are isoclinal (beds on opposite limbs dip at equal angles in the same direction). On the northwest flank of the range, in the northwest corner of the quadrangle, the Paradise Valley chert, which underlies the Harmony formation, is exposed in two anticlines (pl. 1, section *A-A′*) that probably are chiefly isoclinal.

By direct stratigraphic or structural relations the folding of the upper Cambrian rocks in the Hot Springs Range can only be determined to be later than Cambrian and earlier than Tertiary in age, for the oldest rocks that overlie the folded strata are volcanic rocks of Tertiary age. The small granodiorite bodies that intrude the folded rocks are of the same composition as the stock in the Osgood Mountains, which from age determinations by the zircon method is estimated to be Late Cretaceous in age. Probably the folding is, at the latest, earlier than late Cretaceous.

FAULTS IN THE HOT SPRINGS RANGE

Undoubtedly there are more faults in the Hot Springs Range than are shown on the geologic map. Evidence of faulting is common in many places, but without distinctive lithologic markers, most of the faults could not be mapped.

All the faults shown in the Hot Springs Range are vertical or high-angle normal faults. The amount of movement on most of the faults need not have exceeded a few hundred feet to account for the relations between the rocks involved. With a few exceptions most of the faults shown can be grouped into two sets: those which strike north to northeast and those which strike northwest.

North- and northeast-striking faults on the east side of the range have downfaulted, and therefore preserved from erosion, blocks of Tertiary volcanic rocks that formerly overlay the Harmony formation in the range. Similarly, near Box Canyon midway along the east side of the range, a downfaulted block of Valmy formation has been preserved from erosion. Probably the small patch of Valmy at Stone Corral is similarly preserved, although alluvium hides the contact with the Harmony formation. North- or northeast-striking faults are less prominent elsewhere in the range, except in the northwest corner of the quadrangle where northeast faults at the range front probably are responsible for repetition of the Paradise Valley chert.

The major faults along the east front of the range all have the same kind of displacement, with the west or mountain side of the faults displaced upward relative to the east or valley side. In the vicinity of Box Canyon the block of Valmy formation rocks between the parallel faults is down relative to the western or range block but is up relative to the eastern block that preserves the volcanic rocks. Farther south, a narrow block of Harmony formation rocks with some remnants of volcanic rock has been lowered relative to the northwest side of the fault, and the block of volcanic rocks to the southeast has been further downfaulted. This step faulting is probably responsible for the major uplift of the range.

Most prominent of the northwest-striking faults are those on the northwest side of the range that are chiefly at right angles to the folding and intersect and displace a belt of the Paradise Valley chert. Their southeastward extensions are recognizable in the field by offsets in thin limestone members of the Harmony formation.

Many of the faults shown in the Hot Springs Range are certainly of Tertiary age or younger, for they transect the volcanic rocks. For some, such as the two prominent northwest faults in the northern part, there is no direct evidence of their age except that they are later than folds in the Cambrian rocks. Some faults of each set are known to cut volcanic rocks; so it seems probable that both sets are of about the same age.

Some quicksilver deposits of minor importance are associated with faults whose relation to the faults shown on the map is not clearly known. Not shown are some small gold-bearing quartz veins in and surrounding the stock of granodiorite at Dutch Flat, which presumably are related to the intrusive body and were deposited in fractures formed during or soon after emplacement of the granodiorite.

RELATION BETWEEN THE HOT SPRINGS RANGE AND THE OSGOOD MOUNTAINS

An alluvium-filled valley separates the Osgood Mountains from the Hot Springs Range, concealing the boundary between the prevolcanic rocks of the two ranges. Interpretation of the concealed relations is difficult because the Paleozoic rocks of the two ranges are not alike, except for a thrust silver of the Harmony formation in the Osgood Mountains, and there is no clear structural continuity between the ranges.

The Osgood Mountain quartzite and Preble formation beneath the thrust plates of Preble formation are interpreted to be autochthonous (for the most part, in the place where they were deposited). Shales of the Preble formation overlie the quartzite on the northwest flank of the fold near the mouth of Goughs Canyon. Here, then, is evidence that Preble formation may underlie the alluvium and volcanic rocks between the Hot Springs Range and the Osgood Mountains south of the Dry Hills. In the Hot Springs Range the Harmony formation rests depositionally on the Paradise Valley chert, which is younger than the Preble formation; but the nearest exposures of Paradise Valley and Preble formation are 8 miles apart, so their stratigraphic relations are entirely conjectural. The Harmony and Paradise Valley in the Hot Springs Range may be either autochthonous or allochthonous (brought from elsewhere to their present position). In the Sonoma Range (Ferguson, Roberts, and Muller, 1952) the Harmony formation occurs on thrust sheets and also in apparently autochthonous position beneath thrusts; at Battle Mountain (Roberts, 1951) and in the Mount Lewis quadrangle (James Gilluly, oral communication) the Harmony formation is thrust over Lower Cambrian and Ordovician rocks.

Lengthy speculation on possible structural situations that would explain the present relations between the two ranges is fruitless, but two possibilities that seem plausible should be mentioned briefly: (1) the Harmony formation and underlying Paradise Valley chert in the Hot Springs Range are in an autochthonous block on the west side of a major anticline or in a generally synclinal structure, modified by normal faulting along the flanks of the range; or (2) the Hot Springs Range is allochthonous and is on the upper plate of a thrust fault whose outcrop is concealed by alluvium in the valley. Evidence for the first possibility includes the ample room between the two ranges for accommodation of the postulated structure, and the absence of a thrust beneath the Harmony formation or Paradise Valley chert in the Hot Springs Range. Evidence for the second possibility is the position of the Harmony formation in thrust sheets at several places in north-central Nevada including the west side of the Osgood Mountains north of Goughs Canyon, and the lack of exposures in the Osgood Mountains of either the Paradise Valley chert or the Harmony formation in normal position.

Clearly the present structural relations were established prior to the Tertiary erosion and subsequent volcanic activity. Much if not all the Hot Springs Range and at least the southern part and western slope of the Osgood Mountains were blanketed by volcanic rocks that were laid down on an uneven erosion surface cut into rocks that had the same general relationships that they have now. The northern, higher part of the Osgood Mountains may not have been covered by the volcanic flows, for it may have stood as a highland during the time of Tertiary volcanic activity. The flows in the Dry Hills and at Soldier Cap on the west side of the range may lie on the east side of an old valley that was established prior to the episode of volcanism, as suggested by the rather steeply inclined basal contacts that truncate the nearly flat or gently tilted flows and the rapid thickening of the flows westward.

ORIGIN OF THE RANGES

Evidence bearing on the origin of the present form of the Osgood Mountains, the Hot Springs Range, and the bordering alluvium-covered valleys is largely circumstantial. Certain facts suggest, but do not prove, that faulting has been mainly responsible for delimiting the mountain blocks from the basins, but that some of the relief resulted from uplift by slight arching, possibly attended by step faulting.

Separation of bedrock in the mountains from alluvium in the basins by normal faults along the front of the ranges cannot be conclusively demonstrated. Nevertheless, the straightness of a mountain front—particularly if it cuts across the strike of the rocks—is commonly regarded as evidence of a range-bounding fault. Attention is drawn therefore to the east front of the Osgood Mountains, which is fairly straight from about the latitude of Lone Butte to Getchell mine, and more so between the mouth of Granite Creek and Getchell mine. The front is slightly oblique to the strike of the rocks in the range. Similarly, the east and west fronts of the Hot Springs Range are quite straight.

High-angle faults along the east front of the Osgood Mountains are generally parallel to the range, but they seem to be structural elements of the range rather than the boundary between the range and the alluviated valley to the east, for low hills and pediment slopes lie between the range proper and the valley alluvium. Weakly developed facets on some of the spurs along the northern part of the east front of the range and steeply dipping striae on the Getchell fault are evidence of some uplift of the range along this fault, but other data cited previously (p. 75) indicate that perhaps the major movement on the fault was lateral. Other faults east of the present range front may have been responsible for the uplift, and the range front retreated by erosion to its present position. Some evidence supporting this hypothesis is provided by a northeast-striking fault that cuts through alluvium east of the range front in the southeast part of the quadrangle. Another possibility is that the range is bordered by step faults, and the foothills lying east of the Getchell and the Ogee and Pinson faults are part of a steplike block between the higher part of the range and the basin.

The west front of the Osgood Mountains is very irregular, and presents no compelling evidence for a range-front fault. The west boundary of the Osgood Mountains block may, however, be more or less along the east side of the Hot Springs Range, where normal faults form the boundary between volcanic rocks on the valley side and older rocks to the west in the range. The flows on the east side of the Hot Springs Range and those on the west side of the Osgood Mountains are less than a mile apart in places and probably are continuous beneath the valley alluvium.

Some relative upward movement of the Hot Springs Range must be postulated to account for the almost complete removal of volcanic rocks in the range in contrast to their preservation along the east foot of the range. The straightness of the front on both sides of the range is suggestive of a range bounded by faults. Some faults are known on both sides of the range: on the east side, relative upward movement of the range undoubtedly has taken place along the faults between volcanic rocks at the foot of the range and the older rocks, and extensions of these faults may be covered by alluvium; a normal fault probably occurs along the range front in the northwest corner of the quadrangle. Elsewhere, alluvium laps up on the range in many places without an abrupt change in slope, and the ridges are not faceted along the front of the range. However, the possibility that part of the uplift of the Hot Springs Range was accomplished by arching along the axis of the range, accompanied perhaps by step faulting, must also be considered. The flow remnants dip gently east and west away from the range crest toward the valleys, and many are bordered on one side by a normal fault. It can be seen from sections *A–A′* and *B–B′*, plate 1, that the structural relief of the range as shown by the gently dipping flow remnants could have been satisfied by gentle arching. Similar evidence for origin of some of the ranges by upwarping has been found by Willden (in press) in Humboldt County. In the Santa Rosa Range volcanic rocks close around the north end of the range, with west dips on the west and east dips on the east side; in the Black Rock Range, also, volcanic rocks have gentle fold structures.

INTERPRETATION OF THE STRUCTURAL RELATIONS AND GEOLOGIC HISTORY

The structural development of an area can be clearly understood only as part of the geologic history. Therefore, the following discussion, which attempts to interpret the structural position of the Osgood Mountains and Hot Springs Range in a regional framework, includes an interpretation of the geologic history founded on known structural and stratigraphic relations in this and neighboring areas.

PALEOZOIC HISTORY

In an earlier report (Roberts and others, 1958), which summarized the Paleozoic stratigraphy and structural history of north-central Nevada, it is shown that most strata of early and middle Paleozoic age can be assigned to two main assemblages of contrasting facies, regional in extent, and that the facies have been brought together by thrust faults. In eastern Nevada the rocks are a carbonate assemblage that is composed mostly of limestone and dolomite but includes minor amounts of shale and quarzite. They were deposited in a shallow-water miogeosynclinal environment that covered much of western Utah and Eastern Nevada. West of long 116° to 117° and north of lat 39° in central and western Nevada, rocks of the same age are a siliceous assemblage composed dominantly of clastic sediments and chert, with intercalated volcanic and pyroclastic rocks. These rocks were deposited in a eugeosynclinal environment in western Nevada west of the shallow shelf seas.

In the Osgood Mountains quadrangle, Cambrian rocks younger than the Osgood Mountain quartzite and the Comus formation of Ordovician age are characterized lithologically by a combination of siliceous, clastic, volcanic, and carbonate elements; that is, the lithology is neither distinctly a siliceous and volcanic nor a carbonate assemblage, but intermediate. Hence, the concept has developed that these rocks belong to a third, or transitional, assemblage. Taken individually, some transitional units resemble the carbonate assemblage and others the siliceous; but as a whole, a transitional siliceous-carbonate assemblage is recognizable. Boundaries between the assemblages probably oscillated back and forth across central Nevada, and locally deposits of one assemblage may have accumulated in a part of a basin characterized predominantly by another assemblage.

The three assemblages were laid down in a broad north-trending geosyncline in north-central Nevada that persisted until the end of the Devonian (Nolan, 1928, p. 158), when orogenic movements resulted in the emergence of a positive area that trended north-northeast medially across the state. This positive area, which was the locus of intense folding and faulting during the Antler orogeny in the late Paleozoic, has been called the Antler orogenic belt (Roberts, 1949, p. 95; Roberts and others, 1958, p. 2851).

CAMBRIAN SEDIMENTATION

The geologic record in the Osgood Mountains quadrangle begins probably in Early Cambrian time with the deposition of clean sand in an advancing sea that extended over much of the present Great Basin. This sand, representing washed and reworked material derived from a low-lying deeply weathered landmass, accumulated to an unknown thickness probably amounting to several thousand feet and became the Osgood Mountain quartzite in this area. Gradually the sediment being deposited changed from clean sand to finer clastics and carbonate, which became the shale and limestone of the Preble formation.

The record of Cambrian history after Preble time is interrupted in the Osgood Mountains, for the top of the Preble is not exposed and strata of Cambrian age later than the Preble formation are missing, except for a thrust sliver of Harmony formation. In the Hot Springs Range, however, the Paradise Valley chert is evidence that sedimentation continued in the Late Cambrian with the deposition of chert, a little siliceous shale, and limestone, and later with coarse feldspathic sand that become the Harmony formation.

CAMBRIAN DIASTROPHISM

Aside from the initial downsinking that formed the geosyncline in which the sediments began to accumulate, the first clear event in the structural history is the evidence provided by the Harmony formation that Precambrian crystalline rocks were exposed to erosion in the Late Cambrian. We do not know where the land was that furnished the feldspathic sands of the Harmony. It may have been near the site of deposition or the sediments may have been transported a long way by turbidity currents. The episode of active uplift seems to have been of limited extent, for the Harmony formation is restricted to north-central Nevada; no evidence of similar activity is found in Late Cambrian strata in eastern Nevada. Perhaps this was merely a local event, for there is no apparent angular discordance between the sediments of the Harmony formation and the strata of the underlying Paradise Valley chert.

ORDOVICIAN SEDIMENTATION

The geologic record is resumed in the Osgood Mountains with the Comus formation of Ordovician age, which provides evidence of continued deposition of fine clastics and carbonates in an environment that seems to have been intermediate between the miogeosyncline, represented by a predominantly carbonate section in eastern Nevada, and the eugeosyncline, where the chert, fine clastics, and volcanic rocks that characterize the Valmy and Vinini formations were deposited.

Between Ordovician and late Early Mississippian time the record in this area is missing. The nearest positive evidence of marine deposition in the Silurian and Devonian is found to the east and southeast in the Tuscarora Mountains and northern Shoshone Range, where strata belonging to both western and eastern assemblages have been brought together by thrust faulting. The only Mississippian that can be dated with confidence in this part of Nevada is the assemblage of limestone, chert, and greenstone belonging to the Goughs Canyon formation of late Early to early Late Mississippian age, and it has been carried in from elsewhere on a thrust fault. In other words, in the Osgood Mountains, autochthonous strata between the Ordovician and Lower Pennsylvanian are missing, and there is no way of knowing whether this was an area of nondeposition after the Ordovician or whether the rocks were deposited and subsequently eroded or stripped off by thrust faults.

ANTLER OROGENY

A major orogenic episode that is recognized elsewhere in north-central Nevada is clearly expressed in the Osgood Mountains by an unconformity that separates the conglomerate of the Battle formation and the carbonate sediments of the Etchart limestone from the older Paleozoic rocks. In recent years this important structural event has come to be known as the Antler orogeny (Roberts, 1951), its name taken from Antler Peak on Battle Mountain where its effects are so well shown. In the Osgood Mountains the orogeny can only be dated as post-Cambrian–pre-Middle Pennsylvanian, for the Battle formation of Middle Pennsylvanian (Atoka) age is not known to rest on rocks younger than Cambrian. To the east in Eureka County, however, rocks as young as Devonian were involved in the orogeny (Roberts and Lehner, 1955; Roberts and others, 1958).

The Antler orogeny was accompanied by thrust faulting on a grand scale, the first of several episodes of thrusting that took place in the Great Basin at various time between the late Paleozoic and early Tertiary. In this first episode great plates of siliceous clastic, and volcanic rocks of the siliceous assemblage, were sheared off, carried eastward, and thrust over rocks of the same age but of different lithologic character. Merriam and Anderson (1942, p. 1701–1706) showed that in the Roberts Mountains a great thrust fault, which they named the Roberts Mountains thrust, separates an eastern carbonate assemblage from a western assemblage of fine clastics, chert, and volcanic rocks. Gilluly (1954) recognized the thrust in the Cortez and Mount Lewis quadrangles, and Roberts and Lehner (1955) traced it northward through Eureka and Elko Counties approximately along the 116° meredian and showed that it passed into Idaho northeast of Rowland. Although Merriam and Anderson (1942, p. 1706) suggested that the Roberts Mountains thrust may have been of Late Cretaceous or early Tertiary age—that is, part of Laramide orogeny—Kay (1952, p. 1270) reported that in the Toquima Range and Carlin Canyon the thrusting preceded deposition of Pennsylvanian sediments. A similar age for the thrusting was postulated by Dott (1955, p. 2288 and fig. 11). Near Mountain City, conglomerate and limestone of Pennsylvanian age rest on a thrust sliver of rocks of the siliceous assemblage and also on rocks of the carbonate assemblage (K. O. Bushnell and J. R. Coash, oral communication cited in Roberts and others, 1958, p. 2854).

The Roberts Mountains thrust cannot be definitely recognized west of the Shoshone Range, but other thrusts on Battle Mountain, in the Sonoma Range, and in the Osgood Mountains quadrangle probably are contemporaneous with it. Thrust faults on Battle Mountain that are overlapped by the Battle formation (Roberts, 1951) probably are structurally higher than the Roberts Mountains thrust, for the Valmy formation of the western assemblage is in the block beneath the thrusts. The distribution of the Comus and Valmy formations in the Osgood Mountains quadrangle—both of Ordovician age but of different lithology—is the basis for postulating that a thrust fault contemporaneous with, if not an extension of, the Roberts Mountains thrust brought the two facies together. The Comus formation, believed to be an intermediate or transitional assemblage, is regarded as autochthonous or parautochthonous, for it is associated with the seemingly autochthonous Osgood Mountain quartzite and Preble formation of Cambrian age. Unfortunately, the structural relation of the Valmy formation is obscured by later normal faulting and poor exposures, but the interpretation made here is that it represents remnants of a thrust sheet of the siliceous assemblage that overrode the Hot Springs Range and Osgood Mountains. Some support for this structural interpretation is found on the east side of the Sonoma Range, where the Adelaide thrust (Ferguson, Muller, and Roberts, 1951) carries the Ordovician(?) Sonoma Range and Ordovician Valmy formations over Osgood Mountain quartzite and the Preble formation in the lower autochthonous block.

LATE PALEOZOIC SEDIMENTATION

Orogenic uplift in the late Paleozoic was accompanied by erosion and widespread deposition of sediments that overlapped preorogenic rocks of the carbonate, transitional and siliceous assemblages and that have been designated the overlap assemblage (Roberts and Lehner, 1955, p. 1661; Roberts and others, 1958, p. 2821, 2838). Rocks of the overlap assemblage in the Osgood Mountains quadrangle belong to the Antler sequence, which also occurs in parts of Battle Mountain, the Edna Mountains, and the Sonoma Range.

In the Osgood Mountains, parts of the area were once more beneath the sea by Middle Pennsylvania (Atoka) time, and clastic marine limestone was deposited on an eroded surface that exposed rocks as old as the Osgood Mountain quartzite. Uplift continued locally, however, for the limestones grade into and interfinger with coarse conglomerates that must have been derived from a nearby rugged terrain. The thinning and disappearance of conglomerate of the Battle formation westward in the Osgood Mountains suggests that a highland lay nearby east of the present range, and a marine basin lay to the west. Most of the area was probably beneath the sea by Late Pennsylvanian (Missouri or Virgil) time and continued so into the Permian, because remnants of marine deposits represented by the Etchart limestone are found throughout north-central Nevada as well as in the Osgood Mountains. Locally there may still have been uplift and erosion, which would explain why some conglomerates that are identical with the Battle formation contain interbedded limestone containing fossils of Antler Peak age; however, some of these conglomerates may be reworked early Pennsylvanian deposits. The variation in thickness and lithology of the Late Pennsylvanian and Permian deposits indicate that conditions of deposition and the kind and abundance of material being contributed differed considerably from place to place, suggesting an environment of many bays and straits, as in an archipelago.

Marine deposition was interrupted in this region between Early and Middle Permian time, and the previously deposited sedimentary rocks were moderately deformed, uplifted, and subjected to erosion. In Edna Mountain (Ferguson, Roberts, and Muller, 1952), on Battle Mountain (Roberts, 1951; Ferguson, Roberts, and Muller, 1952), and in the Sonoma Range (Ferguson, Muller, and Roberts, 1951) there are erosional remnants of the dominantly clastic Edna Mountain formation, which was laid down in middle Permian (Word) time on an erosion surface cut on the Antler Peak limestone or, where erosion had completely removed the Pennsylvanian formations, directly on Cambrian and Ordovician rocks. The Edna Mountain formation has not been identified in the Osgood Mountains quadrangle, and if deposited it has been completely removed by erosion.

LATE PALEOZOIC OR MESOZOIC OROGENIC HISTORY

Following deposition of the Edna Mountain formation, the rocks were again deformed during the Sonoma orogeny of late Permian age (Silberling and Roberts, oral communication, 1961). The orogeny is well illustrated in the China Mountain area of the Golconda quadrangle where rocks of Pennsylvanian and Early Permian age have been folded and thrust faulted and are overlain unconformably by the Koipato formation of Permian and Early Triassic age. Major structural events that have been assigned to the Sonoma orogeny include the Golconda thrust, a major thrust fault that has been traced throughout the region.

The early Mesozoic history in the Osgood Mountains quadrangle is uncertain, because no Mesozoic rocks have been identified. The historical record was interrupted again after the Upper Pennsylvanian and Lower Permian sedimentary rocks were deposited. Looking beyond the immediate area, however, we find that by the Middle Triassic marine deposits were being laid down in the area now occupied by the Sonoma Range and East Range (Ferguson, Muller, and Roberts, 1951), and sedimentation continued throughout the rest of the Triassic. So, presumably, this was a period of quiescence in the Osgood Mountains area too. Perhaps the Triassic sea covered all the region, but if sediments of Triassic age accumulated in the Osgood Mountains area they have been removed by erosion.

An orogenic episode affecting all of north-central Nevada began in the Early Jurassic (Ferguson and Muller, 1949, p. 9, 13) and may have continued into the Cretaceous and locally into the early Tertiary (Willden, 1958, p. 2396). In the Osgood Mountains the orogenic episode responsible for deformation of the post-Ordovician rocks cannot be dated closer than after the Late Pennsylvanian and Early Permian sedimentation and before the intrusion of granodiorite in Late Cretaceous time. The emplacement of granodiorite is believed to have been a significant part of the orogeny; granodiorite in the Osgood Mountains is of Late Cretaceous age, according to lead-alpha determinations.

The rocks in the Osgood Mountains quadrangle responded to the orogeny by moderate folding and extensive thrust faulting. Generally speaking, the late Paleozoic rocks beneath the thrusts were only moderately folded and some—for example, the Pennsylvanian strata in the prow-shaped bluffs above Hogshead Canyon and those in the northeast corner of the quadrangle—have been tilted without marked folding. The close folding of the Cambrian and Ordovician rocks below the thrust is attributed mainly to earlier deformation, probably during the Antler orogeny. The strata on the major thrust plates are closely folded and may have been deformed during the Late Permian Sonoma orogeny as well as during the Mesozoic orogeny.

Several thrust faults attributable to late Paleozoic or Mesozoic orogenies have been recognized in the Osgood Mountains. Each thrust except the uppermost is overlapped or cut by a higher thrust. This relation in the eastern part of the range is probably an indication of successive pulses of thrusting, but in the west part of the range it probably represents relative movement between a stacked series of thrusts which in the main moved together. The lowermost of the eastern thrusts seems to be the one that brought the plate of Adam Peak formation, a clastic facies of Pennsylvanian and Permian rocks over beds of the Etchart limestone, a predominantly limestone facies of approximately the same age. The Granite Creek thrust was active next and carried Preble formation over conglomerate of the Battle formation, the Etchart limestone, and the Adam Peak formation. Judging by the crosscutting relations, the plates of the imbricate zone of thrusting on the west side of the range were brought into the area next, and the general coextensive distribution and parallelism of the thrusts suggest that they moved together, perhaps riding on a thin sole of shale of the Preble formation. Minor thrusts in the Etchart limestone in the Goughs Canyon region are believed to outline subsidiary plates sheared off the underlying fixed block by movement of the overlying sheets. The Farrel Canyon thrust is the highest plate of the imbricate thrust zone. The age of the Farrel Canyon formation, on the upper plate of the thrust, has not been firmly established, but it may be in part the correlative of the Havallah and or the Pumpernickel formations. In the Antler Peak (Roberts, 1961), Golconda (Ferguson, Roberts, and Muller, 1952), and Winnemucca quadrangles (Ferguson, Muller, and Roberts, 1951), the Havallah formation, a dominantly clastic facies of Middle Pennsylvanian and Permian age, and the Pumpernickel formation of Pennsylvanian age are part of a very extensive thrust plate that overrode the region on the Golconda thrust fault. Therefore, the Farrel Canyon thrust may be equivalent to the Golconda thrust. Possible evidence of a still later episode of thrusting is found on the west side of the Osgood Mountains north of Anderson Canyon, where the Etchart limestone is thrust over Farrel Canyon formation; however, we believe it just as likely that an earlier thrust plate of Etchart limestone was broken and carried on the Farrel Canyon thrust plate.

In the northern part of the Hot Springs Range, north of this quadrangle, the Goughs Canyon formation overrides the Harmony formation on a thrust fault. Presumably thrust faults of late Paleozoic or Mesozoic age extended over the southern part of the range, too, but were removed by erosion. The deformation responsible for folding of the Harmony and Paradise Valley formations in the Hot Springs Range cannot be dated with certainty. Perhaps these rocks were deformed during the earlier Antler and Sonoma orogenies. However, they may have been folded in Mesozoic time. Westward-overturned Triassac rocks have been reported (Ferguson, Muller, and Roberts, 1951) beneath thrust faults of Mesozoic age in the Sonoma Range.

Possibly, although this cannot be proved, some of the high-angle faults were formed during the late Paleozoic and Mesozoic orogenies. Many of the northwest-trending faults that displace the Osgood Mountain quartzite and the rocks of Pennsylvanian and Permian age near the crest of the range in the south-central part of the Osgood Mountains may have been minor tear faults developed during the thrusting and folding; the northwest faults on the northwest side of the Hot Springs Range may be regarded similarly. Possibly some of the faults on which there was Cenozoic movement were also sites of earlier movement.

The Mesozoic orogeny culminated in the Late Cretaceous with the intrusion of granodiorite in the Osgood Mountains and the Hot Springs Range. Some deformation attributable to emplacement of the granodiorite can be recognized adjacent to the Osgood Mountains stock, but most contacts indicate that the intrusive bodies cut the rocks cleanly, without distortion. No special control of emplacement of the major bodies by local structure is apparent, but the main stock in the Osgood Mountains is virtually conformable to the eastward-dipping Preble formation. Some minor dacite porphyry bodies, regarded as satellitic to the granodiorite, occur on or near the contact between Osgood Mountain quartzite and Etchart limestone, which may have guided emplacement of the intrusive to some extent. Similarly, other bodies are near or on the Peak 6837 thrust fault in the north end of the range. Intrusion of the granodiorite was accompanied by transfer of heat and material that metamorphosed the adjacent country rocks, locally converted limestone beds to silicate rocks, and deposited minerals bearing tungsten, lead, zinc, and some copper and gold. Where some of the emanations accumulated in the high parts of partly solidified granodiorite they caused extensive modification of the earlier crystallized minerals.

CENOZOIC HISTORY

EARLY CENOZOIC HISTORY

The history of the Osgood Mountains and the Hot Springs Range following the Mesozoic orogeny is incompletely recorded in the rocks, but during the earlier part of the Cenozoic period extensive erosion and terrestrial and lacustrine deposition seem to have been the main geologic processes in operation. Subsequent erosion has destroyed much of the evidence, but in a few places poorly consolidated sands and gravels are preserved beneath volcanic rocks that fell as ash and flowed as lava over a surface of low relief in the ranges. The episode of volcanism is recorded by rhyolitic tuffs and andesitic and basaltic flows. The tuffs locally underlie later and more extensive andesitic flow rocks, which cannot be closely dated but which probably are of late Miocene to middle Pliocene age; a basalt flow at the south boundary of the quadrangle near the Humboldt River may be of Quaternary age.

CENOZOIC DIASTROPHISM

The last diastrophic episode, which came in the late Tertiary and extended into the Quaternary, is chiefly recorded by high-angle faults that displace the Tertiary volcanic rocks, and by uplift of the ranges. Other high-angle faults in the ranges, many of which transect older structures, may also have formed during the late Cenozoic orogeny, although some that follow earlier structural trends may have had pre-Cenozoic movement. Uplift of the ranges probably was accomplished partly by displacement along north-trending faults near the present front of the ranges, but lack of direct evidence of boundary faults, except in a few places, makes a hypothesis of uplift by arching accompanied by some faulting attractive. Some range-front faults are buried beneath alluvium in the basins. At least one of the range-front faults, the Getchell fault, had an important strike-slip component of movement that may have been nearly as large as later dip-slip displacement. A few of the fractures provided channels for rising hydrothermal fluids that altered the wallrocks, and deposited metal-bearing minerals: arsenic and gold along the Getchell fault, and quicksilver in faults in the Hot Springs Range. Continuation of diastrophism into Quaternary time is expressed by faults that displace valley alluvium and by partly eroded older fan deposits that signify rejuvenation of drainage.

Erosion consequent on uplift is largely responsible for the present topography in the ranges, though some remnants of an old surface of low relief are found at the crest of the ranges. Erosion has cut deeply into the ranges, obviously guided in many places by zones of weakness along high-angle faults. Detritus from the erosion of the ranges accumulated in thick deposits in the basins.

MINERAL DEPOSITS

Mining activity in the Osgood Mountains quadrangle has been directed chiefly toward exploitation of tungsten and gold deposits in the Osgood Mountains. A small production of quicksilver and gold has come from deposits in the Hot Springs Range. In addition, lead-silver and copper deposits of minor importance are known but have yielded little or no production.

For purposes of description the metalliferous deposits are classified mainly according to their metal content and distribution. Each class has a distinctive mode of occurrence and, for the most part, is genetically distinct. Following is the classification adopted here:

1. Tungsten deposits in the Osgood Mountains
    a. deposits in tactite
    b. deposits in altered granodiorite
2. Gold-arsenic deposits in the Osgood Mountains
3. Quicksilver deposits in the Hot Springs Range
4. Gold-bearing quartz veins in the Hot Springs Range
5. Gold-scheelite-cinnabar placer in the Hot Springs Range
6. Minor deposits of lead-silver and copper

Nonmetallic deposits include three barite prospects in the Osgood Mountains and an unexploited quartzite bed of exceptional purity in the southeast part of the Hot Springs Range that might be used as a source of silica.

TUNGSTEN DEPOSITS IN THE OSGOOD MOUNTAINS

Tactites, which were formed by metamorphism of limestone, have been the main host rock for the scheelite that has been produced from mines in the Osgood Mountains. The tactites that have been mined for their scheelite content are along the contact of the granodiorite stock where it has intruded limestone beds of the Preble formation. Minor amounts of scheelite have been produced from alteration zones and bodies of quartz in granodiorite; but such occurrences, though interesting, are of little past or future importance as sources of tungsten ore.

The only tungsten mineral that has been recognized here is the calcium tungstate, scheelite. The ore is almost exclusively tactite that contains accessory amounts of scheelite and minor amounts of metallic sulfides. At some places scheelite is also found in calc-silicate rock and granodiorite, but usually not in sufficient amount for the rock to be considered ore. In addition, at a few places scheelite occupies narrow seams and small pockets of intense alteration in granodiorite, and in the northern lobe of the stock coarsely crystalline quartz in a zone of altered granodiorite contains some scheelite.

DEPOSITS IN TACTITE

Scheelite is an accessory mineral of relatively low concentration in most of the tactite but is sufficiently plentiful at many places so that the tactite can be mined for its tungsten content. The grade of most of the tactite mined ranges from approximately 0.3 percent to approximately 0.7 percent WO_3, and the average grade of all the ore mined contained about 0.5 percent WO_3. Tactite containing less than 0.3 percent WO_3 has been mined where the bodies are situated so that they can be mined in conjunction with material of higher grade.

MINERALOGY AND PARAGENESIS

Scheelite occurs in the tactite as small discrete subhedral to euhedral grains that range from less than 0.5 mm to 2 mm along their greatest dimension. The scheelite grains commonly are scattered through the tactite as if they crystallized at the same time that the tactite was formed; in some specimenes the scheelite grains are enclosed in garnet and, less commonly, in pyroxene. On the other hand, at many places scheelite grains are distributed in thin stringers parallel to the tactite layering as if they were localized along zones of voids remaining after formation of the tactite; in some places the scheelite stringers extend out from veinlets of schelite or scheelite and garnet that cut across layered tactite (fig. 12*B*).

Scheelite occurs in a variety of ways in the tactite. In some specimens it is interstitial to garnet and to garnet and other minerals (fig. 12*A*); in others, equant scheelite crystals occur in quartz-calcite veinlets; in others, scheelite replaces interstitial calcite along calcite-garnet boundaries. Anhedral grains of scheelite are enclosed in some epidote veins that cut tactite. Late-stage quartz commonly replaces tactite and granodiorite at their contact, and some of these quartz replacements contain rich concentrations of coarse scheelite crystals; however, a few of the replacements are barren or have a low scheelite content. Scheelite is associated with green actinolitic amphibole and calcite in an unusual variety of tactite that is represented by the west ore body in the Riley Extension mine and in lesser amounts elsewhere and that is possibly a late modification of a garnet-pyroxene tactite.

The most plentiful gangue minerals of the tactite are the silicate minerals of the host rock, and calcite. In addition, metallic sulfides are erratically distributed in the tactite. The following sulfide minerals have been recognized: pyrite, chalcopyrite, molybdenite, sphalerite and galena. Hobbs (1948,[15] p. 68, 74) has also reported bismuthinite, but none was recognized during this investigation. Most of the tactite and adjoining rocks contain no sulfides except pyrite, but locally all the above-mentioned sulfides are abundant and commonly they occur together.

[15] See footnote 1, p. 6.

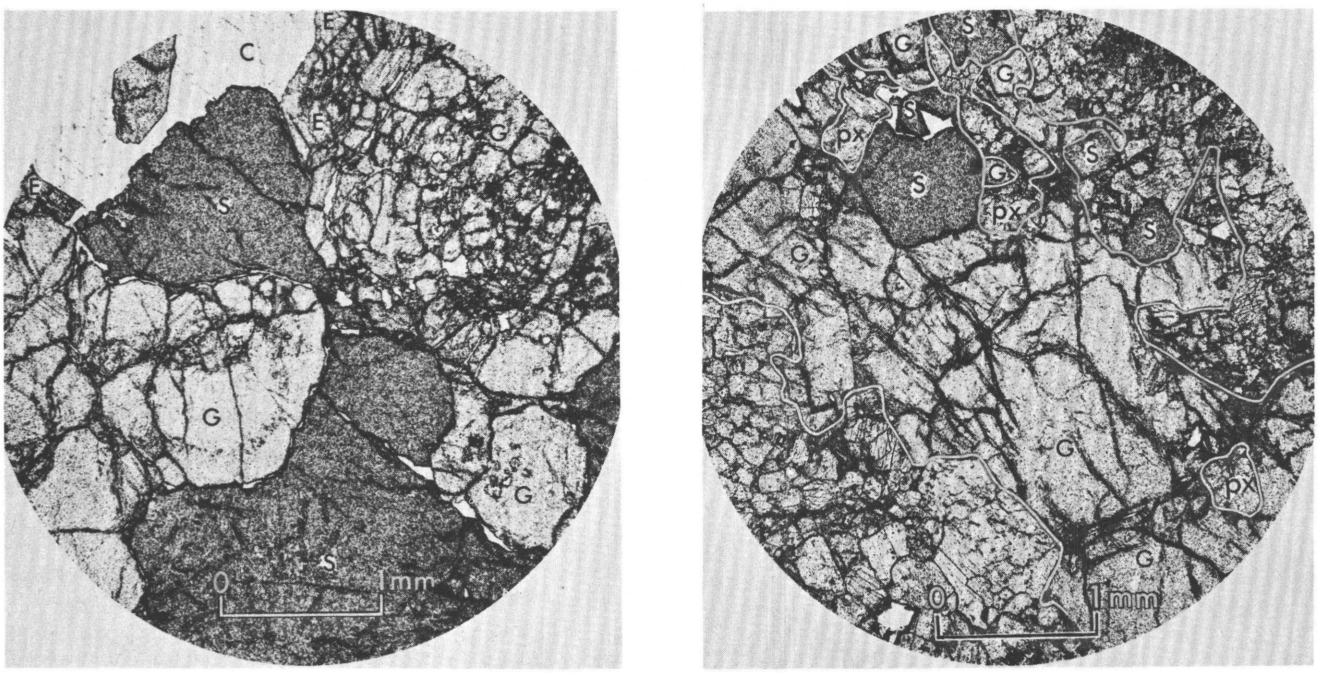


FIGURE 12.—THIN-SECTION PHOTOMICROGRAPHS OF SCHEELITE IN TACTITE. PLANE POLARIZED LIGHT.

A, Scheelite (S) and calcite (C) filling spaces between zoned birefringent garnets (G) and some epidote (E) on boundary between calcite and garnet.

B, Garnet-pyroxene tactite cut by more coarsely crystalline irregular-shaped veinlet of birefringent garnet, accompanied by grains of scheelite. Garnet (G), pyroxene (px), scheelite (S).

Tactite at many places contains a few scattered grains of pyrite and locally it is permeated with pyrite that may amount to as much as 10 to 15 percent of the rock. The crystals of pyrite range from less than 1 mm to as much as 2 cm along an edge. Usually, lesser amounts of chalcopyrite accompany pyrite, but in some places the ratio is reversed. Molybdenite is not common, but small local concentrations have been observed in tactite close to the granodiorite contact, commonly attended by abundant epidote or fibrous uralitic amphibole and late-stage calcite. Sphalerite and galena are restricted in their distribution to small highly concentrated bodies at a few places close to the granodiorite-tactite contact. Sphalerite tends to be more plentiful than galena in an estimated ratio between 3:1 and 5:1. Variable amounts of pyrite and chalcopyrite accompany the sphalerite and galena.

Aside from the specks of pyrite and chalcopyrite that are scattered through some of the contact rocks, the local concentrations of sulfides tend to be on or close to the granodiorite contact. Mostly, they are in the tactite or other contact-metamorphosed carbonate rocks; uncommonly, they are in the granodiorite next to the contact. Relations to the host rocks suggest that the sulfides occupy space made available by a combined filling of intercrystal boundaries and replacement of adjoining grains. No clearly defined veinlike fracture fillings of sulfides in the tactite were observed. The relation of the sulfides to scheelite is obscure at most places. Locally, concentrations of sulfides are attended by larger amounts of fine-grained scheelite. In a few specimens, microscopic examination showed mutual relations strongly suggesting that formation of scheelite preceded the introduction of sulfides.

FORM AND STRUCTURAL RELATIONS

Most of the tactite bodies are tabular, though some are irregular in shape and discontinuous. They range from a few inches to as much as 35 feet in thickness, but the majority are between 5 and 15 feet thick; the thickness may vary along the strike and down the dip. Most of the tactite zones are fairly continuous along the strike, commonly for several hundred feet and in places for more than 2,000 feet, and they are continuous down the dip for hundreds of feet. The tactite bodies are parallel to the granodiorite contact; where the limestone strata are concordant with the contact, the tactite is parallel to the bedding; but where the limestone is discordant, the tactite also is discordant with the bedding, although locally concordant tongues of tactite may extend out along bedding into the wall rocks. The contacts are sharp and well defined on both the footwall and hanging wall. Some of the tactites end abruptly along strike or down dip by a sudden tapering, or bluntly with perhaps one or more stubby protuberances into limestone or calc-silicate rock.

No markedly crosscutting tactite bodies of veinlike habit have been observed. At many places, however, a relatively thick body of tactite that otherwise terminates rather abruptly has thinner fingerlike extensions that suggest localization of tactite-forming solutions by fractures beyond the main part of the body. The tactite bodies are cut by high-angle faults on which the displacement is ordinarily relatively unimportant and measurable in a few tens of feet of normal and strike-slip movement.

In many places the form and size of tactite bodies is directly related to the configuration of the granodiorite contact surface (fig. 13). The contact surface is not smooth but is corrugated with large-scale ridges and troughlike grooves. On maps and sections these irregularities are expressed in the trace of the contact as sharp, commonly right-angle changes of strike or dip. Mostly the corrugations are steeply inclined, but some troughs have a nearly horizontal axis. The thickest parts of several tactite bodies are in the troughlike grooves on the contact surface, and they become thinner away from the axial part of the structures. Where a body of tactite is continuous for several score or hundreds of feet along strike, the tactite is commonly thinner over slight outward bulges or broad ridges on the contact surface than on either side. In places where the contact is relatively smooth, tactite is localized at minor irregularities, whereas the smoother segments of the contact are barren even though limestone is available for replacement by tactite. The corrugations persist downward for as much as several hundred feet; underground development and mining have demonstrated that at several places they plunge steeply but at a different angle from the dip of the contact, and that the tactite bodies follow the plunge of the corrugations.

On a larger scale, also, this correlative relation between tactite and the shape of the granodiorite contact is generally valid. It is apparent from the map (pl. 1) that most of the tungsten deposits are situated in sharp-angled reentrants of limestone along the granodiorite contact.

ORIGIN

The tactites are regarded as having formed from marble and calc-silicate rock by reaction with heated solutions whose source was the intrusive body of granodiorite. The solutions contained tungsten, so that some scheelite was formed simultaneously with the tactite, but scheelite deposition also continued from solutions that followed permeable avenues along the granodiorite contact after formation of the tactite.

The distribution of tactite bodies in relation to irregularities in the granodiorite contact indicates that tactite was formed after the granodiroite was emplaced. The tactite bodies range in width along strike from many feet to a few inches with no correlative change in the bulk composition of the adjacent granodiorite, which may be evidence that the granodiorite at the contact was not directly responsible for the metamorphism of the limestone to tactite. If convection currents were operative, however, they might have obliterated changes in granodiorite at the contact. Channeling of solutions by troughlike structures along the contact would account for the fact that many tactite bodies are situated at reentrants and irregularities in the contact. Mining has shown that the tactite bodies are continuous downward along the contact and do not end as long as the structure is favorable, which lends support to the hypothesis that the reacting solutions were guided along suitable structures. It seems likely that fracturing along the contact, perhaps caused by stresses consequent on emplacement of the stock or by cooling of the intrusive mass, may have controlled the permeability along the zone of replacement, but if so, the formation of tactite has obscured the evidence.

The paragenetic relation of scheelite to the common tactite minerals cannot be decisively established, but the data suggest that some scheelite was deposited when the tactite was formed and that scheelite deposition continued during the waning stages of the magmatic episode.

Deposition of scheelite after the tactite was formed required that the rocks in which it was deposited were permeable to tungsten-bearing solutions—either by reason of voids that remained after alteration or by continued fracturing—and that the solutions were made available to the tactite. The sporadic distribution of scheelite in the tactites is explainable on the basis that the avenues by which tungsten-bearing solutions could enter were wholly or partly closed after the tactite formed. For example, one of the largest tactite bodies, at the Richmond mine, contains less scheelite on the average than any of the other tactites, and large volumes of it are essentially barren. Perhaps a better example is found in the underground workings of the Tonopah mine, where the central part of a thick tactite lens is barren and scheelite mineralization is confined to the borders next to the granodiorite contact.

Composition of the sedimentary rocks at the granodiorite contact also controlled the localization of the scheelite deposits, for although structural controls of permeability probably determined the sites that were accessible to tungsten-bearing solutions, scheelite as

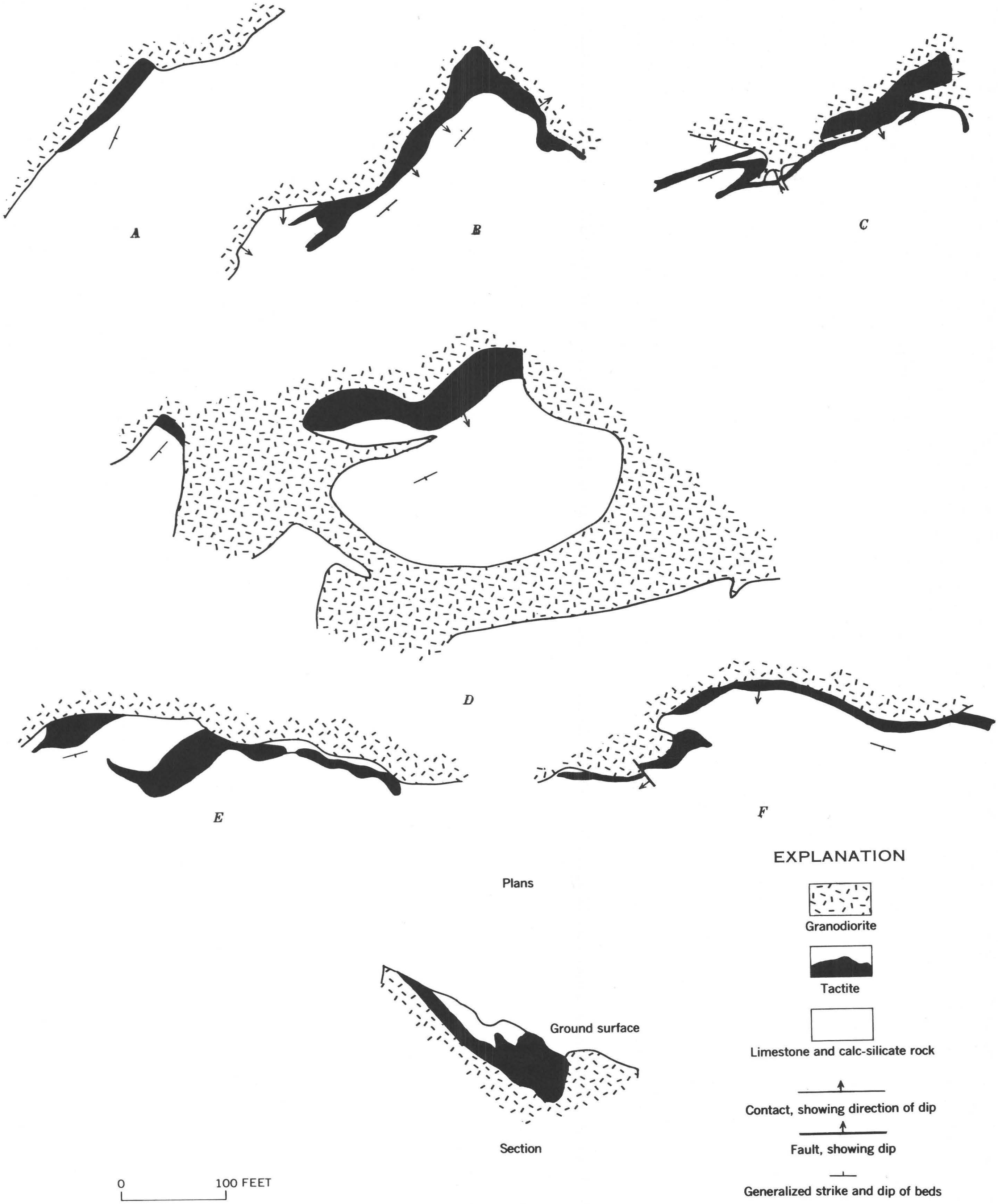


FIGURE 13.—Sketches of some tactite bodies, showing typical shapes and relations to the granodiorite contact.

well as the gangue minerals of the tactite host rock were formed by chemical processes. Limestone bodies at the contact were necessary for formation of tactite, and scheelite is essentially restricted to tactite. It seems unlikely that all contacts against hornfels were completely sealed, or that the hornfels next to the contact was so free from fractures that the rocks were impermeable to the same solutions; yet no tactite or scheelite were formed, because the rocks were chemically inert with respect to the composition of the same fluids that reacted with marble and calc-silicate rocks elsewhere.

The origin of the scheelite in the late quartz that came in along the granodiorite-tactite contact in places is puzzling. Not all of the late quartz contains scheelite, yet at places in silicified tactite, scheelite is unusually abundant. The quartz-forming solutions may have carried tungsten, but possibly earlier formed scheelite in the tactite was mobilized and reprecipitated with the quartz, or remained as a residual in bodies of replacement quartz.

Although at some places scheelite is abundant where sulfide minerals are abundant, there is no concrete evidence that they were formed together. On the other hand, there is considerable paragenetic evidence that scheelite is earlier than the sulfides. Possibly some sites where scheelite was abundantly formed were also favorable for the deposition of sulfides.

DEPOSITS IN ALTERED GRANODIORITE

In the course of mining the tactite deposits, pockets and seams of scheelite have been discovered in the adjacent granodiorite at a few places. Most of the occurrences are too small to be of any economic importance except where they can be mined with the tactite. At the Mountain King mine, however, some seams in granodiorite were so rich that the scheelite was dug out by hand with a pick and collected in powder boxes.

Typically, coarse scheelite crystals are embedded in somewhat porous, light-colored, soft masses of micaceous material. The micaceous material is mostly muscovite, some pale-brown hydrobiotite, and small clear crystals of quartz. Microscopic crystals of apatite with smoky gray centers are fairly plentiful. The associated scheelite is in subhedral to euhedral crystals, some of tabular habit and some with octahedral form. They range from 1 mm to as much as 1.5 cm in size. A few are clear and colorless, but most are gray, dusky yellow, and moderate yellowish brown. With ultraviolet light the yellowish, brownish and colorless crystals fluoresce blue, while the gray ones give off a bluish-white light. The scheelite crystals are concentrated in the central part of ovoid pockets and in the middle of the seams.

The micaceous, scheelite-bearing seams are approximately 2 to 4 inches wide. At the Mountain King mine they are more or less parallel to gently inclined joints in the granodiorite, and may be localized along some of the joints, though some of the seams pass into the adjacent hornfels. Their dimensions are uncertain, but probably they are neither very long nor wide, for apparently they do not extend more than a few feet or a few tens of feet beyond the granodiorite contact. The micaceous pockets in granodiorite are along or within a foot or so of the contact. They are more or less ovoid and 6 to 10 inches in greatest dimension. Their internal surface is drusy with small singly terminated crystals of quartz. Some drusy cavities were seen that have a little white mica on the surface of the quartz crystals, but it is uncertain whether or not these cavities were originally filled with mica. The granodiorite is slightly altered for ½ to 1 inch next to the micaceous seams and pockets, chiefly by slight sericitization of plagioclase.

An unusual deposit of scheelite in the altered zone in the central part of the northern lobe of the Osgood Mountains stock contains scheelite with coarsely crystalline quartz in the most intensely altered part of the body. At the section 5 (Marshall Canyon) pit, crystalline quartz in soft, altered granodiorite ranges from crystals an inch or so long to huge bodies with intergrown individual crystals as much as 4 inches in diameter and 6 inches in length. Some of the quartz contains small subhedral crystals of scheelite. The scheelite seems to be intergrown with the quartz, some of it may be in fractures. Some, however, is within the quartz as intracrystalline phantomlike films, as if it were deposited during formation of the quartz. At places in the pit irregular-shaped masses of sericite rock are found, which are similar to the masses that contain scheelite in seams and pockets elsewhere.

ORIGIN

The relations between the deposits of scheelite in tactite and the scheelite that is in fractures and cavities at a few places in the granodiorite is not clearly established, but presumably they belong to the same general episode of mineralization. The association of sericite aggregates with the scheelite is indicative of the alkaline composition of the solutions that traversed the granitic rock, the sericite having been formed by alteration of the feldspar. Calcium necessary for precipitation of scheelite from an alkaline solution (Lovering, 1941, p. 269–271) could have been obtained from plagioclase in the granitic rock or possibly, as all occurrences are near or at the contact, it came from adjacent calcareous beds. Scheelite that is intergrown with quartz at the section 5 (Marshall Canyon) deposit indicates that the siliceous solutions depositing quartz in the sericitized granodiorite also contained some tungsten.

GOLD-ARSENIC DEPOSITS IN THE OSGOOD MOUNTAINS

The Getchell gold mine, which was developed and successfully mined on a large scale by open-pit methods, had brought considerable attention to the Osgood Mountains before active mining of tungsten deposits began. The gold deposits are situated along the Getchell fault near the base of the northeast side of the range.

A few years before our study of the Osgood Mountains quadrangle, Peter Joralemon made a careful field investigation of the Getchell gold deposit and a detailed laboratory study of the ore. Our work at the Getchell mine was limited to a few days study of geologic features exposed in the open pits, and a brief examination of part of the accessible underground workings. Most of the data contained in this section, including the maps, have been taken from Joralemon's reports (1949[16]; 1951).

DISTRIBUTION

The gold-arsenic deposits are in fractured rocks along the Getchell fault zone and its possible southward extension, the Ogee and Pinson fault. The main deposit, however, is confined to a zone nearly 7,000 feet long at the northern end of the Getchell fault zone. Marginal gold ore is known to occur in the Getchell fault zone at the Riley tungsten mine, and material in a fault zone east of the Pacific mine—possibly an en echelon branch of the Getchell fault—contains gold.[17] The Ogee and Pinson mine south of Granite Creek yielded some low-grade gold ore from fractured chert and hornfels. Some realgar is also reported here by Joralemon (1949,[17] p. 92), which suggests that this deposit belongs to the same episode of mineralization as the Getchell mine.

GENERAL FEATURES

Joralemon (1951, p. 270) has described the general features of the gold deposits as follows:

> The gold ore bodies are sheet-like masses that lie along the various strands of the Getchell fault zone. They extend at least 7,000 feet horizontally and 800 feet down the dip, and vary in width from a few feet to more than 200, averaging about 40 feet wide. The vein pattern, extremely complex at the surface, becomes simpler with depth and, deeper than 800 feet below the surface, all the branch veins apparently join to form a single persistent structure * * *. The veins consist of sheared and mineralized argillite [here called hornfels] and limestone cut by quartz and calcite veins and containing a soft, plastic gumbo that has replaced the sediments. The gumbo is the principal gold-bearer.

He said further (p. 273):

> The gumbo, the important gold-bearer, is unusual in that while it appears to be a fault gouge, it actually consists of minute subhedral quartz crystals embedded in a nearly submicroscopic intergrowth of quartz and amorphous carbon. In places veins of gumbo several feet wide and parallel to the surrounding sedimentary rocks end abruptly against unsheared limestone or argillite.

Joralemon (1951, p. 273) regarded the gumbo as having formed hydrothermally in two stages:

> First was the replacement of wall rock by fine-grained quartz to form a porous aggregate of subhedral quartz crystals in narrow bedding veins. In the second stage, carbon, probably derived from underlying argillite, was introduced into the porous quartz veins and was deposited interstitially to the quartz. * * * Those portions of the veins that contained the most gumbo * * * were the most permeable parts of the veins * * * and the later hydrothermal solutions were there concentrated.

Open, vuggy structures are common, especially in some of the calcite and late quartz veins; uncommonly, some small vugs occur in the gumbo.[18] In contrast, however, there is a "* * * nearly complete lack of certain common epithermal structures such as 'colloform' banding, crustification, and comb structure" (Joralemon, 1951, p. 306).

MINERALOGY AND PARAGENESIS

Nonmetallic gangue minerals of the deposit in addition to quartz, calcite and carbon are sericite and chlorite, and distinctly minor and sporadic barite, gypsum, fluorite, and chabazite. Metallic minerals include pyrrhotite, pyrite, marcasite, chalcopyrite, arsenopyrite, orpiment, realgar, and very minor amounts of scheelite, stibnite, cinnabar, and magnetite. The precious metals are gold and silver. The paragenetic relations of the minerals and their relative abundances are diagrammatically illustrated in figure 14, modified from Joralemon (1951, p. 305).

Concerning the distribution of the metallic minerals, Joralemon (1951, p. 273) said that pyrite and minor pyrrhotite are the earliest sulfides and occur in irregular porous masses in the ore shoots, whereas in the "outlying segments of the veins * * *" pyrite tends to be euhedral.

> Marcasite constitutes about 5 percent of the iron sulfides in the central channelways, but is rare or nonexistent in the outlying areas. It occurs as irregular shells or rims encircling earlier pyrite grains.

[16] Joralemon, Peter, 1949, The occurrence of gold at the Getchell mine, Nevada: Harvard Univ., unpublished doctoral thesis.
[17] Op. cit., p. 91.

[18] Op. cit., p. 58–62.

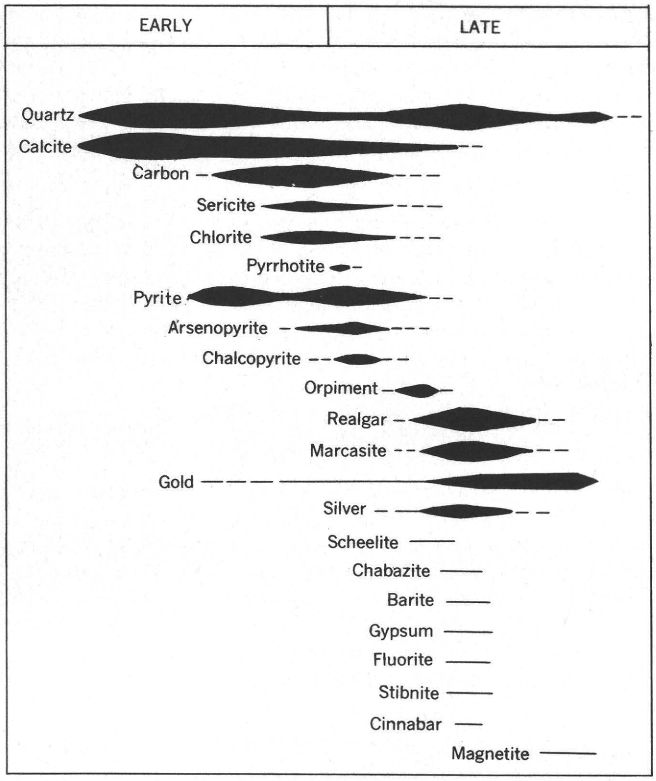


FIGURE 14.—Sequence of mineral deposition at Getchell gold deposit (from Joralemon, 1951, figure 29).

Chalcopyrite is present in extremely small quantities as grains embedded in the gumbo and as blebs in the pyrite.[19]

Arsenopyrite is present in minute euhedral crystals and is almost entirely restricted to parts of the veins in andesite porphyry.—(Joralemon, 1951, p. 273).

One of the most distinctive mineralogical features of the Getchell deposit is the association of the arsenic sulfides, orpiment and realgar, with the gold. Joralemon (1951, p. 273) said that they

* * * are the most abundant of the metalliferous minerals and unlike pyrite, are entirely restricted to the areas within the veins that now contain noteworthy quantities of gold. Orpiment is apparently earlier than realgar and is found only in isolated pockets, characteristically occuring in low-grade sections of the veins * * * [where it] constitutes as much as 40 percent of the mineralized material and is veined and coated by realgar. Realgar occurs throughout the commercial ore bodies in amounts ranging from 1 to 10 percent. It shows such a consistent relation to gold that the presence of realgar is usually a strong indication of good values.

Very finely crystalline stibnite, commonly accompanied by ilsemannite, is restricted to a few isolated pockets and narrower pipelike bodies where orpiment, realgar and marcasite are also present. Cinnabar is very uncommon but coats calcite and chabazite in a few branch veins. (Joralemon, 1951, p. 273–274).

According to Joralemon (1951, p. 276)

Magnetite is apparently the latest of the metalliferous minerals associated with gold, and is the one most closely related to gold. It occurs only in microscopic particles and has been seen only in the gumbo and andesite. * * * It is so closely associated with gold that any polished sections containing a cluster of magnetite grains have also some gold close at hand, generally within a millimeter.

According to Mr. Keith Kunze (oral communication, 1957) the gold ore contains a few one-thousandths to a few one-hundredths of one percent WO_3, and the tungsten minerals scheelite and hübnerite have been identified in ore on which metallurgical tests had been made. Joralemon[20] noted that in the north end of the 600 level scheelite veinlets

* * * less than a millimeter wide cut the andesite porphyry along minor slips. In a 2-foot-wide band in the gumbo near the dike, it occurs as fracture-filling, crusts, and as minute disseminated grains closely related to realgar. * * * Other minor veinlets occur in the wall rock northeast of the turn in the gold vein.

The nature of the oxidized gold ore that was produced from the open pits at Getchell mine is poorly known; Joralemon did not describe it, and none was seen by the writer, for it was mined out and primary sulfide ore had been reached in the open cuts. According to Hardy's (1938, p. 31; 1940, p. 4) descriptions the oxidized ore was siliceous and formed prominent outcrops; much of the silica seems to have been of secondary origin, the pyrite and pyrrhotite were

* * * oxidized to limonite and hematite, and the sulphides of arsenic have been oxidized to a soluble form and removed by meteoric waters.

GOLD

Concerning the distribution of gold at the Getchell mine, Joralemon (1951, p. 276–288) said that economic amounts of gold are restricted to the veins within the fault zone; however, the entire zone containing the ore bodies may be visualized as a block of submarginal gold-bearing rock within which are downward-projecting rootlike ore shoots. The subeconomic parts of the veins between the richer shoots, and the wall rocks for as much as 400 feet on either side of the ore body contain 0.01 to 0.08 ounce of gold per ton. The ore boundaries are economic and at the economic vein wall the gold content increases to more than 0.1 ounce per ton.[21]

According to Joralemon's laboratory investigations (1951, p. 276–281) the bulk of the gold in the economic parts of the deposit is in microscopic particles that

[19] Op. cit., p. 79.

[20] Op. cit., p. 84–85.

[21] Op. cit., p. 86.

range from a fraction of a micron to nearly 1 mm, and the smaller grains are more abundant than the larger. He found that visible gold is sparsely disseminated through all the rock types but that most of it is intimately associated with the fine-grained quartz-carbon matrix of the gumbo. In the gumbo some of the gold is uniformly distributed in small isolated particles, but it also is concentrated in pods or lenses as much as 100 microns wide and several millimeters long and is intimately associated with magnetite and carbon. Inclusions of visible gold are also found in all the major sulfides of the ore zone, though the amount is probably less than in the gumbo. Of the sulfides, pyrite and marcasite are the principal hosts. To account for submarginal values found in the rocks outside the ore shoots, where there is essentially no visible gold, Joralemon (1951, p. 281–292) reasoned that the gold may be mainly in solid solution in pyrite and carbon.

SILVER

The gold: silver ratio in bullion assays varies from 2:1 to 134:1 and averages 10:1 for the entire bullion production (Joralemon, 1951, p. 294). Joralemon (1951, p. 294) observed microscopic metallic grains in the Getchell ore that he concluded were native silver, although the particles were so small that conclusive chemical tests were not possible. No other silver-bearing minerals have been recognized, nor does silver seem to occur, except very rarely, as electrum, the natural silver-gold alloy.

The following observations on the occurence and distribution of silver were made by Joralemon (1951, p. 295–296):

> In general the amount of silver in the ore varies inversely with the amount of gold. Extremely rich portions of the ore body contains less silver than the portions of average grade. * * * Native silver persists with no large decrease in amount to the deepest levels of the mine. * * * Silver particles show the same variations in grain size as the gold particles. * * * Like gold, silver occurs in the carbonaceous matrix of the gumbo and in sulfides, particularly the porous pyrite-marcasite intergrowths.

ORIGIN AND CLASSIFICATION OF THE DEPOSIT

The general sequence of deposition of the minerals in the Getchell mine ore zone is illustrated in figure 14. Joralemon (1951, p. 299) concluded that:

> The changes in the nature of deposition with time were brought about largely by changing conditions of deposition. The changes were gradational and the minerals characteristic of the early stage were being deposited in the outlying areas at the same time that the late stage minerals were forming in the ore shoots. At the earliest stage of deposition at the known depth horizon, pyrite containing low-grade gold was deposited, with some quartz, calcite, sericite, and chlorite, near the major channelways. As the flow of hot solutions continued, more and more heat was carried into the zones of mineralization, the intensity-character increased, pyrite and lean gold were deposited toward the centers of mineralization by replacement or re-solution of those minerals that had been stable during the earlier and feebler stages then and there existing. Realgar, marcasite, and the economically important gold, which latter was becoming richer with time, were deposited in the main channelways at about the same time that the solutions in the outlying portions of the veins and wall rock were depositing disseminated pyrite and included submicroscopic gold in noncommercial amounts.

Joralemon (1951, p. 292) regarded the rootlike ore shoots as representing the sites of original channelways along which most of the gold-bearing fluids passed throughout the period of gold deposition:

> These richer shoots coincide with the zones of extensive development of gumbo, suggesting that even before gold deposition began, these areas were most permeable to the mineralizing solutions.

The gumbo, being unconsolidated, was more permeable than the enclosing rock, and therefore afforded more favorable channels for transportation of the gold-bearing fluids. The intense shattering of rocks in the ore zone, the vuggy structures, the unconsolidated nature of the gumbo, and the relatively shallow blanket of richer ore with its rootlike downward projections suggest that the Getchell deposit was formed under a relatively shallow cover (Joralemon, 1951, p. 308).

In many ways the Getchell deposit is similar to the epithermal gold-realgar deposits at Manhattan, Nev. (Ferguson, 1924), 150 miles to the south. Both have a high gold-silver ratio and abundant realgar and orpiment in addition to carbon, pyrite, stibnite, fluorite, and cinnabar; lead, zinc, and copper sulfides are very scarce or absent. D. E. White (1955, p. 143) and Joralemon (1951, p. 306) called attention to characteristics of the Getchell deposit that are similar to hot spring deposits at Steamboat Springs, Nev., where siliceous muds precipitated at shallow depth in the spring system contain small amounts of gold, and where cinnabar and aggregates of minute stibnite needles are also being deposited.

The Getchell ore bodies can be classed as epithermal deposits, and Joralemon (1951, p. 309) suggested that they may represent a gradation between the depositional environments of typical epithermal gold deposits and lower intensity stibnite, realgar and cinnabar deposits.

AGE OF MINERALIZATION

The episode of mineralization that formed the Getchell deposit is later than the emplacement and solidification of the granodiorite stock of Late Cretaceous age, for the ores are localized in a fault zone that cuts the granodiorite. The deposit also is younger than andesitic dikes that postdate the granodiorite and have been mineralized in the ore zone. Alluvium of Recent and possibly Pleistocene age overlays the mineralized zone. On evidence from geologic relations in the immediate area, therefore, the mineralization is of Tertiary age.

Indirect evidence suggests that the Getchell deposit may be related to a widespread episode of mineralization of late Tertiary age in Nevada. Ferguson (1929, p. 131–141) has suggested a classification for the precious metal deposits of Nevada based on the gold-silver ratios of the ores, which is supported by certain correlative features that indicate that the classification is valid from a genetic standpoint (Nolan, 1933, p. 625). In this classification, deposits in which the ratio of silver to gold is greater than 1 to 1 seem to be of Miocene (pre-Esmeralda) age whereas those in which gold exceeds the amount of silver are of late upper Miocene or Pliocene (Esmeralda and post-Esmeralda) age (Ferguson, 1929, p. 131–136). The high gold-silver ratio of the Getchell deposits puts it in the second class, and the mineralogical peculiarities strongly suggest that it is genetically similar to the Manhattan deposit, which belongs to the later gold-rich deposits.

QUICKSILVER DEPOSITS

Three quicksilver deposits of minor importance are known in the southern part of the Hot Springs Range within the Osgood Mountains quadrangle (Bailey and Phoenix, 1944, p. 90–91). Other deposits in the northern part of the Hot Springs Range have also been dedescribed by Bailey and Phoenix (1944, p. 101–106) and by Willden (in 1963).

The deposits are small, and as of 1955 only one, the Dutch Flat mine, had produced more than a token amount of quicksilver. One other, the K and K mine, is a potential small producer. The third, the Last Chance mine, appears to be no more than a poor prospect. The average grade of the ore at all these properties is low, probably less than 10 pounds of quicksilver per ton of ore and more nearly on the order of 5 to 6 pounds per ton, although small shoots and highly restricted pockets of higher grade material have been found in the Dutch Flat and K and K mines.

The deposits are in fractured altered feldspathic sandstone and shale of the Harmony formation. Cinnabar, the only ore mineral of mercury that has been recognized in the deposits, fills fractures and also occupies spaces between mineral grains in sandstone. In some fine-grained rocks disseminated subhedral cinnabar crystals that are several times larger than the average grain size of the rock may have been deposited by replacement. The gangue minerals of the altered host rocks are the quartz and sericite and minor amounts of vein quartz, pyrite, calcite, limonitic iron oxide, and jarosite. Paragenetic relations are not completely clear and the entire sequence of deposition is not represented in all the deposits. Cinnabar is later than quartz and pyrite, for cinnabar fills fractures that contain brecciated quartz with inclusions of pyrite. There is no evidence that any of the quartz succeeded cinnabar. Calcite was introduced after cinnabar, as it cements breccia zones containing cinnabar, and also veins and to a minor extent replaces cinnabar, and encrusts the walls of fractures and small cavities. Weathering processes caused oxidation of pyrite and stained the rocks with iron oxide and in places formed jarosite.

Although the rocks that contain the cinnabar have been altered, the alteration is not necessarily closely related to cinnabar mineralization, for at most places the altered rocks are associated with gold-bearing quartz veins and contain no cinnabar. Mostly, the alteration has a direct spatial relation to the granodiorite stock at Dutch Flat, and is possibly a late-stage phenomenon of granitic intrusion in the region.

GOLD-BEARING QUARTZ VEINS IN THE HOT SPRINGS RANGE

Near Dutch Flat on the western side of the Hot Springs Range the small stock of granodiorite and the surrounding sedimentary rocks of the Harmony formation are cut by gold-bearing quartz veins on which prospects and a few small mines have been located.

The gold deposits are in bodies of quartz in the granodiorite stock, on its periphery at the contact with the sedimentary rocks, and in the sedimentary rocks beyond the stock for approximately 2,000 feet northward and southward. The quartz bodies range from veinlets to well-defined veins and occur in two ways: (1) in the sedimentary rocks along faults or fault zones that strike north-northeast to northeast and dip moderately to steeply northwest or southeast and (2) in the granitic rock in fractures that strike north-northeast to north, dip gently to moderately west, and are parallel to a joint system in the granitic stock. Some of the prospects are on faulted zones with no quartz, but the larger properties were all developed on quartz bodies.

Most of the quartz veins are from 3 inches to 2 feet thick, and veinlets and stringers that range from less than 1 inch to 3 inches in thickness are plentiful. Veins as much as 5 to 8 feet thick are unusual, but were seen in a few places. The veins vary considerably in thickness and individual bodies are seldom continuous for long distances, but several discontinuous bodies may occupy a zone and constitute a minable "vein." The veins are parallel to the fractures that they occupy and obviously their emplacement was controlled by earlier fracturing. The veins have sharp clean walls, show all the features indicative of fracture filling, and give no evidence of replacement origin. Commonly they are brecciated and may be displaced by later faults. The quartz is white, except where stained by brown and yellow iron oxide, and commonly is more or less vuggy. Small clear terminated prisms of quartz project into the cavities. Some specimens show comb structure. Only a few inclusions of wallrock in quartz were seen.

The wallrocks adjacent to the fractures and veins are more or less altered, whether they are granodiorite, sedimentary rocks of the Harmony formation, or hornfels formed from the sedimentary rocks at the granodiorite contact. Although the stock at Dutch Flat is somewhat altered everywhere, alteration is noticeably more intense where the granodiorite is fractured and veined by quartz. In the sedimentary rocks and hornfels, alteration is most intense close to the quartz veins and to faults that have no visible quartz. The effects of the alteration are essentially the same in the granitic or sedimentary rocks and have resulted in the formation of sericite and kaolinite. The alteration effects are described in more detail on pages 46–49.

Samples of vein material from three of the properties were assayed for gold and silver,[22] and values ranging from a trace to 0.05 ounce of gold per ton and 0.3 to 1.5 ounces of silver per ton were reported. The assays indicate a ratio of gold to silver of 1 to 3; these samples cannot be considered representative of the ore mined, but no other data on grade of the ore are available. Small stopes in some of the mines indicate that parts of some veins contained small relatively high-grade ore shoots. Presumably the gold is free and occurs in the quartz; no gold-bearing minerals have been recognized. A specimen of quartz containing visible gold in limonite-filled fractures was found by one of us on the dump of a shallow prospect shaft on the ridge southwest of the granodiorite stock. No silver minerals were recognized, and probably the silver is contained in galena.

The quartz contains very minor amounts of the following metallic sulfide minerals: stibnite, galena, chalcocite(?), pyrite, sphalerite, and chalcopyrite. In addition veins in the mines at locations 4 and 5 (fig. 15) contain a few grains of scheelite at widely scattered places. Mostly, the sulfides are sparsely scattered through the quartz as individual grains 1 to 3 mm in maximum dimension, less commonly as irregular shaped bodies as much as 20 mm in size. Some sulfides occupy cavities in the quartz, and sulfides also occur along fractures. Stibnite is one of the commonest sulfides, and several large pieces of stibnite were found on the dumps, but none of comparable size were seen underground. Galena and light-brown sphalerite occur together. A gray metallic mineral uncertainly identified as chalcocite is fairly common but occurs as exceedingly minute individual grains. Pyrite is also fairly common but seemingly not more plentiful than the other sulfides, and chalcopyrite is rare. The paragenetic relations of the sulfides to one another could not be satisfactorily determined, but all are later than the quartz.

Distribution of the veins strongly suggests that they are genetically related to the granodiorite stock at Dutch Flat. The veins cut the granitic rock as well as the metamporphosed rocks that border the stock, which suggests the veins were formed after the stock was emplaced and had cooled sufficiently to solidify and maintain open fractures. The character of the veins and their accompanying minerals are indicative of intermediate or mesothermal conditions of formation, though the association of stibnite seems to mean that they were formed in an environment that approached the epithermal type. None of the veins is known to cut the closely associated volcanic rocks of Tertiary age, and the volcanic rocks are not affected by the alteration that modified the sedimentary rocks and intrusive body, which is further evidence that the veins were formed earlier than the volcanic episode. The volcanic rocks rest on a surface that is only about 100 feet higher than the minimum altitude at which the roof of the stock may have been, and therefore it is assumed that a considerable thickness of rock which contained the veins was removed by erosion, for surely the stock was emplaced several hundred, possibly more than a thousand, feet below the surface.

GOLD-SCHEELITE-CINNABAR PLACER

Placer deposits at Dutch Flat on the west side of the Hot Springs Range (fig. 15) contain significant amounts of gold, scheelite, and cinnabar. The following description has been abstracted from an account published elsewhere by Willden and Hotz (1955). Placer gold was discovered in the area by F. G. Wendell in 1893 (Vanderburg, 1936, p. 94); cinnabar probably was also noted in the early placer operations. The original discovery of scheelite in the placer is not known, but scheelite was recovered along with cinna-

[22] Assays for the authors by Union Assay Office, Inc., Salt Lake City, Utah.

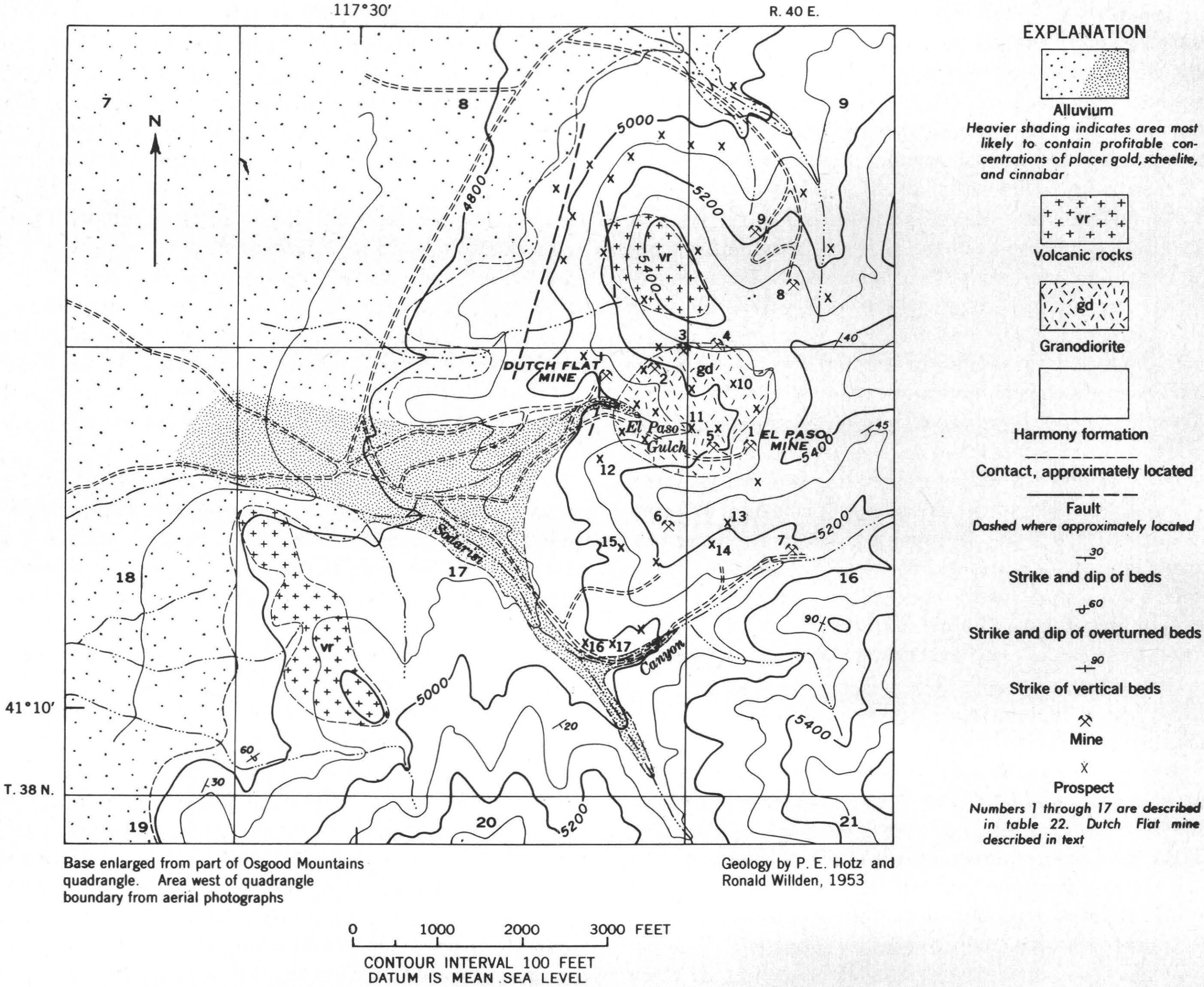


FIGURE 15.—Map of the Dutch Flat district.

bar and gold in a small washing plant in 1952. The only published production figures are those given by Vanderburg (1936, p. 94) and Smith and Vanderburg (1932, p. 55) for placer gold. Gold production in the first year after discovery was valued at about $75,000 and the total placer production in the district at about $200,000, nearly all of which was recovered by working with rockers (Vanderburg, 1936, p. 94).

Two types of placer deposits are present in the areas: stream deposits containing water-transported material, and slope-wash deposits containing material weathered from rocks in the immediate vicinity. The stream-transported material occurs in the bottoms of the larger canyons and on the alluvial fan below the mouth of Sodarisi Canyon (fig. 15). The stream deposits range in depth from 10 feet in the upper parts of the canyons to 35 feet at the mouth of Sodarisi Canyon, with an average depth of about 20 feet. The depth of the alluvial fan below the mouth of the canyon is known from a few scattered prospect pits to range from 12 to 90 feet.

The slope-wash deposits are found on both sides of El Paso Gulch to a height of as much as 50 feet above the creek bottom. The slope-wash deposits, as the term is used here, includes some stream-transported material which was left on the sides of the canyons as the streams cut down through bedrock. Some high assays reported from the slope-wash material are probably due to such accumulation of placer material from the stream channels. The slope-wash deposits range in depth from 5 to about 25 feet, with an average vertical depth to bedrock of 12 feet.

Sampling of the placer deposits has been sporadic and in general inadequate to delimit channels or pay streaks. Smith and Vanderburg (1932) report that the placer covers an area 8,000 feet long and 300 to 2,000 feet wide. Twenty acres of this ground sampled in 1904 (Vanderburg, 1936, p. 94) gave an average value in gold of 31 cents per cubic yard based on a price of $20.67 per ounce. Assays of the placer material give values of gold ranging from 20 cents to more than $1.00 per yard (at $35 per oz.), mercury ranging from 12 to 60 cents per yard based on the price of mercury in March 1954 ($185 per flask), and scheelite ranging from 60 cents to $2 per yard based on a price of $39 per 20-pound unit of WO_3 (Willden and Hotz, 1955, p. 666). Don B. Sebastian of San Francisco, Calif., had the property under lease in 1954. Preliminary samples of the alluvial fan and lower stream gravels, covering an area of about 50,000 square yards, indicate an average grade of $1.50 in combined gold, mercury, and tungsten based on the prices of the metals in August 1954 (mercury at $290 per flask) and a 60-percent WO_3 concentrate (Willden and Hotz, 1955, p. 666). Values up to $30 per yard for some of the slope-wash material have been reported, but such high values are probably a result of sporadic nuggets in the sample.

Some of the scheelite is in quartz fragments, particularly in the slope-wash material. This association is regarded as an indication that at least part of the scheelite was derived from erosion of quartz veins, even though the quartz veins exposed in the mine workings carry very little scheelite. Calcareous shale and sandy limestone beds occur in the Harmony formation in the area, and possibly contact-metamorphic tungsten deposits in these calcareous beds were the source of part of the scheelite in the placer. A small amount of garnet in the placer lends some support to this idea, but no contact-metamorphic tungsten deposits are exposed in this area.

The cinnabar grains range in size from minute dust to nuggets about 1 cm in diameter. The larger grains are not single crystals of cinnabar but consist of aggregates of cinnabar, quartz, and feldspar. Cinnabar commonly makes up 70 to 80 percent by volume of the aggregates. It shows a closer correlation of grain size with distance of transport than either scheelite or gold. The larger grains are found only in the upper part of El Paso Gulch, whereas the material from the alluvial fan is rarely larger than 0.5 mm and is usually very fine grained, with average grain size smaller than 0.02 mm. The shear zone of the Dutch Flat mine containing cinnabar provides an adequate source for the cinnabar found in the placer.

The gold in the placer occurs as coarse angular grains, which are sometimes attached to a quartz matrix, and as fine angular free grains. Vanderburg (1936, p. 96) reported that the purity of the gold is 940 fine and states that the largest nugget found had a value of $180. A sample of gold from concentrates of slope-wash material assayed by C. W. Hammond (written communication 1955) of the Nevada Mining Analytical Laboratory was 954 fine. The samples of placer material from the fan that were examined contained no gold. Of the many quartz veins known in the Dutch Flat district, those that were assayed contained some gold, and rich pockets of ore have been reported in some of the mines, so there is an adequate source within the district for the gold in the placer.

Scheelite occurs as nearly equant grains that range in size from 0.05 mm to 1 cm. The larger grains are found in the slope-wash material from the sides of El Paso Gulch and from stream deposits in El Paso Gulch above the junction with Sodarisi Canyon. Most scheelite grains in samples of the upper part of the alluvial fan are not more than 1 mm in size, and their average size is about 0.5 mm. The scheelite fluoresces blue white and is presumed to be free of molybdenum.

Several other minerals are associated with the gold, scheelite, and cinnabar, especially in the fine size fractions of the placer concentrates. These are magnetite, ilmenite, hematite, zircon, and pyrite, with small amounts of hornblende, monazite, garnet, and augite. These minerals are scarce in the size fractions of concentrates larger than about 1 mm, but they become increasingly abundant in the fine-size fractions, until in the size fraction smaller than 0.1 mm, only 25 percent of the total heavy minerals consist of gold, scheelite, and cinnabar.

The concentration of the minerals is largely due to the proximity of the placer deposits to the sources of the placer minerals. Greater distance of transportation of the minerals would result in progressive loss of cinnabar and, to a lesser extent, scheelite due to comminution of these brittle minerals by abrasive action during transport in the stream.

MINOR DEPOSITS OF LEAD AND COPPER

In addition to the minor amounts of copper, lead, and zinc that are associated with the tactite at many places, two small lead deposits and showings of copper mineralization of little or no economic importance are known in the Osgood Mountains and the Hot Springs Range. All are in Paleozoic sedimentary rocks and, except for a lead deposit at the Richmond mine, relatively far from exposures of granitic rock.

LEAD DEPOSITS

The only deposit that is reported to have had production is at the Richmond mine on the northwest side of the Osgood Mountains. Small seams and bunches of galena accompanied by sphalerite, pyrite, chalcopyrite, and considerable scheelite occur in metamorphosed limestone and tactite close to the granodiorite contact. According to Hess and Larsen (1921, p. 303) the mine was operated for many years as a silver mine with a small production. Presumably the silver was in galena, because no other silver minerals have been recognized. The sulfide mineralization seems to have been an episode that was related to the deposition of scheelite, although apparently galena and sphalerite were formed later than the scheelite.

In contrast, a small deposit of galena with subordinate amounts of sphalerite at the Silver Hill prospect at the south end of the Hot Springs Range is far from any granodiorite intrusive body. This deposit is in a fault zone in sedimentary rocks of the Harmony formation, and the sulfide minerals fill fractures and spaces between breccia fragments in the country rock. Supergene alteration of the galena and sphalerite and secondary deposition have formed some anglesite, cerussite, and hemimorphite, which accompany the primary minerals.

COPPER DEPOSITS

Some showings of copper mineralization in the Osgood Mountains have been explored by small pits and short adits where green stains in the rocks have attracted the attention of prospectors. The deposits are in calcareous and dolomitic rocks of various formations of Paleozoic age, and are generally several thousand feet beyond the granodiorite stock. None are known in the Osgood Mountain quartzite.

All the deposits consist of small seams and irregular bunches of chalcopyrite, usually with some pyrite, and some are associated with small amounts of galena. Vein quartz may or may not be present. Secondary copper minerals including malachite, chrysocolla, and rarely azurite, and green stains of indefinite mineral composition are also common.

BARITE DEPOSITS IN THE OSGOOD MOUNTAINS

Three small barite deposits are known in the Osgood Mountains. The largest is on the southeast side of the range, near the mouth of Hogshead Canyon. Barite is also exposed in a prospect pit approximately 2 miles to the northeast, near the Blue Bell mine at the mouth of Felix Canyon. A small deposit is on the northwest flank of the range north of Anderson Canyon. None of the deposits is known to have had commercial production of barite, and only the one at the mouth of Hogshead Canyon appears to be a potentially commercial deposit.

The deposits are geologically similar in several respects. They are relatively far from the granodiorite stock, ranging from 6,000 feet to as much as 15,000 feet from the nearest granodiorite contact. They replace dolomite and limestone chiefly, and chert to a lesser extent, in strata of Cambrian and Ordovician age; none are known in younger strata. Except for a little vein quartz locally, the barite seems to be unaccompanied by other ore or gangue minerals. Mostly the barite is dark gray, but some is gray to light gray. Commonly it is medium grained and equigranular, but it also occurs as globular or nodular concretions which have an internal radiating coarsely fibrous structure. Some aggregations of tabular crystals have crested surfaces where the crystals terminate on the outside of the nodules. The globular forms range from a few millimeters to as much as several centimeters in diameter; in places a carbonate bed may contain many scattered small separate globules of radiating barite.

The age of the deposits cannot be clearly established. They are certainly later than the strata of Cambrian and Ordovician age which they replace. The small deposit near the Blue Bell mine is in an area where there has been some metamorphism and copper metallization that are probably related to the emplacement of granodiorite, but the relations of the barite to the copper and metamorphism are not known. Gianella (1940) has described similar barite deposits in Lander and Eureka Counties which he regards as of hydrothermal origin, and he suggests that they may range from Jurassic to late Tertiary in age.

SILICA

The Valmy formation west of Stone Corral contains a quartzite unit which consists of quartz in a silica cement. The only deleterious constituents observed are traces of colorless mica and occasional minute grains of apatite and zircon. A chemical analysis of a random chip sample of the quartzite exposed in the low ridge west of Stone Corral is presented below:

SiO_2	99.6	TiO_2	0.01
Al_2O_3	.20	P_2O_5	.00
Fe_2O_3	.08	MnO	.01
FeO	.04	H_2O	.10
MgO	.03	CO_2	.05
CaO	.09	Sum	100.24
Na_2O	.01	Sp gr (lump)	2.57
K_2O	.02	Sp gr (powder)	2.64

[Analyzed by methods similar to those described in U.S. Geol. Bull. 1036–C. Analysis by U. S. Geol. Survey]

The quartzite bed is about 150 feet thick and is well exposed for approximately 3,000 feet along the strike. The exposure stands as much as 50 feet above the general ground level, is close to the east side of the low ridge, and could be quarried with relative ease. It is readily accessible by 10 miles of unpaved road from Golconda.

MINING FUTURE OF THE REGION

The main mineral resources that are known in the Osgood Mountains quadrangle are the deposits of tungsten ore and the gold-arsenic reserves in the Osgood Mountains.

The reserves of tungsten ore contained in the Osgood Mountains scheelite deposits have been considerably reduced by the high production since 1951, and probably most if not all of the ore that was readily obtained by open-pit methods is gone. There remain, however, important known deposits that can be mined underground at the Pacific, Riley, Riley Extension, and Tonopah mines. From the available data it does not seem that proximity of the Getchell fault zone to these mines will limit the downward continuation of these tactite deposits. The dips of both the contact and the fault are in the same direction, and it appears that the fault will not intersect the contact until much greater depths, if at all. Irregularities in the granodiorite contact, which can be defined by exploratory drilling, are probably the most important factors in determining the extent of known deposits or in predicting new ore bodies. At the Granite Creek mine considerable ore remained to be mined when the deposit was studied, but it was fairly clear that the bottom of the main deposit had been reached.

Exploration of the underground continuation of tactites mined at the surface in the Tip Top pits west of the Granite Creek mine may be profitable, and the steep slope and depth of the canyon should make these deposits exploitable by relatively short, direct adits. Similarly, additional reserves may lie beneath Pit 2 at the Kirby mine.

The possibility of finding new scheelite-bearing tactite bodies at the surface where limestone is in contact with the granodiorite seems remote, for the area has been rather thoroughly prospected. Tactite float is known, however, at a few places in the canyon of Osgood Creek, and possibly warrants some investigation (see p. 112).

Exploitation of the gold deposit at the Getchell mine depends on the ability of an operator profitably to mine and extract gold from a low-grade ore whose high arsenic content presents a difficult metallurgical problem. No other deposits of comparable size have been found along the Getchell fault zone, although there has been considerable exploration along it by the Getchell company. Possibly the Getchell deposit, like other epithermal precious metal deposits, will become of lower grade with depth and hence completely uneconomic.

It seems unlikely that any important production can be expected from the deposits of quicksilver in the Hot Springs Range, but the gold-scheelite-cinnabar placer at Dutch Flat might yield a moderate amount of quicksilver and tungsten as well as gold. Probably no more than token amounts of galena and sphalerite will be produced from the few small deposits of these minerals. Barite and possibly silica may eventually be produced from the area in modest quantities.

MINES AND PROSPECTS

TUNGSTEN DEPOSITS

The tungsten deposits of the Osgood Mountains are, with a single exception, in metamorphic rocks on the contact of a stock of granodiorite in the northern part of the range. A small, economically unimportant deposit occurs within altered granodiorite in the central part of the northern lobe of the stock. The location of the deposits is shown on figure 16. Most of the deposits and the most productive mines are distributed intermittently along the east contact of the northern and southern lobes of the stock; a large deposit and some less important ore bodies are on the south contact of the southern lobe; some small deposits are on the west contact of both lobes.

HISTORY AND PRODUCTION

Scheelite was discovered in the Osgood Mountains in about 1916 (Hess and Larsen, 1921, p. 247) after the attention of prospectors had been directed toward contact-metamorphic scheelite deposits by the high prices paid for tungsten at that time and the establishment of tungsten mills near Lovelock, Nev., and Bishop, Calif. When Hess and Larsen made their reconnaissance examination of contact-metamorphic scheelite deposits in 1917 and 1918, several tactites in the Osgood Mountains had been prospected for scheelite, and claims had been located, but no ore had been produced (Hess and Larsen, 1921, p. 301–304). Some of the deposits were on claims that had previously been prospected for copper and silver.

Active development and production of tungsten ore from mines in the Osgood Mountains began in the summer and fall of 1942 during the general stimulation of production, increased prices, and a Federal purchase program brought about by increased demands for tungsten during World War II. The principal mines were developed and mined by Getchell Mine, Inc., and U.S. Vanadium Corp. In 1942 the Getchell company remodeled its gold mill to enable milling and concentration of scheelite ores coming from their mines, and to treat ores from other properties on a custom basis. In 1945 a mill was erected at the Riley mine by the Northern Nevada Mining Corp., but only a small amount of ore was treated until the mine and mill were purchased by U.S. Vanadium Corp. later in the same year. Active development and mining continued in the district until the summer of 1945, when the demand for high domestic production of tungsten was lessened with the termination of World War II. All except the Riley mine had ceased operation in July 1945, but the Getchell mill continued to process stockpiled ore into 1946. A small production of scheelite from the Riley mine was maintained until early 1948.

From 1948 through 1950 the scheelite mines in the Osgood Mountains were idle. Then, in 1951 mining was resumed in response to a Federal purchase program, which raised the price of tungsten concentrates to $63 per unit of WO_3 and was designed to stimulate domestic production of tungsten to supply the rapidly expanding consumption consequent on the increased use of tungsten carbide, especially for military purposes. Full-scale development and mining of the tactite deposits was still going on in 1954 and 1955 when they were examined by the U.S. Geological Survey. At that time Getchell Mine, Inc., owned and operated four of the producing mines, one was operated by U.S. Vanadium Corp. (later the Union Carbide Nuclear Co.), and two belonged to independent operators. All the ore mined in the district was milled and concentrated at the Getchell mill, the mill that was built at the Riley mine in 1945 having been previously dismantled.

Table 20 is the recorded production from the Osgood Mountains tungsten deposits.

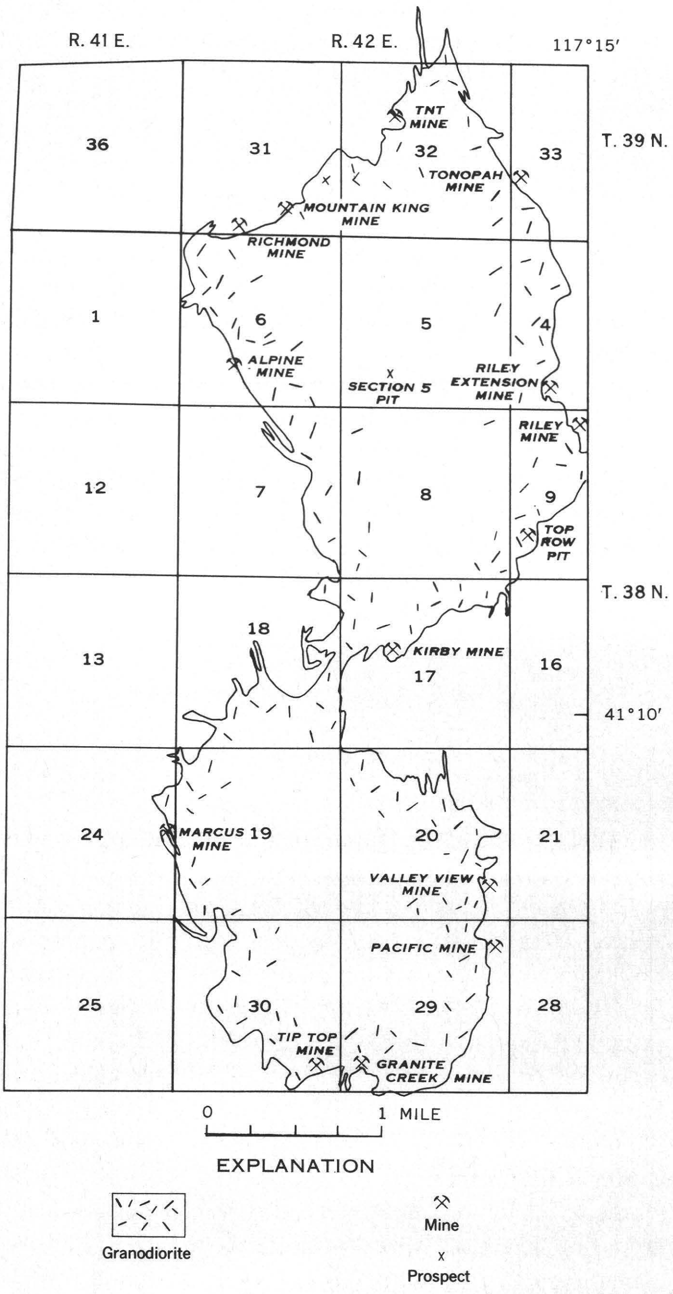


FIGURE 16.—Index map of the Osgood Mountains tungsten deposits.

GRANITE CREEK MINE

Tactite has been mined at several places in Granite Creek along the south boundary of the Osgood Mountains granodiorite stock, on the steep south side of the canyon where the intrusive is in contact with limestone of the Preble formation. Most of the production has been from the Granite Creek mine; less important deposits west of the Granite Creek mine have been mined from surface pits known collectively as the Tip Top mine.

The Granite Creek mine is in the SW¼ sec. 29, T. 38 N., R. 42 E., (fig. 16) approximately one-half mile above the mouth of Granite Creek. The mine is easily accessible by a good unpaved road.

The Granite Creek mine, owned by Getchell Mine, Inc., was first active in the fall of 1942 and continued in operation until the summer of 1944. A crosscut was driven on the 300 level to the tactite (pl. 3) and a drift was run along the strike (Newman,[23] 1955, p, 16), from which the tactite zone was stoped to the surface. Production during the period of slightly more than 2 years amounted to 88,000 tons of ore averaging 0.5 percent

[23] Newman, W. J., 1955, Tungsten mining at the Getchell mine: Nevada Univ., unpublished Mining Engineer thesis, 43 p., 5 figs., 3 tables.

WO_3 (Hobbs and Clabaugh, 1946, p. 15). Mining was resumed in December 1950, and in 1951 a drift 1,700 feet long with the portal at the canyon mouth was driven on the 500 level to intersect the ore body below the 300 level (pls. 2, 3). An intermediate level (400 level) was then driven off raises from the 500 level (pl. 3) and the ore body was stoped up to the 300 level (Newman, 1955, p. 19–20). In 1955 mining was being actively pursued at all levels, and production for the 4½ years from December 1950 through June 1955 exceeded 188,000 tons. Total production from 1942 was 276,000 tons.

The country rock at the Granite Creek mine (pl. 2) is limestone of the Preble formation which has been intruded by granodiorite. Mostly the limestone strikes from N. 40° E. to N. 60° E. and dips from 35° to 60° southeast. Locally it departs from this general attitude and in places has been squeezed and tightly folded on a small scale, but no major folds have been recognized. Some minor faults were observed underground in the western workings on level 300 that dip steeply east or southeast along the limestone-granodiorite contact. In the mine the limestone is metamorphosed to calc-silicate rock, marble, and tactite. Tactite occurs along only part of the granodiorite-limestone contact, an indication that special conditions were necessary for its formation. The distribution of scheelite is almost entirely restricted to the tactite, though minor amounts are found in silicated limestone near the granodiorite contact.

The gross trend of the granodiorite contact in the Granite Creek mine area is easterly, and the contact dips steeply south; but in detail it has irregularities and sends off dikelike apophyses southward into the country rock. The main scheelite-bearing tactite body at the Granite Creek mine is at the contact between limestone and granodiorite in the angle of a reentrant that is formed by a narrow granodiorite offshoot on the west and broader protrusion on the east. Between the surface and the 300 level the granodiorite contact is

TABLE 20.—*Production of tungsten ore, Osgood Mountains*

Production, 1943–1945, Hobbs and Clabaugh, 1946, p. 14–27; Getchell-owned properties and Riley mine, 1950–1955, from owners; Richmond mine, 1954, Valley View mine, 1951–1955, Mountain King and T.N.T. mine, compiled by U.S. Bur. Mines and published with permission of owners

Property	Year	Tons of ore	Avg. grade, in percent WO_3	Contained units of WO_3
Getchell Mine, Inc:				
Tip Top mine	1954–55	37, 000	[1] (0.3)	11, 100
Granite Creek mine	1942–45	88, 000	. 5	44, 000
	December 1950–June 1955	188, 000	. 5	94, 000
Pacific mine	February 1952–June 1955	226, 700	. 3	68, 010
Kirby mine	1942–43	32, 000	. 43	13, 760
	1952–53	34, 550	. 43	13, 820
Top Row pit	1954	2, 800	(. 3)	840
Riley Extension mine	1951–June 1955	188, 800	(. 3)	56, 640
Tonopah mine	1950–June 1955	19, 750	(. 3)	5, 925
Alpine (Porvenir) mine	1943	8, 000	. 5	4, 000
	1950–53	43, 700	(. 3)	13, 110
Total		869, 300		325, 205
Union Carbide Nuclear Co: (Formerly U.S. Vanadium Corp.)				
Riley mine	1943–44	80, 000	. 63	50, 400
	1945	8, 500	. 70	5, 950
	1946–48			
	1952–June 1955	264, 600	. 57	150, 822
Richmond mine	1942–43	31, 500	. 5	15, 750
	1954	2, 800	. 2	560
Total		387, 400		223, 482
Valley View mine	1943–44	1, 500	(. 5)	750
	1951–55	67, 300	. 5	33, 650
Total		68, 800		34, 400
T.N.T., Inc.:				
Mountain King and T.N.T. mines	1951–55	18, 650	. 6	11, 190
Total		18, 650		11, 190
Total, all mines		1, 344, 150		594, 277

[1] Figures in parentheses are estimated average grade.

troughlike, with the keel plunging approximately 50° S. 13°E.; the trough is slightly overturned to the west, so that on the eastern side granodiorite overlies the tactite and metamorphosed limestone. The reentrant angle persists down to the 400 level, but the eastern side of the trough is very short. On the 500 level the contact, though somewhat concave southward, no longer has the reentrant angle and troughlike form. Development and drill-hole exploration underground indicate that the limestone in the reentrant is isolated from the main body of limestone by a narrow east-west granodiorite dike that cuts through the limestone a short distance south of the mine workings and is continuous with the granodiorite on each side of the reentrant.

When the mine was examined in 1955, it appeared that the limestone was a downward-tapering roof pendant whose root lay not far below the 500 level.

The form of the tactite deposit can be most clearly visualized from the map and sections through the mine shown in plate 3. The main tactite body was continuous from the surface to the 500 level, and lay in the keel and lower part of the troughlike structure. It varied considerably in width, ranging from 5 to 25 feet and averaging between 15 and 20 feet above the 400 level and approximately 10 feet below that level. It was most persistent in depth on the west side of the trough, where it was followed continuously from the surface to the 500 level, though between the 400 and 500 levels its length became considerably shorter. It was longest on the 300 level, where it lay on both sides of the keel, and tapered towards the surface and in depth. The tactite body exposed in the 500 level workings is small and of irregular shape.

The tactite parallels the granodiorite contact rather closely and tends to be most persistent and wider where the strike of the limestone and the granodiorite contact are parallel, although tactite was also formed on the 300 level, where the contact is about perpendicular to the strike of the beds. The tactite terminates along strike by pinching or ending abruptly in blunt tongues or slightly interfingering contacts against calc-silicate rock and marble. In most places there is no obvious geologic reason for the termination, such as a sharp change in strike and dip of the granodiorite contact, change in lithology, or control by faulting.

Underground drill-hole exploration of the block southeast of the mine workings between the 300 and 500 levels showed that some tactite containing scheelite occurs at the contact of limestone with an offshoot of granodiorite. The body probably is continuous with the tactite in the drift at the end of the southeast crosscut on the 300 level (pl. 3) and is present approximately 130 feet southeast of the 400 level. The size and extent of the body is virtually unknown.

Exploration southwest of the mine workings failed to discover tactite along the contact with the long southward-projecting dike of granodiorite on the 300 level. A short crosscut and drift west of the southwest drift on the 300 level exposed a small body of tactite that is in a position similar to that occupied by a small mass of tactite exposed at the surface in the end of the reentrant of limestone between two granodiorite tongues west of the mine (pl. 2).

TIP TOP MINE

Discontinuous bodies of tactite occur along the granodiorite-limestone contact on the south side of Granite Creek for about 2,000 feet west of the Granite Creek mine. Except for a little prospecting, the deposits west of Granite Creek mine were not worked until the summer of 1954 when Getchell Mine, Inc., commenced open-pit mining, which continued until June 1955. About 37,000 tons of ore was produced from three open pits, the bulk coming from two of them. The average grade of the material mined is not known. The two larger pits are shown on plate 2; a third one is on the crest of the ridge southeast of the upper pit shown on plate 2. The workings, called collectively the Tip Top pits, are in SE¼SE¼ sec. 30, T. 38 N., R. 42 E.

The country rock is granodiorite, and limestone and hornfels of the Preble formation. The sedimentary rocks strike N. 50° E. to N. 70° E. and dip 50° to 60° SE. Locally there is some minor folding but no large-scale folds have been recognized. The granodiorite contact trends west with many slight local variations in strike, and with some south-trending dikelike offshoots, from the Granite Creek mine to the middle pit of the Tip Top mine (pl. 2). At the west end of the area shown on the map (pl. 2) the contact assumes a southwest trend which continues to the crest of the ridge (Hobbs and Clabaugh, 1946, pl. 2). In its large-scale relations then, the granodiorite contact cuts across the strike of the sedimentary rocks, but locally the strikes of both are more or less parallel. Observations of the contact in the open cuts show that it dips southeast to south at about 60° to 70°; where strike of the contact is parallel to the strike of the sedimentary rocks it dips more or less in conformity with the bedding.

The limestone is recrystallized to marble, calc-silicate rock, and tactite adjacent to the granodiorite. The tactite bodies are discontinuous, occurring only where the granodiorite intrudes limestone; the hornfels contacts are barren. Limestone is recrystallized to marble for as much as several hundred feet from the contact; calc-silicate rock occurs sporadically along the contact and is confined to a zone within a few feet of the contact. At a few places limestone and marble near the contact are complexly crumpled. The tactite bodies are confined to the granodiorite contact and everywhere are next to the granodiorite, between the intrusive rock and the limestone. The tactite bodies are tabular and as long as 400 feet. The largest of those exposed in the open pits are about 6 feet thick, and they average about 4 feet; however, the mined-out parts may have been wider.

The tactite is a brown garnet rock with accessory pyroxene and at places contains some coarsely crystalline calcite. The contact with granodiorite is sharp and clean, and that between tactite and limestone or its metamorphosed equivalents is also abrupt. At many places the granodiorite immediately adjacent to the contact is strongly impregnated with quartz, contains visible pyrite and chalcopyrite, is richer in dark minerals, and contains hornblende in addition to biotite. Narrow reaction zones and indefinite contacts between calc-silicate rock and pyroxene-bearing granitic rock are observable in places. In addition to scheelite, the tactite contains at places a little chalcopyrite and pyrite, and some is stained with green secondary copper minerals. The tungsten content of the tactite is not known.

The tactite bodies in the Tip Top pits occupy parts of the contact that are relatively smooth. Limestone adjacent to the tactite bodies is not obviously different from that adjacent to barren parts of the contact, but the area is not well enough exposed to eliminate entirely compositional or structural variations in the host rock as factors of possible importance. It may be significant that the important tactite bodies seem to be situated at positions along the contact, where it closely parallels the strike of bedding in the limestone.

MARCUS MINE

The Marcus mine is 2 miles northwest of the Granite Creek mine in NE¼SW¼ sec. 24, T. 38 N., R. 42 E. (pl. 1), at an altitude of approximately 7,250 feet. A road from the Granite Creek mine was built in 1955 by Getchell Mine, Inc. Before 1955, the Marcus mine had been explored by several trenches and open cuts and two short adits, but no ore had been produced. In July 1955, Getchell Mine, Inc., commenced open-pit mining and when the property was visited in 1955 a few hundred tons of ore had been produced.

The Marcus deposit is on the west contact of the southern lobe of the granodiorite stock. The rocks adjacent to the granodiorite are argillaceous hornfels and interbedded limestone of the Preble formation (pl. 1). At the Marcus mine a bed of limestone is partly enclosed between the main body and a northwestward-projecting tongue of granodiorite. Tactite containing scheelite has been formed in limestone at the east contact of the granodiorite apophysis. Beyond the tactite the limestone is marbleized and converted to calc-silicate rock.

The excavation made by Getchell Mine, Inc., was on a tactite body described by Hobbs and Clabaugh (1946, p. 27) as being 2 to 10 feet wide and at least 75 feet long. The tactite is a medium-grained garnet rock containing calcite, epidote, quartz, a little idocrase, and scheelite. Hobbs and Clabaugh (1946, p. 27) estimated the ore to contain approximately 1.7 percent WO_3. A tactite body that is exposed in a short adit downhill approximately 120 feet to the south may or may not be a continuation of the same body. Two small isolated exposures of tactite of lower grade, about 600 and 1,000 feet north of the upper deposit, were also reported by Hobbs and Clabaugh (1946, p. 27) but were not seen during this later study.

PACIFIC MINE

The Pacific mine is at the front of the Osgood Mountains approximately three-fourths of a mile north of the mouth of Granite Creek, in eastern part sec. 29, T. 38 N., R. 42 E. (fig. 16). An excellent graded dirt road from the property joins the main road to Getchell mine approximately 1 mile to the east, and also connects with the Granite Creek road to the south.

Before 1951 the Pacific mine was a prospect with no recorded production. Hobbs and Clabaugh (1946, p. 18–19) estimated a reserve of 40,000 tons of tactite containing 0.5 percent WO_3, and predicted that larger masses of tactite would be discovered underground. From 1951 through 1956 the Getchell Co. mined more than 279,000 tons of tungsten ore from an open pit at the Pacific mine (Newman, 1957, p. 91). Ore was mined underground from two levels, and by early 1957 stopes from the upper level had broken through into the floor of the open pit at two places. As of July 1955, more than 4,500 tons of ore had been mined by underground methods.

The Pacific deposit (fig. 17) is on the east contact of the southern lobe of the granodiorite stock. Limestone of the Preble formation that strikes north to N. 15° W., with an average dip eastward of about 40°, is intruded by granodiorite. The contact trends generally north and dips 55° to 60° E, possibly becoming steeper with increased depth. Little unmetamorphosed limestone is to be found at the Pacific mine, as most has been metamorphosed to calc-silicate rock composed of interlayered diopside-plagioclase hornfels, calcite-wollastonite rock, and some layers of marble. East of the carbonate unit is a zone of strongly crushed and sheared black hornfels, which is at least 600 feet wide where it is exposed in the adit of the lower or 300 level of the underground workings. The eastern part of the limestone and calc-silicate rock unit is also somewhat broken where it is exposed underground. The faulting is along a zone that seems to be en echelon with the southern extension of the Getchell fault zone.

The scheelite deposit is confined to tactite, and the tactite is at the contact between granodiorite and calc-silicate rock, but tactite is not found everywhere that granodiorite intrudes carbonate beds or their metamorphosed equivalent. The main tactite body is confined to a relatively short distance along the contact south of a hook-shaped southward projection of granodiorite which formed a trough that plunges steeply south. The tactite body was tabular, with its thickest

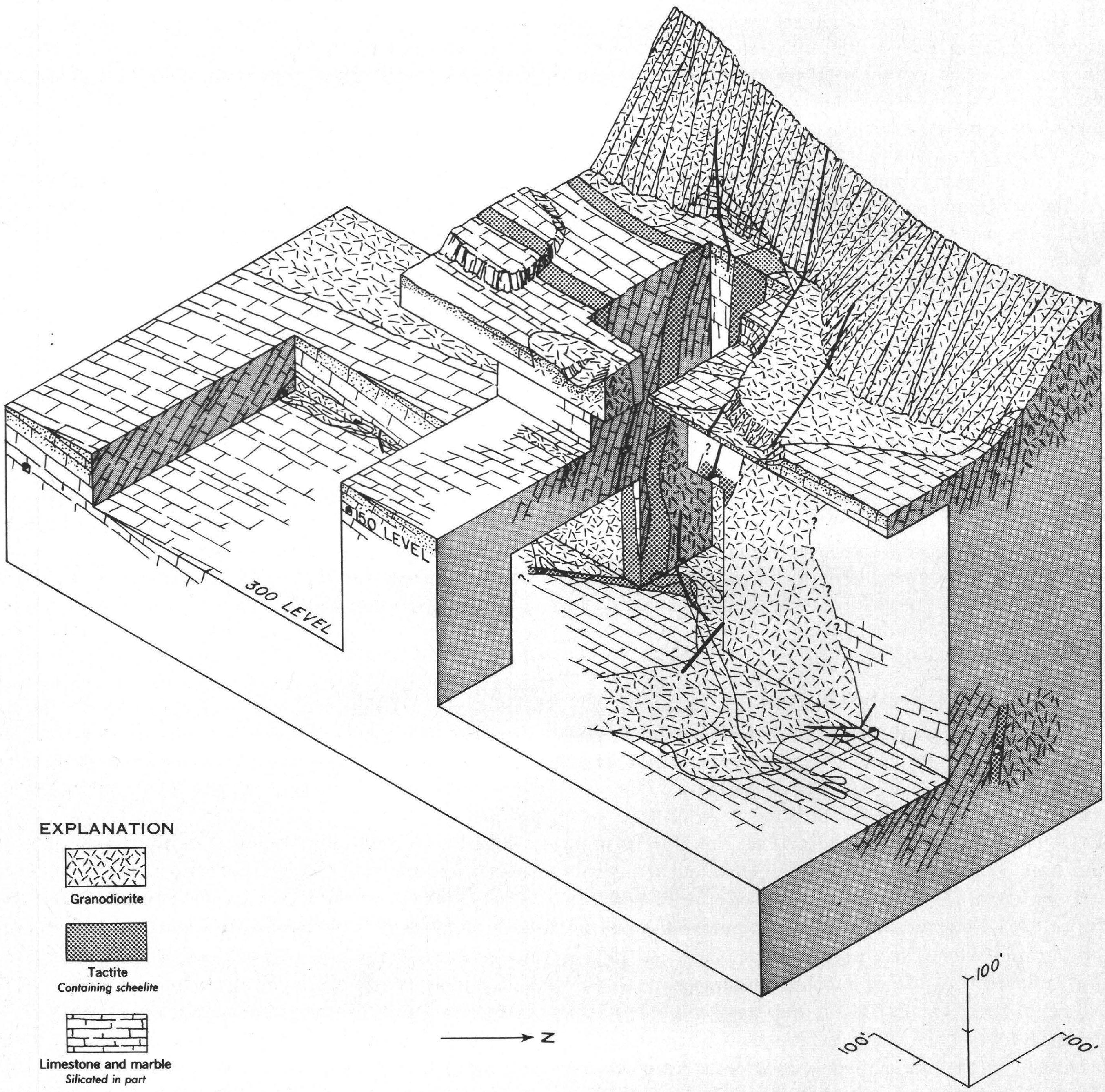


FIGURE 17.—Block diagram of the Pacific mine.

part near the trough, and became thinner and pinched out southward. It was mined down dip for approximately 550 feet to the lower level (300 level) of the underground workings; most of it was mined from an open pit. A smaller body of tactite was mined approximately 600 feet north of the main deposit where apparently another jog in the contact formed a trough that plunged steeply northeast. This body was mined out from the surface workings.

VALLEY VIEW MINE

The Valley View mine is in low hills at the east front of the Osgood Mountains in SE¼SE¼ sec. 20, T. 38 N., R. 42 E. (fig. 16), near the mouth of Bunch Creek, and joins the property of the Pacific mine on the north. A good unpaved road 1.4 miles long connects the property with the main road to Getchell mine approximately 5½ miles south of the Getchell Mill.

According to Hobbs and Clabaugh (1946, p. 24) the property was originally prospected for silver and copper before scheelite was recognized in the district. Harold's Club Mining Co. of Reno, Nev., leased the property in 1943 and commenced mining tungsten ore, which, beginning in July 1943, was shipped to the Metals Reserve Co. stockpile at Getchell mine. Ore production between the summers of 1943 and 1944 was from underground workings that included a 225-foot adit along the contact zone south from Bunch Creek and more extensive workings along the contact north from the creek. The property had a relatively small production of approximately 1,500 tons during this period (Hobbs and Clabaugh, 1946, p. 24). From 1944 to 1950 the property was idle, but in 1951 mining operations were resumed by the Valley View Mining Co., lessees, and 67,318 tons of ore whose average grade was 0.5 percent WO_3 was mined from 1951 through 1955. The ore was trucked to the Getchell mine, where it was milled on a custom basis. Total production from 1943 through 1955 was approximately 69,000 tons.

Between 1951 and 1955 some ore was mined from shallow open cuts in tactite along the contact south of Bunch Creek but the main operation was in the tactite north of the creek (pl. 4). Here the tactite was mined in a deep narrow open cut that was more than 250 feet deep in the summer of 1955; the tactite body has been mined for a depth of nearly 300 feet down the dip.

Country rocks at the Valley View mine are granodiorite, limestone, calc-silicate rocks, marble, hornfels, and a little biotite schist that have been formed by metamorphism of limestone and pelitic sedimentary rocks of the Preble formation by the granodiorite intrusive body. The tactite ore body is on or near the contact between granodiorite and limestone.

The tactite is like that elsewhere in the district, composed of reddish-brown garnet and dark-green pyroxene; some is massive and some is composed of alternate layers of pyroxene with relatively coarse garnet and narrower layers of fine-grained pyroxene with little or no garnet. Scheelite occurs as grains a millimeter or less in diameter in the tactite. In the wall next to the bottom of the pit the tactite is veined by gray greasy-looking quartz that encloses fragments of tactite and dark biotite-rich streaks. The quartz contains specks and veinlets of chalcopyrite and pyrite. In a crushed zone along the granodiorite-tactite contact in the north end of the pit, probably subsidiary to the fault along the west side of the open pit (pl. 4), tactite is altered to a porous crumbly rock composed of a soft dark-greenish-yellow to moderate-greenish-yellow material, identified as a mixture of nontronite and quartz. Presumably the nontronite was formed by alteration of the tactite, for it veins, encloses, and apparently replaces garnet-rich tactite. Some jasperoid with cavities coated by finely crystalline quartz is associated with the altered rock.

In the surface workings south of Bunch Creek a tactite layer is exposed between the granodiorite and limestone. In part, the surface workings reached to the depth of an old adit (not shown on pl. 4) that followed the layer southward. The layer averages 2½ to 3 feet in thickness and is exposed for approximately 175 feet at the surface. The main ore body was north of the creek, and as it was originally exposed at the surface it extended northward from near the portal of the north adit as a thin tabular body ranging from 2 to 7 feet in width for about 300 feet, where it expanded into an irregular-shaped body of mixed hornfels and tactite 200 feet long and 50 to 100 feet wide (Hobbs and Clabaugh, 1946, p. 25). The inset in plate 4 taken from Hobbs' and Clabaugh's report (1946, pl. 7), illustrates the shape and extent of the tactite at the surface before open-pit mining began. Judging from the original surface expression of the tactite and its present distribution in the pit, the wide northern end of the tactite body plunges northeastward at about 50°, becoming narrower with increasing depth. From the relations visible in 1955, the main tactite deposit that was mined from the open pit was separated from the footwall granodiorite for most of its length by a layer of calc-silicate rocks and marble 5 to 15 feet thick, and it was overlain by similar rocks. It was in contact with granodiorite only at its northern termination and south of the pit, where it narrowed from its maximum thickness. It is worth noting that the expanded part of the tactite is where the granodiorite contact makes an abrupt change in direction of strike from the somewhat irregular but generally northerly strike pursued in the southern part of the mine to a roughly easterly trend. The result is that the granodiorite contact changes from a generally concordant attitude with respect to the bedding of the sedimentary rocks to a strongly discordant relation, and the tactite was formed in a pocket partly enclosed by granodiorite.

Two conspicuous faults are exposed in the walls of the main open cut. They strike slightly west of south and dip steeply east. The west fault cuts through granodiorite in the north face of the open cut, where it strikes approximately N. 10° W. and dips 60° E. Southward it is on the central contact between limestone and a tongue of granodiorite east of the limestone and tactite. In the north adit a wide faulted zone that is probably a continuation of the west fault forms the east wall of the tactite layer (Hobbs and Clabaugh, 1946, p. 24). The tactite in the pit dips in the same direction as the fault but the angle of dip seems to be about 10° less than the dip of the fault, and therefore the fault may limit the downward extent of the tactite (pl. 4, section *A–A′*). Possibly this part of the fault follows the original intrusive contact of the granodiorite tongue, but even so, the tactite may be limited in depth by convergence of granodiorite from the hanging wall and footwall sides. The east fault separates limestone and hornfels on the east from the granodiorite on the west in the pit. Its strike is essentially parallel to that of the west fault, but it is nearly vertical or dips steeply east. It is interesting to speculate that if the fault on the east side of the pit intersects the granodiorite-tactite contact beneath the pit, there might be a down-faulted tactite body to the east at depth. Although the faults cannot be traced beyond their exposures in the open cut they are, judging from the strong crushing along them, probably persistent for considerable distances north and south of the mine. They are chiefly parallel to the strong zone of faulting along the east front of the range and it seems reasonable to suppose that they are related in age and origin.

North of the main open cut a blunt projection of granodiorite extends eastward across the north-trending sedimentary rocks. On the north side of the projecting granodiorite, where it is in contact with limestone, a zone of thin discontinuous tactite crops out for 200 to 250 feet. One of the tactite bodies is explored by an adit 110 feet long, and is approximately 3 to 8 feet wide. Hobbs and Clabaugh (1946, p. 25) estimated the tactite along this zone to be 3 to 10 feet wide and to have a low tungsten content of 0.3 percent to 0.4 percent WO_3.

KIRBY MINE

The Kirby mine includes three open-cut operations in NW¼ sec. 17, T. 38 N., R. 42 E. (fig. 16), in the valley of Julian Creek (also commonly known as Kirby Creek or Ranch Creek). Two of the open cuts are on the south side of the valley, and one is on the north side, near the bottom. They are easily accessible by 1¾ miles of good dirt road that joins the main road to Getchell mine near the Riley mine.

In 1917 the property, in part at least, was on claims held by Fayant and Blaine, and was briefly described by Hess and Larsen (1921, p. 301–302) as a potentially important scheelite prospect. The claims were subsequently relocated by Mr. Kirby and in 1942 the property was sold to Getchell Mine, Inc., which mined approximately 32,000 tons of tungsten ore containing more than 12,000 units of WO_3 from underground and open-pit workings in a deposit south of the creek at an altitude of about 6,300 feet. The ore contained an average of 0.43 percent WO_3 (Hobbs and Clabaugh, 1946, p. 17). Mining ceased in the summer of 1943 and the property lay idle until 1952, when the Getchell company resumed mining and developed two smaller open-cut operations on deposits that are, respectively, lower down the south side of the valley, and near the valley bottom north of the creek (pl. 5). The operations were closed down in 1953 after having produced 34,550 tons during this second period of operation. Total production from 1942 to 1952 was 66,550 tons.

The tactite deposits of the Kirby mine area are on the southeast contact of the northern lobe of the granodiorite stock, near the narrow central waist of the stock (fig. 16). The granodiorite intrudes shale and limestone of the Preble formation (pl. 5), which have been metamorphosed to hornfels, marble, and calc-silicate rock. Tactite is restricted to the contact between granodiorite and limestone and is a scheelite-bearing garnet-pyroxene rock like that found elsewhere in the district.

The average strike of the sedimentary rocks is N. 50° E., and they dip moderately southeast. The contact of the granodiorite trends generally northeast, also, but is irregular and strongly discordant locally. Dikelike apophyses extend out into the sedimentary rocks from the main granitic body. It is clear from the map (pl. 5) that localization of the important tactite bodies is related to abrupt bends in the contact, which form reentrants of calcareous rocks in the granodiorite.

The principal deposit of tactite, here called the No. 1 pit, was in the southwest part of the Kirby mine area where the granodiorite contact makes a right-angle turn. This deposit has been essentially mined out. A pendantlike body of tactite projected downward into the granodiorite (pl. 5, section *A–A′*) and tapered out westward beneath limestone that overlies the granodiorite. West of the No. 1 pit, thin tabular bodies of tactite probably not more than about 5 feet thick occupy the gentle eastward-dipping granodiorite-limestone contact. They are nearly parallel to the slope of the hill and therefore the wide prominent outcrops represent only small volumes of ore. An attempt was made to strip and mine these thin bodies of tactite in 1943, but the operation was unprofitable and was abandoned (Hobbs and Clabaugh, 1946, p. 18–19). Hobbs and Clabaugh (1946, p. 18) estimated

* * * 1,500 tons of measured ore containing 600 units of WO_3, 2,000 tons of indicated ore containing 900 units of WO_3 and 2,000 tons of inferred ore containing 800 units of WO_3

from the tactite bodies on this hillside.

A small tactite body is exposed in a small open cut approximately 450 feet southeast of the No. 1 pit. The deposit is situated in a shallow reentrant of calc-silicate rock and hornfels in granodiorite. The tactite has been nearly all mined out, and a small lenticular body of mixed tactite and calc-silicate rock remains. The country rock in the reentrant is mostly hornfels, and the volume of tactite probably is not large.

During the operations in 1952–53 an open-pit mine, here called the No. 2 pit, was developed on a fairly large body of tactite on the south slope of Kirby Creek at an altitude of about 6,100 feet. The tactite body is tabular and dips approximately 60° SE., parallel to the northeast granodiorite contact and discordant to the bedding of the limestone. Its widest part and northeast termination is where the granodiorite contact makes a short right-angle bend; here it is about 10 feet thick. It wedges out gradually southwestward along strike and ends below the top of the cut, a horizontal distance of about 110 feet and about the same distance vertically. A septum of hard epidote-rich granitoid rock separates the tactite from the granodiorite footwall. Limestone in the hanging wall has been converted to calc-silicate rock and marble. The granodiorite exposed in the lower east wall of the open cut contains many dark streaks rich in hornblende and biotite that are arranged subparallel to the contact with the overlying sedimentary rocks.

A tactite layer about 2 feet thick is in a small exposure of calc-silicate rock approximately 110 feet south of the upper rim of the No. 2 pit. Bedrock between here and the open pit is concealed by soil and colluvium, but probably this is an isolated body of tactite separate from the tactite in the open cut.

Tactite was also mined from a small open cut on the lower north slope of Kirby Creek, here called the No. 3 pit. This deposit was also situated at a right-angle bend in the granodiorite contact. It is almost mined out and only some thin remnants of the original body are left. The rocks intruded by the granodiorite are mainly hornfels with some calcareous interbeds, and limestone is subordinate. A little tactite has formed in the hornfels, presumably from interbedded calcareous layers, but the main deposit seems to have been a small body on the contact between granodiorite and limestone. The unreplaced limestone exposed in the pit is thick-bedded to massive marble. Some of the remaining tactite contains much glassy quartz, some of which is coarsely crystalline. Some of the quartz contains scheelite.

The tungsten resources of the Kirby mine area are definitely limited, and probably the major production from the area is past. Some tactite remains to be mined from the deposits west of the No. 1 open cut, but the amount is limited by the sheetlike form of the bodies, which are not expected to have much extension in depth. The steeply dipping tactite body exposed in No. 2 pit has a limited strike length. Its extension in depth is not known. Its position on the hill makes further open-pit mining expensive, and underground development may be required to mine the body profitably.

TOP ROW PIT

The Top Row pit is on the east side of the Osgood Mountains, approximately 0.25 mile south of Summer Camp Creek in SW1/4 sec. 9, T. 38 N., R. 42 E. (fig. 16). It is at an altitude of approximately 5,750 feet and is situated on the southeast side and near the top of a small knob about 600 feet west of the road from Riley mine to Julian Creek and the Kirby mine. A short steep road makes the mine accessible from the Julian Creek Road. In June 1954, Getchell Mine, Inc., opened a pit on a small tactite body and mined approximately 2,800 tons of tactite. Because the tactite body is small, further development has not been undertaken.

The tactite is on the east contact of the northern lobe of the granodiorite stock. Argillaceous hornfels is the principal rock type in this area, but scanty amounts of thin lenticular limestone are interbedded with the argillaceous rocks. A small body of limestone in a reentrant angle along the slightly irregular granodiorite contact was replaced by scheelite-bearing garnet-pyroxene tactite. A little tactite is left, probably less than 25 feet long and 5 feet wide at the surface; other calcareous interbeds have been converted to calc-silicate rock that contains no scheelite.

RILEY AND RILEY EXTENSION MINES

The Riley and Riley Extension mines are in the northern part of sec. 9 and southern part of sec 4, T. 38 N., R. 42 E. (fig. 16), respectively, at the east front of the Osgood Mountains between Hansen Canyon on the north and Summer Camp Creek on the south. Parts of the tactite deposit are east of the quadrangle boundary. Both properties are served by good dirt roads that join the main road to Getchell mine approximately one-fourth of a mile to the east.

The Riley mine, formerly known as the Dernan property, was prospected during World War I, but little if any tungsten was produced. J. E. Riley leased the property from heirs of the Tom Dernan estate in the fall of 1942, and developed and operated it as the Riley mine (Northern Nevada Mining Co.) until 1945. About 80,000 tons of ore averaging 0.63 percent WO_3 was mined from surface pits and sold to Metals Reserve Co. from June 1943 to July 1, 1944. A mill was constructed on the property and beginning in May 1945 about 8,500 tons of ore, reported to average 0.70 percent WO_3 ore, were treated up to July 31, 1945 (Holmes, 1946, p. 3). In 1945 the United States Bureau of Mines conducted a diamond-drill exploration of the Riley mine, testing the underground extent of the ore to depths of 500 feet down the dip from surface exposures, and greatly enlarged the known reserves (Hobbs and Clabaugh, 1946, p. 21; Holmes, 1946).

In October 1945, the U.S. Vanadium Corp. purchased the property and continued operation of the mine until 1948. Production during this period is not known, although in 1947 the mine was the second largest producer of tungsten concentrates in Nevada (Minerals Yearbook, 1947, p. 1191). The mine was idle from 1948 to August 1952, when mining was resumed by the same company. In 1955 the name of the property owners was changed to Union Carbide Nuclear Co. From 1952 to June 1955, approximately 264,600 tons averaging nearly 0.5 percent WO_3 was produced from combined surface and underground operations.

The northern extension of the Riley mine or body is on property owned by Getchell Mine, Inc., which began underground development and mining at the Riley Extension mine in 1951. By June 30, 1955, about 161,000 tons of ore had been produced and mining and further development was still being actively pursued. In November 1953 a small open-cut mine was started in another ore body on the ridge south of Hansen Canyon at an altitude of about 6,000 feet. The operation ended in the summer of 1954 after approximately 27,800 tons of tactite averaging 0.5 percent WO_3 had been produced.

Since mining started at the Riley mine in 1943, total combined production from the Riley and Riley Extension mines to June 1955 is estimated to have been more than 550,000 tons, not including the small open-cut operations at the Riley Extension mine. This, then, was the largest single tactite deposit in the district.

Before 1952, the Riley mine was entirely a surface operation, and the ore was mined from a series of open pits. Since August 1952, mining has been conducted underground and continued from surface pits as well (pl. 6). Underground development and mining (pl. 7) was initiated by driving an adit in the hanging wall on the 200 level, from which drifts were developed northward and southward on the ore. Tactite above the level was mined from shrinkage stopes. An inclined shaft was sunk down dip from the 200 level, and drifts were driven on three levels from which stopes were developed to mine the ore. At the south end of the deposit a short incline was sunk from pit 4 in a northwest direction and a level and stope were developed to mine the underground continuation of the tactite in pit 3.

Underground development of the Riley Extension mine by Getchell Mine, Inc., was started in 1951 on the northern continuation of the Riley mine tactite body (pl. 7). The underground workings are connected with the Riley mine on the 200 level and by stopes from the 300 level and 400 sublevel. A 2¾-compartment vertical shaft was sunk to 183 feet, with a station 152 feet below the collar from which cross cuts were driven to the ore body and raises put up to the 200 level.[24] Intermediate levels were driven off these raises, from which shrinkage stopes were developed. In 1955 an inclined shaft was being sunk down dip from the 400 level to develop more ore at depth.

The tactite deposits of the Riley and Riley Extension mines are on the east contact of the northern lobe of the granodiorite stock. Limestone and interbedded shale of the Preble formation have been intruded by the granodiorite. The Riley mine has one main tactite zone. Ore has been mined from three tactites at the Riley Extension mine: the main tactite, called the east vein, is an extension of that in the Riley mine; a second important underground ore body is west of and stratigraphically beneath the first; and a third was mined at the surface still farther west at the granodiorite contact.

The granodiorite contact trends generally north, but it has abrupt changes in direction that, from south to north, carry it progressively farther west (pl. 6). North of pit 4 the contact strikes practically west approximately 400 feet to pit 3. From pit 3 of the Riley mine north to the little valley west of the vertical shaft at the Riley Extension mine the general strike of the granodiorite contact is a few degrees west of north, and the dip is from 30° to 55° east, predominantly near 40°. Along this course the contact has several small sharp

[24] Newman, 1955, p. 22–23.

bends. West of the vertical shaft the granodiorite ends in a blunt northward projection and the contact swings westward approximately 900 feet and partly encloses a broad reentrant of sedimentary rocks between the projection and the main part of the granitic body. The sedimentary rocks strike generally south, dip moderately east and are for the most part parallel to the granodiorite contact, except locally where the granodiorite transects the structure.

The sedimentary rocks adjacent to the granodiorite at the Riley mine are practically all limestone or metamorphic rocks derived from limestone; at the Riley Extension mine a wide body of hornfels with some interbedded limestone occurs in the reentrant of sedimentary rocks where the granodiorite contact swings west.

Some of the limestone is rather pure, thick bedded to massive, medium to coarse grained; much is impure, fine grained, and thin bedded. Metamorphism of the limestone near the granodiorite contact has formed marble, calc-silicate rock, and small bodies of wollastonite rock as well as the tactite that contains scheelite. The hornfels is dark and fine grained and has a more or less well defined platy parting. Close to the granodiorite contact it grades into a purplish or brownish fine-grained, commonly somewhat schistose biotite-quartz-plagioclase rock. At many places in both mines a stratum of this schistose rock that ranges from a few inches to as much as 10 feet in thickness is the footwall between the tactite ore body and granodiorite. Schistose hornfels forms the hanging wall as well as the footwall of the "west vein" in the Riley Extension mine. The rock east of the limestone belt is dark gray, soft, and crumbly brecciated hornfels or argillite that marks a fault zone along the front of the range.

Dikes of hard fine-grained dacite porphyry cut all the rocks at the surface and underground.

The "east vein," a tactite zone mined in both the Riley and the Riley Extension mines, extends along the granodiorite contact for approximately 1,600 feet at the surface and on the 200 foot level (pls. 6, 7). Mostly it is between marble or calc-silicate rock and the underlying granodiorite or schistose biotite hornfels. The tactite includes reddish brown medium- to coarse-grained garnetite that locally contains some green clinopyroxene of the diopside–hedenbergite series, banded rocks composed of alternating reddish brown garnet and green clinopyroxene layers, and, less commonly, rocks composed almost wholly of green clinopyroxene with minor amounts of garnet. Locally, especially at the footwall contact but also extending well into the tactite zone, tactite is partly replaced and veined by irregular masses of gray glassy quartz. Granodiorite also contains an abundance of this late quartz where it is the footwall of quartz-rich tactite.

Scheelite is almost completely restricted to the tactite and occurs as small discrete grains that are generally more plentiful in the garnetite and garnet-rich layers of garnet clinopyroxene rock than in the pyroxene layers or the pyroxene tactite. Scheelite is more abundant, coarser, and commonly has fairly well defined crystalline forms in the quartz-rich tactite. The tactite contains a few small grains of pyrite and chalcopyrite at many places; concentrations of these sulfides in small masses, measurable in inches, locally constitute as much as 30 percent of the tactite. Generally the sulfide concentrations are in places where glassy quartz is abundant. The tactite also contains small concentrations of molybdenite in places. Sphalerite, commonly accompanied by a little galena, is fairly common but not as plentiful as pyrite and chalcopyrite.

The contacts of the tactite in the Riley mine and its continuation in the Riley Extension mine are sharp and definite. Even the hanging-wall contact is abrupt, and in places transects the bedding of the limestone or its metamorphosed equivalent. Generally, a zone of calc-silicate rock overlies the tactite zone and in contrast therewith has a highly variable thickness and indefinite outer limit. In some places pure marble or marble containing wollastonite adjoins the tactite zone and may lie between tactite and the zone of calc-silicate rocks.

The most continuous body of tactite in this zone was in the northern part of the Riley mine and the Riley Extension mine where a roughly tabular mass has been developed and mined for approximately 800 feet down dip (pls. 6, 7). The tactite, which pinches and swells along strike and down dip, ranges from 2 or 3 feet to as much as 20 feet in thickness and is 10 feet thick on the average.

At its north end the tactite ends rather abruptly by thinning and fingering out along the strike. Presumably the termination coincides with the ending of the apophysis of granodiorite that underlies the north end of the tactite zone, for limestone favorable for replacement is present where the tactite ends. At its south end the tactite thins down and ends where the granodiorite footwall swells eastward. The swell, which is between pits *1B* and *1C* at the surface (pl. 6), pitches northeastward in depth, thereby causing the tactite to terminate farther north at successively deeper levels in the Riley mine. On the 500 level of the Riley mine (pl. 7) the tactite is thin, and drill-hole data east of the tactite on the 400 level of the Riley Extension mine indicate that the body may be too narrow to mine economically less than 100 feet below the level (pl. 6, section *A-A'*).

The tactite bodies are less continuous in the southern part of the Riley mine, where they are localized by irregularities of the granodiorite contact. Sharp bends in the contact form troughlike structures that plunge slightly north of east at an angle of about 40°. The thick parts of the tactite follow the troughs, while the thin or barren sections are on the ridges between the troughs where the contact bends outward (pl. 6).

The granodiorite contact and the tactite are generally west of the Getchell fault zone, however, at the south end of the property in pit 4, granodiorite and the ore zone end against the fault. Northward the contact is progressively farther west of the fault. Underground there is no evidence that the ore north of pit 4 is cut off by the fault; mostly the fault and granodiorite contact have nearly the same eastward dip (pl. 6, sections *C-C'* and *D-D'*).

At the Riley Extension mine another tactite zone, called the "west vein" by the operators, was discovered underground approximately 125 feet west of and 100 feet stratigraphically below the tactite that is the continuation of the Riley mine tactite (section *A-A'*, pl. 6). This lower tactite does not crop out at the surface but possibly the limestone lens that is partly exposed on the hillside approximately 900 feet N. 70° W. of the shaft (pl. 6) is part of the same layer, which at depth is changed to tactite. The footwall is schistose biotite hornfels and the hanging wall is calc-silicate rock and impure thin-bedded limestone.

The tactite of the "west vein" is different from the tactite of the "east vein" and most other tactites in the area, for it is a greenish-gray massive fine-grained rock composed of matted pale-green actinolitic amphibole and calcite; a little typical garnet-pyroxene tactite is found in places but is not common. Grains of scheelite are scattered through the tactite, which also contains some small grains of pyrrhotite and chalcopyrite. The boundaries of the tactite zone are less definite than those of the garnet-pyroxene tactites elsewhere, and scheelite also is not as strictly confined to the tactite. During the underground mapping of the "west vein", it was difficult in many places to determine the limits of the ore without an ultraviolet lamp.

The granodiorite apophysis that is below the "east vein" tactite in the Riley Extension mine overlies the hanging-wall rocks of the "west vein" (pl. 6, section *A-A'*); the main granodiorite body is west, south, and beneath the mineralized "west vein" zone. The granodiorite contact surface has, therefore, the shape of a broad northward-plunging trough. Presumably the ore was localized in the only available limestone above the trough and extended upward along bedding from the granodiorite and ended below the present surface.

The small deposit mined in the open cut at the Riley Extension mine is in a limestone lens in hornfels near the granodiorite contact (pl. 6). The scheelite is in garnetiferous tactite, and some is in pale-green amphibole tactite. The ore, which has been mostly mined out, appears to have been largely confined to a lens of massive limestone in a body of thin-bedded impure limestone.

TONOPAH MINE

The Tonopah mine, owned by Getchell Mine, Inc., is in the SW¼ sec. 33, T. 39 N., R. 42 E., on the northeast contact of the northern lobe of the granodiorite stock (fig. 16). The open pits are on the lower slopes of the range, approximately 600 to 1,200 feet west of the large South Extension pit of the Getchell gold mine. The portal of the underground workings is in the west wall of the gold pit.

Mining at the Tonopah property did not begin until 1950. Before then, only the most northern pit had been opened in 1943 as a tungsten prospect (Hobbs and Clabaugh, 1946, p. 19). Hess and Larsen (1921, p. 304) mention the Jack mine, which may be the same as the Tonopah, where a body of tactite along the northeastern part of the contact was prospected for copper and molybdenum. Getchell Mine, Inc., started open-pit mining at the Tonopah in 1950 and produced 12,650 tons of tactite. The surface operations were idle when the property was examined in 1955, but underground exploration and development were being carried on (pl. 8). The underground operations began in 1953 when a 500-foot adit, which was originally driven to explore the footwall of the gold-bearing zone, was extended to intersect the granodiorite contact. A small tactite body was encountered and mined from overhead stopes. A drift northward on the contact of a dikelike offshoot for approximately 400 feet failed to find minable tactite, so a crosscut was driven westward. The crosscut encountered a tactite body, which was being developed in 1955, that is the down-dip extension of an ore body that had been mined on the surface. By July 1, 1955, production from the underground workings had been 7,100 tons of tactite. The total production from the surface and underground was 19,750 tons.

At the Tonopah mine, granodiorite is intrusive into limestone and shale of the Preble formation. Much of the limestone is impure and thin bedded and has interbeds of hornfelsed shale. The beds strike generally north to northeast and dip east, but locally they are distorted by small-scale complex folds. The granodiorite contact strikes approximately N. 50° W., dips moderately to steeply northeast, and cuts across the bedding of the sedimentary rocks (pl. 8). In the southeastern part of the map area the contact has a more northerly strike and is mostly parallel to the bedding. Furthermore, in this southeastern area the granodiorite approaches the Getchell fault zone and is displaced by a minor fault related to that zone of faulting; a bit farther south, beyond the area of pl. 8, granodiorite is on the footwall of the main Getchell fault zone.

The tactite is fine- to medium-grained pyroxene-garnet rock that commonly has a crudely defined color banding caused by variations in amounts of pyroxene and garnet in parallel laminae. Scheelite occurs in the tactite as small grains, few of which are more than 2 to 3 mm in diameter, that tend to be concentrated in narrow zones parallel to the rock layering. At a few places tactite next to the granodiorite is impregnated with glassy gray quartz, and in some of these places the granodiorite is also silicified. The sulfide minerals, pyrite, chalcopyrite, and molybdenite, are much less common in the tactite than at some of the other deposits. Some green incrustations of a secondary copper mineral, possibly chrysocolla, coat fracture surfaces in tactite at some places in the pits.

Limestone adjacent to the tactite has been converted to calc-silicate rock and marble. The shale has been metamorphosed to fine-grained dark hornfels.

Tactite is exposed in mine pits at the surface for approximately 1,000 feet along the granodiorite contact. In the exposures the tactite layer ranges from 1 foot to as much as 15 feet in thickness, is parallel to the granodiorite contact, and is transverse to the bedding of the limestone. Mapping shows that tactite is discontinuous along the contact, and the tactite bodies appear to be separated by barren rock; but poor exposures preclude accurate delineation of the relations.

Exploration at depth is not sufficient to clearly determine the form of the tactite, but from the available data the bodies seem to be tabular and continuous down dip, at least to the 500-level underground workings. The tactite bodies are not guided by any obvious strong structural controls; unlike most of the other deposits in the district, they are not localized at any pronounced bends in the granodiorite contact. A change in direction of the contact may, however, be responsible for the localization of the tactite body that was mined at at the south end of the 500 foot level and that is exposed in the southernmost of the main open cuts (pl. 8). The wide tactite body exposed in the west end of the 500 foot level may be related to a change in dip of the granodiorite contact, for underground the contact dips steeply west, whereas at the surface it dips moderately northeast.

ALPINE (PORVENIR) MINE

The Alpine mine is near the crest of the Osgood Mountains at an altitude of about 7,600 feet in SW¼ sec. 6, T. 38 N., R. 42 E. (fig. 16). It is about 4 miles by road from the Getchell mine, and can be reached by a road that branches south from the Burma Road at the crest of the range.

In 1943 the property was operated as the Porvenir mine by W. C. Rigg of Winnemucca, Nevada, and about 8,000 tons of tactite averaging 0.5 percent WO_3 was mined from an open pit at the south end of the deposit (Hobbs and Clabaugh, 1946, p. 25). The property was idle from 1945 until 1950, when Getchell Mine, Inc., resumed open-pit operations and produced 43,700 tons of ore. The property was inactive in 1954, but some tactite still remained in the pits.

The mine (pl. 9) is on the west contact of the northern lobe of the granodiorite stock with rocks of the Preble formation. The granodiorite contact strikes approximately N. 30° W. and dips steeply east. Inasmuch as the contact—at least at the outcrop—dips toward the stock, the sedimentary rocks chiefly underlie the intrusive body.

The host rock is a limestone unit in a predominantly hornfels area. The limestone, which is thin bedded and rather impure, has been converted to tactite or calc-silicate rock near the granodiorite contact. Contact metamorphism has affected the limestone as much as 700 or 800 feet from the granodiorite contact and has resulted in bleaching and recrystallization of the purer carbonate layers; the impure layers are calc-silicate rock. The hornfels formed from shaly rocks is dark and commonly spotted with porphyroblastic cordierite. Near the contact it is a purplish fine-grained biotite-rich hornfels. Much of the tactite that remains in the pits is composed of coarsely crystalline brown garnet with prominent prisms of epidote and fine-grained green actinolitic hornblende; some is mostly garnet with small amounts of calcite, and some is predominantly epidote and calcite, with subordinate amounts of actinolitic amphibole. Garnet-pyroxene tactite seems to be scarce but may have been more abundant in the material that has been mined. The tungsten content is reported to have ranged from 0.4 to 0.5 percent WO_3, and some parts contained more than 1 percent WO_3 (Hobbs and Clabaugh, 1946, p. 26). The tenor of the remaining tactite is not known.

Tactite occurred intermittently for approximately 1,300 feet along the contact, but the main production came from two pits, at the north and south ends of the property (pl. 9). The bodies were tabular or flattened lenses that lay parallel to the granodiorite contact. The tactite that remains in the open cuts ranges from approximately 3 to 20 feet in thickness. An erroneous impression of the amount of tactite remaining is obtained from the map, because the tabular tactite bodies are nearly parallel to the slopes of the hillside and the walls of the cuts. The depth to which the remaining tactite continues is uncertain. With favorable structural situations it could presumably continue in depth as far as the granodiorite is in contact with limestone.

The tactite body that was mined in the South Pit is situated where the surface trace of the granodiorite contact makes a right-angle turn. In effect, the granodiorite surface forms an inverted trough that pitches steeply northeastward, and the tactite is in the trough, on the footwall side of the contact. A small tactite body has been exposed in a cut 100 feet or so north of the north rim of the South Pit, where the contact also makes another right-angle turn. For approximately 600 feet, to the North Pit, the granodiorite contact is straight and barren of tactite. The North Pit tactite body cannot be correlated with any very obvious structural feature, for the trace of the contact is essentially straight. A structural feature that may be at least partly responsible for localization of tactite in the North Pit is visible, however, in the south wall of the pit near the bottom, but is too small to be shown on the map. The tactite layer in the south wall is above a short, blunt, nearly flat-lying tongue of granodiorite that extends 10 feet or so into the footwall and, if continued along the strike, would form a narrow shelf or trough.

RICHMOND MINE

The Richmond mine in SW¼ sec. 31, T. 39 N., R. 42 E., is on the west side of the Osgood Mountains (fig. 16), in the upper part of Anderson Canyon at an altitude of approximately 6,500 feet. Two tactite deposits, on the east and west sides of the canyon, were mined from open cuts (pl. 10), and the deposit on the east side was also mined partly by underground workings. The property is accessible from the Getchell mine by Burma Road, a distance of approximately 5 miles.

Hess and Larsen (1921, p. 302–304) briefly described the property, which had been operated as a silver prospect up to the time of their examination. Sometime prior to the summer of 1942, U.S. Vanadium Corp. purchased the property and it was leased to W. C. Rigg of Winnemucca, Nevada (Hobbs and Clabaugh, 1946, p. 19). Between the summers of 1942 and 1943 more than 31,500 tons of tactite was mined from open-cut and underground operations on the east tactite deposit. The average tungsten content of the tactite was approximately 0.5 percent WO_3 (Hobbs and Clabaugh, 1946, p. 20). The only reported activity on the west tactite deposit was open-pit mining, which was attempted from June to August 1954, but the operation ceased in August because the low tungsten content of the tactite made further mining unprofitable. Approximately 2,800 tons of ore with an average grade of 0.2 percent WO_3 were mined during that period.

The tactite deposits of the Richmond mine are on the north side of a blunt westward projection of the northern lobe of the granodiorite stock, at the contact of granodiorite with a wide bed of limestone in the midst of hornfels mapped as Preble formation (pl. 1). The contact strikes approximately N. 75° E. and is for the most part vertical. On the west side of the canyon in the southwest corner of the map area, the contact makes nearly a right-angle turn where a short protrusion of granodiorite extends northward (pl. 10). The limestone bed, bounded on the east and west by hornfelsed shaly rocks, strikes into the granodiorite. Tactite has been formed at the contact of granodiorite with the limestone and occurs on both sides of the valley; the larger tactite body is on the west side in a reentrant in the granodiorite. Bedrock is not exposed in the valley between the two tactite bodies, but the west body does not appear to continue east of the lower pit, and the east ore body is terminated on the west by hornfels. A small exposure of hornfels on the west side of the canyon suggests that limestone—and therefore, also tactite—does not underlie the covered area.

The tactite bodies are tabular, and they strike and dip parallel to the granodiorite contact and are mostly perpendicular to the bedding of the limestone that they replace. Their outer limit is as clearly defined as at the granodiorite contact and is unrelated to the bedding. On the limestone side they are joined by a zone of calc-silicate rock and marble a few feet to a few tens of feet wide that grades into limestone. In this zone, bedding control is more evident, and layers of marble alternate with layers of calc-silicate rock.

Characteristically, the tactite is a medium- to coarse-grained granular brown to greenish-brown garnetite with not much more than 10 to 15 percent of accessory minerals, including interstitial calcite, quartz, a little chlorite, and fibrous pale green amphibole. Fairly large grains of pyrite are locally plentiful in the east tactite body. Most conspicuous and characteristic is the absence of pyroxene, which is present in most other tactite bodies in the district. Locally the tactite contains considerable quartz, accompanied by epidote.

At places on the granodiorite contact of the east tactite body small masses of quartz that contain pyrite, galena, and sphalerite as well as a few grains of scheelite replace the granodiorite and tactite. No sulfides were seen in the tactite as suggested by Hess and Larsen's

(1921, p. 30) statement that the "* * * silver ore is in considerable part on the limestone side of the contact zone * * *".

The east tactite body was approximately 210 feet long and averaged nearly 30 feet in thickness (pl. 10; Hobbs and Clabough, 1946, p. 20, fig. 28). Although it is the smaller of the two, it has been the most extensively mined. It was mined from an open cut and by underground methods in which stopes were developed to the surface. Some tactite remains in pillars and as thin remnants lying on the granodiorite contact surface. The underground workings were inaccessible in 1954, but Hobbs and Clabaugh (1946, p. 19–20) reported that they consisted of an adit 120 feet below the floor of the cut, and a sublevel below the open cut and 85 feet above the adit level. According to their report (1946, p. 20)

> * * * most of the ore body is cut out on the main adit level by fine-grained dike rock which takes the place of limestone at the contact, and little ore is indicated below the sublevel. The average tungsten content of the tactite was about 0.5 percent WO_3.

The west tactite deposit is about 300 feet long and from 30 to 140 feet thick. Although it is larger than the last deposit, its average tungsten content is so low that attempts to mine it profitably have been unsuccessful. The tactite is mostly an almost pure garnetite with sparsely distributed small grains of scheelite. Hobbs and Clabaugh (1946, p. 20) reported that in adits the greatest concentration of scheelite appeared to be at the contact with granodiorite; surface samples ranged from 0 to 0.58 percent WO_3, but the distribution was very erratic, and more than half the samples contained less than 0.15 percent WO_3 (Hobbs and Clabaugh, 1946, plate 9). An attempt was made to mine the west body in open cuts during the summer of 1954, but the operation was abandoned after about 2,800 tons of tactite averaging 0.2 percent WO_3 were produced.

MOUNTAIN KING MINE

The Mountain King mine is east of the Richmond mine, on land leased from the Southern Pacific Land Co. The claims were held by R. L. Taylor, M. V. Taylor, K. Nielsen, and L. Thomas, members of a private company, T.N.T., Inc., which was operating the mine. The workings, which include three open cuts, are in SE¼ sec. 31, T. 39 N., R. 42 E. (fig. 16), on the upper western slope of the Osgood Mountains from an altitude of about 6,800 feet to the crest of the range at about 7,000 feet, and are easily reached from Burma Road, about 3.5 miles from the Getchell mine. The workings at the crest of the range are referred to in the report by Hobbs and Clabaugh (1946, p. 28, plate 1) as Ed Knight's claims.

Mining of these deposits was started in 1951, and a small production was maintained that totalled about 18,000 tons in June 1955. The tungsten content of the ore produced ranged from 0.47 percent to as much as 0.85 percent WO_3, and the weighted average is 0.6 percent WO_3, slightly higher than the average grade of ore produced from other mines in the district. The ore was trucked to the custom mill at the Getchell mine.

In this area the granodiorite contact maintains the northeast strike it has at the Richmond mine, and continues with slight irregularities in its surface trace to the eastern part of the map area, where the strike becomes slightly more northerly (pl. 10). It dips southeast at steep to moderate angles. The argillaceous hornfels and interbedded limestone of the Preble formation, which are intruded by the granodiorite, have a more northeasterly trend than at the Richmond mine and hence are only slightly truncated by the intrusive rock.

The sedimentary rocks adjacent to the granodiorite in this area are predominantly argillaceous hornfels but include some interbedded chert and thin interbeds of limestone and impure shaly limestone. Tactite is restricted to places where the granodiorite is intrusive into shaly limestone beds; the purer limestone along the contact has been converted to marble containing some tremolite. The tactite zone at the easternmost mine pit on the ridge crest is only about 5 feet thick, and other layers only a few inches thick interbedded with hornfels are not shown on the map. These were originally thin limestone members interbedded with argillaceous and cherty rocks. The localization of the tactite bodies, other than proximity of favorable beds to granodiorite, seems to have been controlled by composition and bedding; they follow the bedding of the sedimentary rocks. In the two pits adjacent to Burma Road the tactite bodies are thickest next to the granodiorite and become narrower and pinch out northeastward along the strike of the beds.

The tactites are brown garnetite containing small amounts of quartz. Scheelite grains are scattered through the garnetite and many are intergrown with the quartz. In places the tactite is cut by seams of greenish nontronite, which is an alteration product of the garnet.

In the face of the open cut adjacent to Burma Road, on the east side, scheelite also occurs along seams which are parallel to a prominent close-spaced set of joints in the granodiorite that strike N. 10° E. and dip 28° W. The alteration occurs in the granodiorite and in the hornfels as well. In the granodiorite soft masses of white mica accompanied by chlorite, crystalline quartz.

and tiny euhedral crystals of smoky apatite have been formed, and within it are local concentrations of crystalline scheelite. In places next to the altered zone there are cavities in the granodiorite as much as 6 inches long that are lined with quartz crystals. Where the seams of alteration cut the hornfels the rock is silicified, and it contains small amounts of fine-grained white mica, some minute crystals of smoky apatite, goethite, and limonite perhaps derived from pyrite, and in places incrustations of green secondary copper minerals. No concentrations of scheelite were observed in the alteration seams in hornfels.

T.N.T. MINE

The T.N.T. mine is on the east slope of the Osgood Mountains in the northern part of the range at an altitude of approximately 6,500 feet in NW¼ sec. 32, T. 39 N., R. 42 E. (fig. 16), 0.9 mile due west of the Getchell mine. It is 1.2 miles by road from the Getchell mine, partly over Burma Road, and partly by a very steep mine road.

The mine is owned by T.N.T., Inc., and was started in July 1954, operated for only short time as an open-pit operation, and discontinued in December the same year. The recorded production from August 4 to September 29 was approximately 600 tons of ore whose weighted average grade was 0.87 percent WO_3. The ore was trucked to the custom mill at the Getchell mine.

The workings consist of a small open cut and an exploration adit 345 feet long beneath the ore zone (fig. 18). The exploratory adit and two diamond-drill holes were partly financed by the Defense Minerals Exploration Administration, a Federal agency.

The mine is on the northwest contact of the northern lobe of the granodiorite stock where it begins to narrow rapidly. This part of the contact has a highly irregular surficial trace (pl. 1), and the deposit is in a prong of sedimentary rocks with granodiorite on three sides. At the deposit, the contact strikes northwest and dips steeply northeast. Rocks of the Preble formation, which are mostly argillaceous hornfels but include a few thin discontinuous beds and lenses of limestone, are intruded by the granodiorite. In general the sedimentary rocks strike northeast and because of close folding they dip steeply west and east; at the mine, possibly because of disturbance by emplacement of the granodiorite, bedding attitudes diverge widely from the general pattern.

In contrast to deposits elsewhere in the district, scheelite at the T.N.T. mine is not in a well-defined garnetite tactite zone. A small lens of limestone about 100 feet long and 70 feet thick interbedded with argillaceous hornfels is partly metamorphosed to a clinopyroxene-plagioclase hornfels containing small masses and veinlets of sulfide minerals, mostly pyrrhotite and chalcopyrite, and lesser amounts of dark-brown sphalerite and a few grains of galena. Part of the calc-silicate rocks contains small grains of scheelite in addition to the sulfide minerals; the scheelite grains are scattered through the rock and not necessarily associated with the sulfides. Scheelite is restricted to the calc-silicate rock but its distribution is not well defined; however, it does not seem to occur in the adjacent carbonate rocks that have been partly bleached and recrystallized.

The small size of the limestone body and seemingly erratic distribution of scheelite preclude important production from the T.N.T. mine in the future.

SECTION 5 (MARSHALL CANYON) PIT

Exploratory pits were opened by Getchell Mine, Inc., in the summer of 1955 on a scheelite prospect near the head of Hansen Canyon, known locally also as Marshall Canyon, in SW¼ sec. 5, T. 38 N., R. 42 E. (fig. 16). The operation was suspended after a few hundred tons of scheelite-bearing rock were produced. Previously the area had been prospected for silver.

The prospect is near the center of the northern lobe of the granodiorite stock (pl. 1), in the southern part of an area of sericitic alteration p. 46, 88). A few irregular-shaped bodies and 3- to 5-foot veins of coarse glassy quartz cut the granodiorite in the southern part of the altered area. Some of the quartz is coarsely crystalline prisms. In places the veins contain cubes of pyrite as much as 4 to 6 inches on a side. In addition it contains erratically and generally sparsely distributed scheelite in subhedral to euhedral grains as much as 5 mm across, a few larger crystals a few centimeters in dimension, and phantomlike films that coat intracrystalline surfaces in the quartz. In places a boxwork of iron-stained jasperoid and small quartz crystals contains a few crystals of scheelite as much as 1 cm across. Scheelite apparently does not occur in the altered granodiorite but is confined to the quartz bodies.

PROSPECTS IN THE CANYON OF OSGOOD CREEK

On the south side of the canyon in NW¼ sec. 20, T. 38 N., R. 42 E., the contact between granodiorite and limestone has yielded a little tactite float that contains a few grains of scheelite, and in addition, irregularities in the contact deserve some attention. Small-scale exploration has also revealed a small amount of tactite that is low in scheelite on the north side of Osgood Creek where the granodiorite contact has a small irregularity, near the line between secs. 17 and 18, in SW¼ sec. 17, T. 38 N., R. 42 (pl. 1). This and the former locality were also mentioned by Hobbs and Clabaugh (1946, p.

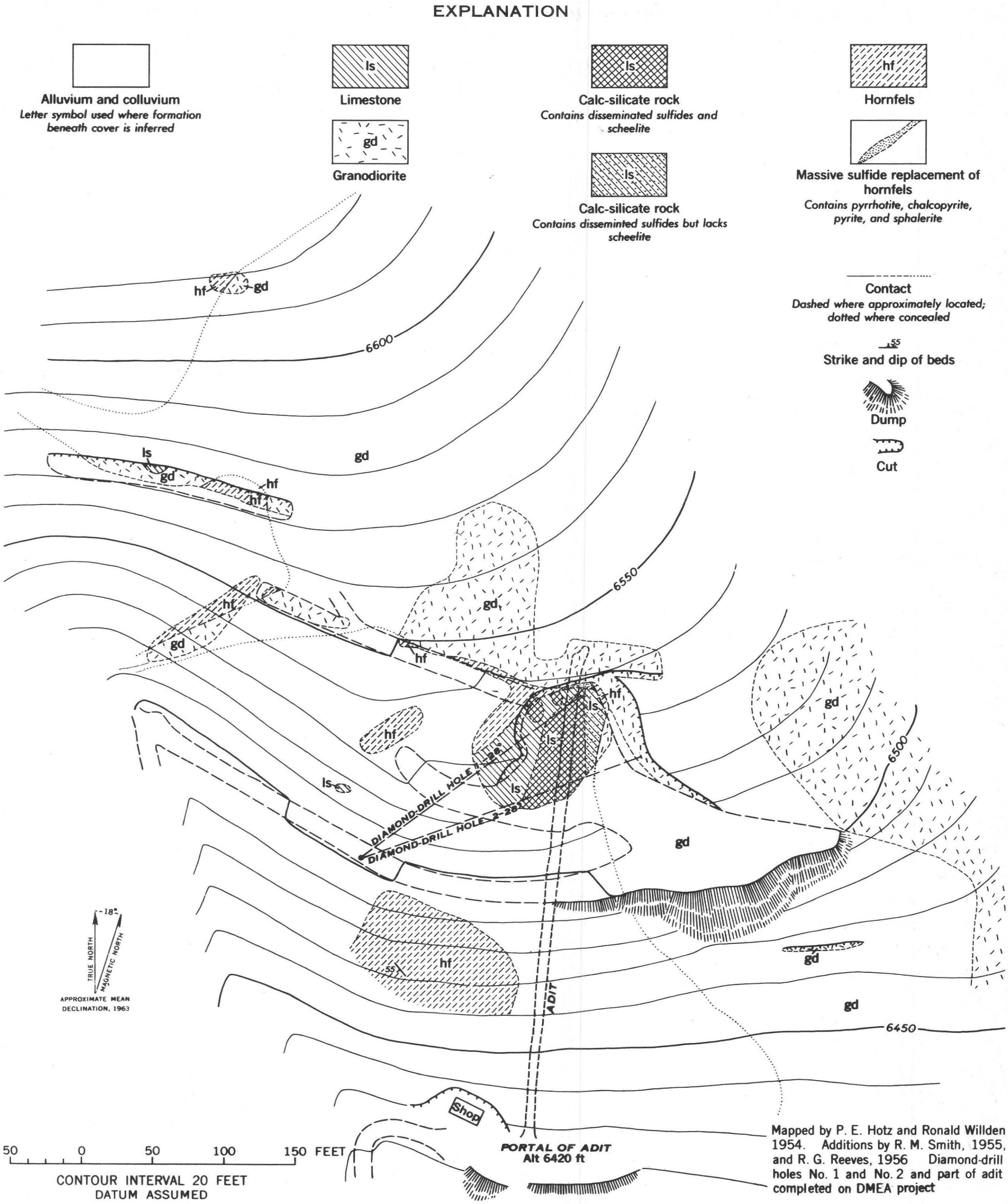


FIGURE 18.—Geologic map of the T.N.T. mine.

27). Tactite float is also known near the west contact of the southern granodiorite body east of Adam Peak, in eastern part NE¼ sec. 24 T. 38 N., R. 41 E. (pl. 1). This area is accessible from Granite Creek by a bulldozed road from the Marcus mine.

GOLD DEPOSITS

GETCHELL MINE

The Getchell mine is at the foot of the Osgood Mountains on the northeast side of the range (pl. 1). Mining operations have been carried on along a zone more than 7,000 feet long, most of which is in the western part of sec. 33, T. 39 N., R. 42 E.; parts of the zone are also in SW¼ sec. 28, SE¼ sec. 29, and NE¼ sec. 32, same township and range; at its south end the zone is in the northern part of sec. 4, T. 38 N., R. 42 E. The mine is 26 miles north of Golconda via good blacktop and graded dirt roads, and 15 miles northwest of Redhouse, the nearest railroad station.

HISTORY AND PRODUCTION

The Getchell gold deposit was discovered in the fall of 1934 by two prospectors, Ed Knight and Emmet Chase, on the site of a prominent siliceous outcrop that had been known for many years but that had failed to yield any gold to those who previously had tested it by panning. Knight and Chase interested Noble H. Getchell in the prospect, and it was sampled and assayed. The assays showed that the samples contained between 0.1 and 0.2 ounces of gold per ton and an equivalent amount of silver. Later that year Getchell purchased Knight's interest and called the prospect to the attention of Mr. George Wingfield. Wingfield bought Chase's share of the property early in 1935, and Getchell and Wingfield organized Getchell Mine, Inc., with Wingfield as president, Getchell as vice-president and general manager, and T. L. Wilcox as secretary-treasurer. The outcrops were prospected by several short adits that intersected from 60 feet to 90 feet of ore, and churn-drill holes showed that the ore extended to at least 1,100 feet down the dip of the vein, and that the grade warranted further work. Therefore, development was continued with drill holes, adits, shafts, and drifts (Joralemon, 1949,[25] p. 8).

The following historical and production data were taken from the U.S. Bureau of Mines Minerals Yearbooks, 1938 to 1951, inclusive. Production started in February or March 1938 with ore mined from an open pit, and by the year's end the Getchell mine had established itself as the leading mining enterprise in Humboldt County. The first dividend was declared in September of the same year. By 1939 Getchell mine was the leading gold producer in Nevada, a position that was maintained until 1943 when byproduct gold from the copper pits at Ely exeeded Getchell's output. The bulk of the production during the first five years was from oxidized ores, but as mining progressed more and more arsenic sulfide ores had to be treated. By September of 1943 the oxide ores were virtually exhausted and the ore had to be roasted before the gold was recovered by cyanidation. In 1941, however, a Cottrell electric precipitating unit had been installed to save the arsenic that was liberated by roasting the sulfide ore, and in 1943–45, when government wartime restrictions forced the shutdown of many gold producers, Getchell mine was permitted to continue operation as a producer of "strategic" arsenic. In 1943 arsenious oxide was being produced at the rate of 10 to 25 tons per day from furnace fume. In 1945, however, gold production became increasingly difficult because of rising costs of labor and supply, and shortages of labor and material. On about May 1, 1945, mining of the gold ore was suspended and activities were confined to underground development, reconstruction and expansion of the plant, and metallurgical research. Discovery of additional sources of oxidized ore at the Getchell mine and at the Ogee-Pinson lease 7½ miles south of Getchell enabled gold production to be resumed in 1948 and continued through 1950. In 1949 the Getchell plant was again an important gold producer and rose to second place in the list of Nevada gold operations. A description of the milling operation is given by Huttl (1950). In 1951, however, the Getchell mine gave up its gold operations entirely and converted its plant to treat tungsten ores. Following the close-down of tungsten production in 1957, plans for resuming the gold operations were being formulated; this depended on the solving of metallurgical problems involving separation of gold from the sulfide ores.

The gold ore has been mined by power shovels from three large open pits developed along the fault zone. When operations ceased, open-pit mining had progressed to approximately 100 feet maximum depth below the original surface, and the oxidized ore was virtually exhausted. Considerable underground exploration and development have been carried out that include a vertical shaft (the S24 shaft), from which development work was conducted on the 800 and 1,000 levels beneath the southern part of the South Extension pit; a 45° inclined shaft from the 400 level in No. 4 tunnel, whose portal is in the west side of the South pit, and from which workings were developed on the 600 and 800 levels; and the North shaft, which is connected with the

[25] Joralemon, Peter, 1949, The occurrence of gold at the Getchell mine, Nevada: Harvard Univ., unpublished doctoral thesis.

600 level at the north end of the mine by a haulage drift 700 feet long. Considerable underground exploratory and development work has been done on the 400 and 600 levels at the north end of the mine preparatory to mining the north ore body. The North shaft and inclined shaft from No. 4 tunnel are connected by a drift on the 600 level approximately 2,800 feet long.

Production data for the Getchell gold mine are incomplete, but the available figures are presented in table 21. The information has been obtained from yearly volumes of the Minerals Yearbook, except where noted in the table. According to Joralemon.[26]

> * * * at the close of 1945 the mine had produced 428,571 ounces gold with a gross value of $15 million. Most of this production had come from three large open pits.

TABLE 21.—*Gold, silver, copper, and lead production from the Getchell mine*

Year	Tons	Gold (ounces)	Silver (ounces)	Copper (pounds)	Lead (pounds)	Total value
1938	159,857	23,574	2,964			$827,006
1939	278,975	49,288	2,559	----------	----------	1,726,902
1940–41	(1)	----------	----------	----------	----------	----------
1942	295,624	46,625	8,442	2,600	6,200	1,638,608
1943	257,572	35,047	10,603	64,000	5,700	1,242,933
1944	(1)	----------	----------	----------	----------	----------
1945	----------	10,752	2,264	----------	4,000	378,274
1946–47	None	----------	----------	----------	----------	----------
1948	(1)	----------	----------	----------	----------	----------
1949	151,695	18,785	2,000	----------	----------	[2] 656,038
1950	153,516	----------	----------	----------	----------	[3] 689,992

1 Production figures not available.
2 Eng. and Mining Jour., 1950, v. 151, no. 6, p. 122.
3 Gross yield Jan. 1 to June 30 (Mining Record, v. 61, no. 39, p. 1, Sept. 1950).

GEOLOGY

The gold deposits of the Getchell mine are in fractured rocks along the Getchell fault zone (pl. 11). Areal mapping has shown that the predominant country rocks of the mineralized zone are part of the Preble formation of Cambrian age. They include dark hornfels and lenticular bodies of thin-bedded limestone. Dikes of fine-grained andesite porphyry cut the sedimentary rocks. In the southern and central parts of the mine area granodiorite forms the walls of the fault zone, but does not seem to have been one of the host rocks for the ore. At the northern end of the mine area, in the vicinity of the North shaft, a somewhat conical body of rhyolite tuff lies east of the fault zone.

On the west side of the faulted and mineralized zone the average strike of the sedimentary rocks is about parallel to the east contact of the granodiorite stock. Dips vary somewhat, but on the average the bedding is inclined about 50° E. At the north end of the mineralized zone, and beyond, the beds strike more northwest and west and dip northeast and north.

East of the zone of faulting and mineralization the beds are mostly concealed by alluvium, and their structure is unknown. A few poor exposures suggest that the strata strike parallel to the fault zone.

> The Getchell fault zone is the dominant structural feature of the mine area [pl. 11]. It trends in a northerly direction and is composed of a persistent footwall strand dipping moderately eastward, with steeper, arcuate hanging-wall branches. (Joralemon, 1951, p. 270.)

As discussed on page 75 there is evidence that the fault had a strike-slip movement with left-lateral displacement, which was followed by later normal movement. Joralemon (op. cit.) believes that

> the steeper, hanging-wall branches were formed at that time [during the episode of normal movement] and occur above flexures in the footwall fault where the dip flattens * * *.

The gold ore bodies are tabular deposits unlike veins in the usual sense, but because they are sheetlike bodies of mineralized rock whose form and distribution are related to fractures of the Getchell fault zone, they are most conveniently referred to as veins. The following descriptions are derived from Joralemon's unpublished report[27] with some editorial changes by the present authors.

The gold veins are localized almost entirely within the Getchell fault zone, lying against the footwall of the numerous echelon branches and the cymoid loops (McKinstry, 1948, p. 315–319). The ore bodies are lenticular or tabular in shape, and in places they swell to nearly 200 feet in width. Most of the veins are linked by relatively narrow, low-grade stretches along cymoid loops (pl. 11). Between junctures the veins swell irregularly, rolling with the fault both vertically and laterally. The shapes of the veins are shown in the composite vein plan (pl. 11).

At the south end of the mine the fault zone contains at least five veins. The largest vein persists throughout the South Extension pit. Near the surface it is steeply dipping to vertical, but in depth it joins the footwall branches of the fault zone and resumes the normal easterly dip (pl. 11, section *A–A′*). It appears to be the main vein on the surface, although it does not follow the major fault but is a wide steep hanging-wall branch from the main fault to the west (pl. 11). The small vein between the two major ore bodies in the South Extension pit has a westerly dip. The main vein and this minor branch seem to lie along tension fractures formed, perhaps, by normal faulting. Two mineralized branches of the main fault lie west of the pit and are not well exposed. Toward the north these footwall branches join and form the main vein exposed in the South pit. At the south end of the South pit, the strong vein that persists through the South Extension pit swings west and rejoins the main footwall vein.

26 Op. cit., p. 9.

27 Op. cit., p. 50–55.

In the South Extension pit, then, the major footwall fault is occupied by a minor vein from which arcuate hanging-wall veins branch and rejoin.

In the South pit the vein structure is simple. One strong vein persists for 1,000 feet along the main footwall fault. It is bounded on the east by a zone of from 2 to 50 feet of barren blue gouge. One small lenticular vein that lies 700 feet into the footwall from the main vein is not exposed at the surface (pl. 11, section B–B′). It dips steeply east and probably occupies a tension fracture. Drill holes prove that it is a discontinuous lens and does not join laterally with the main vein, either to the north or to the south.

The veins in the north end of the mine area are similar in form and disposition to those in the South Extension pit. At the surface in the North pit the main vein ends, and no northern continuation has been found although the northerly extension of the Getchell fault has been thoroughly explored. Underground workings bear out this observation. Although the main Getchell fault zone continues to the north without deflection it contains no vein. In this area, however, the major fault is no longer parallel to the bedding; rather, the sediments strike at angles from 50° to 80° to that of the fault. Underground workings have shown that the strong vein frays to the north, splitting into a series of minor veins developed along shears parallel to the bedding which with westerly strike departs from the main fault at a moderate angle. Here as in the South Extension pit, a persistent but narrow footwall vein swings toward the west, and from it wider, less persistent, arcuate hanging-wall veins branch and rejoin, thus surrounding a series of narrow wall-rock horses. Both in dip and in strike these veins are persistently lenticular and discontinuous. The northern part of the ore body contains three major hanging-wall branches and several minor veins from 1 to 2 feet thick.

The persistence of the veins with depth is poorly known.[28] Most of the hanging-wall branch veins in the South Extension pit join the main, footwall vein within 100 to 200 feet below the surface. Churn-drill holes show that the two major veins in the area of the South Extension pit persist as separate units to at least 700 feet vertically beneath the surface, but they appear to approach one another with depth because they are 350 feet apart at the surface and less than 100 feet at a depth of 700 feet. At the north end of the mine, the veins also appear to unite with depth.

[28] Op. cit., p. 54.

Vein structure becomes simpler with depth, as several veins merge into one. But, as Joralemon [29] pointed out, there is no obvious reason why the vein structure should not persist to at least 1,000 feet vertically beneath the surface, or 1,500 feet downdip.

Joralemon [30] noted the following facts concerning the distribution of the gold ore at the Getchell mine. The distribution of gold in economic amounts is similar to the distribution of realgar, and therefore realgar is commonly a good indicator of the presence of gold ore at the Getchell mine. Gold is not present throughout the fault zone, but is restricted to the lenticular veins that generally occur along the footwall of the fault zone. The economic boundaries of the ore bodies are sharp, whereas the actual mineralogic boundaries are gradational. The wallrock for at least 100 feet and in places for as much as 400 feet from an ore body contains between 0.01 and 0.08 ounce of gold per ton, but farther out the wallrock contains only traces of gold. At the economic vein wall the gold content increases abruptly to more than 0.1 ounce of gold per ton; however, the increase in gold content at the vein wall is insignificant, as Joralemon [31] pointed out that a change in grade from 0.06 ounce to 0.12 ounce per ton, which is the difference between waste and minable ore, is a change in gold content of less than 0.0002 percent of the rock.

Changes in attitude of the veins apparently have little effect in the gold content, but commonly the wider parts of the vein are the richer parts. Gold occurs in the sedimentary rocks of the fault zone, in altered dike rocks, and in the gumbo. It almost always is more abundant in the gumbo than in the other rocks. Nearly every mineral of the ore is host for at least some of the gold. Pyrite, marcasite, and the carbonaceous matrix of the gumbo are the most important gold bearers, but gold also occurs sparsely in realgar, arsenopyrite, and coarse quartz.

The gold ore bodies at the Getchell mine have been determined from exploration and development to occur for at least 7,000 feet horizontally and 800 feet downdip, and they range in width from a few feet to more than 200 feet with an average width of about 40 feet (Joralemon, 1951, p. 270). At the north end of the mine the ore bodies turn northwest from the main Getchell fault, more or less in conjunction with a bend in the strike of the sedimentary rocks, and end rather abruptly. At the south end the complex of branching veins becomes simpler with the joining of several

[29] Op. cit., p. 55.
[30] Op. cit., p. 86–94.
[31] Op. cit., p. 86–87.

branches, and southward the ore body gradually tapers down to uneconomic proportions, though churn-drill hole exploration suggests that gold ore may lie as much as 500 feet south of the South Extension pit.[32]

From development and exploration data, Joralemon[33] (1951, p. 276, fig. 3) prepared a longitudinal projection showing the distribution of the gold values. Of the vertical distribution he said:[34]

> This projection indicated that the rich gold values become increasingly rare with depth. A section in the vein that extends about 150 feet down the dip from the surface is largely of richer than average ore, and much of it contains more than 0.3 ounce of gold per ton. Deeper than 150 feet, the richer portions become rarer and, instead of covering large parts of the vein in a blanket like form as they do near the surface, they project downward in narrow, steeply pitching shoots. The ore shoots thus have the form of a shallow, extensive blanket with an irregular rootlike lower surface. The roots extend downward at least 300 feet along the dip of the vein from the surface, the great north ore body persists at least 900 feet along the dip of the vein and, at that depth, contains richer gold values than at any other place in the mine * * *.
>
> Much of the vein stuff between these roots is of lower but still mineable grade. This material, containing from 0.1 to 0.3 oz gold per ton, persists to the deepest level of exploration, decreasing only slightly in amount with depth. It may be expected to continue several hundred feet deeper before it pinches into roots similar to those of the richer but shallower ore shoots.
>
> Barren and uneconomic vein material occurs in a large zone that separates the north ore bodies from the ore in the South pit, and again in the footwall veins west of the South Extension pit.

OGEE AND PINSON MINE

The Ogee and Pinson mine is on the south side of Granite Creek in the northern part of sec. 32, T. 38 N., R. 42 E., about 7½ miles south of the Getchell mine (pl. 1). From a small open pit, approximately 4,150 tons of oxide ore assaying 0.186 ounce of gold was mined under a lease agreement in 1949, and treated at the Getchell mill (Huttl, 1950, p. 62). Apparently some prospecting had been done previously on the property.

The property is on a fault that possibly is a southern continuation of the Getchell fault zone, or a branch of the same general system which separates rocks of the Preble formation from those of the Comus formation. The open pit from which the gold ore was mined is in strongly fractured cherty rocks and possibly silicified hornfels, apparently east of the fault. According to Joralemon[35] irregular zones of low-grade ore occurred along bedding planes and minor fractures in argillite on the footwall of the fault, and realgar was reportedly found, though none is visible now.

OTHER OCCURRENCES OF GETCHELL-TYPE GOLD DEPOSITS

According to Joralemon[36] the wide shear zone exposed in the open cuts of the Riley tungsten mine contains "* * * several tens of feet of marginal gold ore." In addition, black, siliceous gouge is exposed in short adits and prospect pits a few hundred feet east of the quadrangle boundary in sec. 4, T. 38 N., R. 42 E., north of Hansen (Marshall) Canyon, between the Riley Extension mine and the South Extension pit of the Getchell mine. Possibly this material also contains gold values, for it resembles material in the Getchell pits and at the Riley mine, but no assay data are available. Joralemon also reported that fault material exposed in a trench several hundred feet east of the Pacific mine contained "* * * nearly 0.1 ounce of gold per ton."

GOLD-QUARTZ VEINS IN THE DUTCH FLAT DISTRICT

Presumably most of the gold mines and prospects in the Dutch Flat district were located and worked after the discovery of placer gold in the district in 1893 (Vanderburg, 1936, p. 94), but their history is unknown. Excepting the Dutch Flat quicksilver mine, all the properties have been abandoned for several years. Their production is not known; the only published production data for the district are the figures given by Vanderburg (1936, p. 94) for placer gold and by Bailey and Phoenix (1944, p. 90) for quicksilver. The ruins of two small mills for recovery of gold from quartz remain. At a few of the mines the scarcity of vein quartz on the dump is out of keeping with the size of the underground workings, and suggests that the vein material was sent to a mill for treatment.

The mines and prospects are situated in the granodiorite stock at Dutch Flat (pl. 1, fig. 15) and in rocks of the Harmony formation as far as half a mile north and south of the center of the stock. About 40 workings were visited. Most of these are no more than prospects with a short adit or pit that was dug on a fractured or altered zone; in many no quartz was visible. Most of the properties warranted only a brief inspection, but a few that had more extensive workings were mapped with tape and compass. Except for one mine, the El Paso, we were not able to learn the names of the properties, as they are numbered on the map (fig. 15) and so listed in the description, table 22.

32 Op. cit., p. 90.
33 Op. cit., p. 93–95, pl. 30.
34 Op. cit., p. 94–95.
35 Op. cit., p. 92.
36 Op. cit., p. 91.

TABLE 22.—*Gold mines and prospects in Dutch Flat*

Number on map (fig. 15)	Location (T. 38 N., R. 40 E.)	Geology	Workings
1 (El Paso mine)	South center, NW¼NW¼ sec. 16.	Discontinuous broken quartz stringers in 4-ft fault zone striking N. 20° E. dip steeply west in altered sandstone and shale of Harmony formation. Stibnite in vein quartz. Sample from stope assays 0.05 oz gold, 1.5 oz silver per ton.	Drift along fault zone, stoped to surface.
2	NE¼NE¼ sec. 17	North-striking, gentle westward-dipping quartz vein pinches downdip from 2 to 3 ft on upper level to 0 on lower level. Wallrock is altered granodiorite. Sheared stibnite on dump, none seen underground.	Vein developed on two levels.
3	NE corner sec. 17	Granodiorite near portal, rest of workings in slightly altered to intensely altered hornfels of Harmony formation. Workings follow several faults containing coarse, vuggy quartz veins and stringers. Quartz on dump contains some galena, pyrite, and chalcopyrite.	About 345 ft of drifts and crosscuts.
4	Near SW¼SW¼ sec. 9	Workings cross granodiorite contact with hornfels and sandstone of Harmony formation. Some alteration. Very little quartz; most in prominent N. 50° E., 60° NW. fault zone in main adit. Traces of scheelite in quartz.	Three adits totaling about 340 ft. No stoping.
5	W part NW¼ sec 16	Country rocks are altered granodiorite faulted against Harmony formation underground. A little quartz in veins near portal of long (south) adit and near its end contain traces of scheelite and chalcopyrite.	Two adits and a 75-ft winze.
6	SE¼NE¼ sec. 17	Sandstone and shale of Harmony formation, altered in few places, cut by nearly continuous northeast-striking fault zone containing discontinuous quartz veins and brecciated quartz. Sample assayed 0.02 oz gold, 0.4 oz silver.	Lower adit (drift) 280 ft long, stoped above and below, with raise to short (65 ft) drift. Estimate more than 1,000 tons removed from stopes, but dumps are small.
7	South center, NW¼ sec. 16.	Shale and sandstone of Harmony formation, more or less altered. Northwest- to northeast-trending faults contain a little brecciated quartz. Quartz on dump contains traces of sphalerite. Sample assayed trace of gold, 0.3 oz silver.	Adit 450 ft long follows several fractures that contain quartz.
8	South center, SW¼ sec. 9	Discontinuous quartz veins 2 in to 2 ft wide in N. 5° to 10° E., 55° NW, fault zone. Quartz is mostly in part of fault south of shaft, not much north of shaft. Quartz from dump contains small amounts or galena, sphalerite, and stibnite.	150-ft-long inclined shaft to depth of 120 ft; five levels with total length of 300 ft. No important stoping.
9	Near center, SW¼ sec. 9	Fault zone, N. 50° E., 80° SE., containing quartz stringer in unaltered sedimentary rocks of Harmony formation.	Adit 307 ft long. Drift 122 ft long perpendicular to adit follows fault. No stoping.
10	NW¼NW¼ sec. 16	Vuggy, iron oxide-stained quartz vein 3 ft wide, strikes N. 10° E., dips 40° W. in altered granodiorite. Small amounts stibnite and pyrite in quartz.	Incline 25 ft long.
11	Near line between NW¼ sec. 16, and NE¼ sec. 17.	Altered granodiorite. No clear structure or vein.	Adit approximately 240 ft long; vertical shaft 60 ft deep.
12	Center, NE¼ sec. 17	N. 35° E., 80° W. fault zone 4¼ feet wide in altered hornfels of Harmony formation; 4- to 6-in. quartz vein. Quartz white, slightly vuggy; piece from dump contains visible particles of gold.	Inclined shaft 40 to 50 ft deep.

TABLE 22.—*Gold mines and prospects in Dutch Flat*—Continued

Number on map (fig. 15)	Location (T. 38 N., R. 40 E.)	Geology	Workings
13	SW¼NW¼ sec. 16	Quartz vein 2 to 6 ft wide in altered rocks of Harmony formation. Vein strikes N. 10° E. to N. 30° E., dips 70° to 75° SE.	Two shallow shafts, short adit.
14	Approximately 300 feet southwest of No. 13.	North-trending alteration zone in sedimentary rocks of Harmony formation. Some narrow veins and veinlets of crushed, white, vuggy quartz that contains traces of stibnite and pyrite.	Prospect pit, 40-ft incline, adits 45 ft and 285 ft long.
15	Near south boundary, NE¼ sec. 17.	Adit in altered sedimentary rocks of Harmony formation follows N. 10° E., 20° W. fault containing quartz veinlets. Shaft sunk on breccia zone 3 ft wide, N. 35° E., 80° W., that contains broken quartz.	Adit 35 ft long; 25-ft shaft.
16	Near center, SE¼ sec. 17	Fault strikes N. 40° E., dips steeply northwest, in slightly altered sedimentary rocks of Harmony formation, and contains thin discontinuous quartz stringers.	Couple of shallow pits; 45-ft adit.
17	300 ft east of no. 16	Fault, N. 30° W., 10° NE. in slightly altered sedimentary rocks of Harmony formation.	Adit 25 ft long.

QUICKSILVER DEPOSITS IN THE HOT SPRINGS RANGE

Three deposits of quicksilver are known in the Hot Springs Range within the Osgood Mountains quadrangle. They are included in the Dutch Flat (Florence) district by Bailey and Phoenix (1944, p. 90–91), although only one of the properties is near Dutch Flat.

DUTCH FLAT MINE

The Dutch Flat mine is in the northern part NE¼ sec. 17, T. 38 N., R. 40 E. (pl. 1; fig. 15) on the west side of the Hot Springs Range near its base at an altitude of approximately 5,000 feet. The property is about 18 miles northeast of Winnemucca and 14 miles north of Golconda and is accessible by 19 miles of dirt road from the Southern Pacific and Western Pacific railroads and U.S. Highway 40 at Golconda.

Cinnabar was probably noted in placer operations as early as 1893 (Vanderburg, 1936) but was not discovered in place until 1939. The Dutch Flat Mining Co. was organized by E. R. Rogers, Ed. Hazlett, and J. W. Tuft in September 1940, and ore was developed and mined from underground workings (Bailey and Phoenix, 1944, p. 90). Intermittent production since 1942 has amounted to less than 2 flasks of quicksilver a year (Willden and Hotz, 1955, p. 663).

The mine is on a fault zone that transects slightly metamorphosed shale and feldspathic sandstone of the Harmony formation a few hundred feet west of a small stock of granodiorite. The fault zone strikes approximately N. 10° E., dips from 20° to 35° SE., and has an average width of about 5 feet over a length of 900 feet. At the south end of the zone, shallow open-cut bulldozer operations have discovered only traces of cinnabar in altered sedimentary rocks and an altered dike. The main cinnabar mineralization seems to be in a small area at the north end of the fault zone, where about 700 feet of underground workings have been developed to explore and mine the mineralized zone (fig. 19).

In the underground workings cinnabar seems to be chiefly distributed in light-gray fine-grained altered hornfels that has been fractured and sheared. The cinnabar occurs as small disseminated grains, but at a few places small concentrations of cinnabar are visible, and some shear surfaces are coated with a thin film of cinnabar "paint". Some cinnabar obviously is in small discontinuous veinlets. It seems to be restricted to the light-gray altered rock, for none was seen in brown altered hornfels. The light-gray rock tends to be in the upper part of the fault zone adjacent to the hanging wall, though in places the full width of the fault zone is occupied by the light-gray altered rock. Locally the crushed rock has been impregnated with calcite cement which makes it more resistant than the surrounding material. Cinnabar is commonly plentiful in these calcite cemented masses, and pyrite is present in visible amounts.

Bailey and Phoenix (1944, p. 91) estimated that the retorted ore averaged approximately 20 pounds of quicksilver to the ton. The ore zone has not been systematically sampled, but even at its northern end, where it is most strongly mineralized, it is unlikely that ore of this grade could now be mined consistently. The

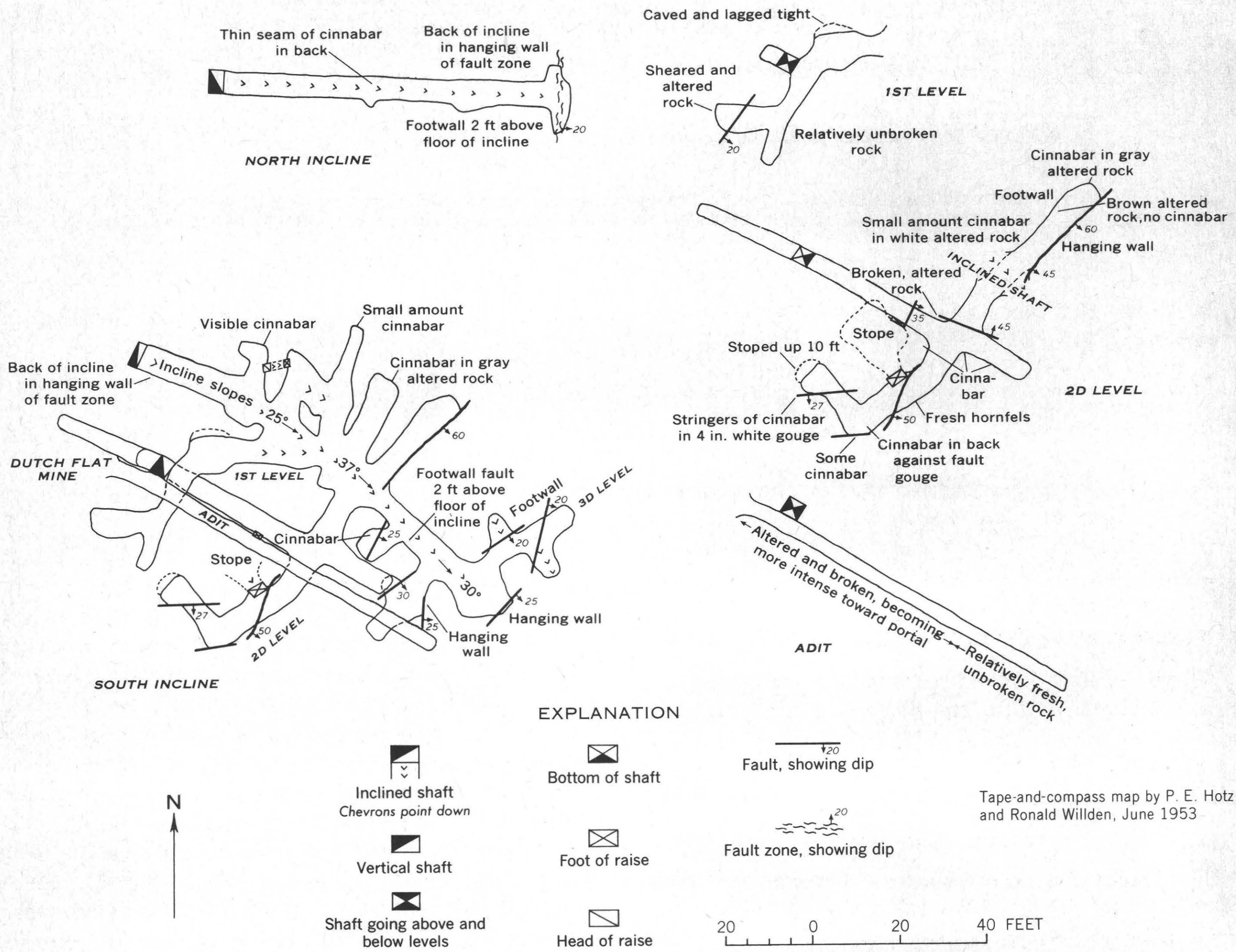


FIGURE 19.—Map of the Dutch Flat mine.

past history of the underground development and mining has shown that the ore shoots are small and discontinuous and considerable dilution of the ore with barren or low-grade material is unavoidable, because of the narrowness of the ore zone.

LAST CHANCE PROSPECT

The Last Chance prospect is in NW¼SE¼ sec. 32, T. 38 N., R. 40 E., approximately 3.5 miles south of the Dutch Flat mine (pl. 1). It is at an altitude of approximately 5,000 feet on the east side and near the head of an unnamed westward-draining canyon near the south end of the Hot Springs Range. The prospect is accessible from the Dutch Flat–Golconda road by a barely passable road. The only known production was in 1941, when one flask of quicksilver was recovered from ore treated in the retorts at Dutch Flat (Bailey and Phoenix, 1944, p. 91).

The property has 2 short adits and 2 shallow shafts that were driven to explore an alteration zone in sandy rocks of the Harmony formation. The rocks exposed in the adits are light-gray and brown soft altered sandstone and sandy shale with streaks of vermillion and brown iron oxide and veinlets of harder limonite. A few veinlets of white quartz cut the rock. No cinnabar is visible and none was obtained by panning a sample taken from one of the adits.

The altered zone has an indefinite width of about 20 feet. The zone trends slightly west of north, nearly parallel to the strike of the beds, which dip moderately to gently east. Alteration and iron oxide stains were also seen in rocks which are exposed at the surface approximately 0.2 mile south of the prospect on the southwest side of the saddle at the head of the canyon and which may be a continuation of this zone. The alteration zone is not visibly related to faulting, however, it is parallel to the strike of a fault approximately 400 feet to the west (pl. 1).

K AND K (RED DEVIL) PROSPECT

The K and K quicksilver property is at the south end of the Hot Springs Range in sec. 5, T. 37 N., R. 40 E., at an altitude of about 4,800 feet pl. 1). It is easily accessible by 13.5 miles of dirt road from Golconda. Presumably this is the property referred to by Bailey and Phoenix (1944, p. 91) as the Red Devil prospect. Cinnabar was discovered here in 1941 on land owned by the Southern Pacific Railroad, from whom the property, consisting of about 197 acres, was purchased by Major Estridge of Winnemucca, Nev. An inclined shaft about 40 feet long was sunk immediately after the discovery but no production of quicksilver was recorded. A vertical shaft 20 feet deep on a ridge one-half mile northeast of the incline may also have been sunk at this time. The property was acquired by Humboldt County after the death of Estridge, and Charles Kassebaum purchased it in July 1950 from the county. Aside from deepening the incline a few feet and doing some bulldozer work, the new owner performed no additional exploration or development work. However, a small gas-fired Rossi-type retort and condenser were installed in 1953; the dump from the incline was treated, and a small amount of quicksilver was obtained. In January 1955 the property was leased by Lilburn Davey and Cleto, Frank, and Chris Bengoa. They drove an adit 310 feet long and a raise from the adit to the incline shaft. During the summer and fall of 1955 the lessees erected a small jaw crusher, a 5-stamp mill, and two concentrating tables at Stone Corral, about 2½ miles east of the mine, where they planned to concentrate the cinnabar prior to roasting.

The country rocks are feldspathic sandstone, some shale, and a few thin limestone members—all of the Harmony formation. The prevailing strike of beds is north and the moderate dips are east, but there is some evidence of close folding and disruption by faulting. The mineralized zone is about 200 feet east of a fault that strikes north and dips at a moderate angle east and has been traced for about 4,000 feet south of the mine shaft. In 1955 the mineralized zone had been explored by an inclined shaft and an adit at the discovery site, but its length and continuity were unknown.

Cinnabar occurs in a faulted zone in soft altered feldspathic sandstone and shale. Where it is exposed in the adit the zone of alteration is about 90 feet wide. The rocks are most altered in the faulted zone, where cinnabar mineralization is also most intense. In the faulted zone the fractures strike from N. 10° W. to N. 30° W. and dip from 35° to 80° E. Visible amounts of cinnabar occur in brecciated quartz and calcite veinlets along a faulted contact between sandstone and shale in the adit, and in a sublevel between the adit and inclined shaft. In the inclined shaft cinnabar grains are in fractured and altered rock, and some apparently barren altered rock yields fine-grained cinnabar when crushed and panned.

LEAD-ZINC DEPOSITS

The lead-zinc deposits are at the Richmond mine in the northern part of the Osgood Mountains and the Silver Hill prospect at the south end of the Hot Springs Range. Neither has any recorded production.

RICHMOND MINE

Before development and operation as a tungsten property, the Richmond mine (pl. 1) in SW¼ sec. 31, T. 39 N., R. 42 E., on the northwest side of the Osgood Mountains was operated in a small way for many years as a silver mine. According to Hess and Larsen (1921, p. 303) small irregular bunches and lenses of quartz carrying some galena and considerable scheelite occurred on the limestone side of the granodiorite contact. The ore was mined from underground workings, the portal of which is in the upper part of Anderson Canyon. The original workings were later extended to develop and mine the tungsten ore. Specimens on the dump are like those found in places at the surface near the granodiorite contact and are composed of silicated limestone containing small seams and bunches of galena intergrown with smaller amounts of dark-brown sphalerite associated with pyrite, and some chalcopyrite in a gangue of milky quartz and coarsely crystalline calcite.

SILVER HILL MINE

The Silver Hill mine is at the south end of the Hot Springs Range on the west side of a small ravine in center SE¼ sec. 5, T. 37 N., R. 40 E., 0.25 mile southeast of the K and K quicksilver property (pl. 1). The workings consist of a short adit with a small stope, and a shallow vertical shaft to the north from which some short exploratory crosscuts and an inclined winze were driven. The adit and stope were driven on small west-trending mineralized fractures and brecciated zones in sandstone of the Harmony formation.

Fine-grained galena and subordinate amounts of sphalerite and traces of chalcopyrite accompanied by some vuggy quartz fill the fractures and spaces between breccia fragments of country rock and earlier quartz with but slight replacement. Anglesite and cerussite which are formed by supergene alteration of the galena, partly replace galena and fill openings in it and the vein quartz and country rock. In places fractures in the wallrocks are filled and coated with white to greenish radiating aggregates of the secondary zinc mineral hemimorphite.

UNNAMED PROSPECT

A prospect whose name is not known is on the east side of the Osgood Mountains on the south side of the ridge between Osgood and Julian (Kirby) Creeks, in NW¼SW¼ sec. 17, T. 38 N., R. 42 E. The workings include an inclined shaft about 50 feet long and an open cut 20 feet long, which are in light-gray sugary-textured recrystallized limestone of the Preble formation. The prospect is approximately 600 feet from the narrow "waist" of granodiorite between the southern and northern lobes of the stock. No ore was seen in place, but some of the limestone on the dump contains a small amount of galena.

COPPER DEPOSITS

Some small exploratory cuts and adits have been made on showings of copper in the Osgood Mountains, but no important or even promising deposits have been discovered. None are known in the Hot Springs Range.

BARITE PROSPECTS

BARIUM AND BARUM GROUPS OF CLAIMS

The principal barite deposit is covered by the Barium and Barum groups of claims, five in each, which almost cover a low hill at the east front of the Osgood Mountains in the eastern part sec. 12, T. 37 N., R. 41 E. (pl. 1), near the mouth of Hogshead Canyon. The property, originally located by P. V. Sanders and W. M. Pettit in September 1940, is held by the Baroid Division of National Lead Company. No recorded production of barite from the property has been made; the principal activity has been surface exploration and trenching done for assessment work.

The country rocks are limestone, dolomite, shale, and black chert of the Comus formation of Ordovician age. On the eastern and southeastern slopes of the hill the rocks are mainly interbedded limestone, dolomite, and shale; the summit area and western slope is black chert cut by veinlike bodies of brown jasper and quartz. The country rocks are contorted and are not well enough exposed to permit the structure to be clearly deciphered: at the largest open cut on the southeast slope of the hill, beds strike generally east-west and dip moderately to the south, whereas elsewhere in the area strikes are north and dips are moderately east or steeply west. Barite occurs mainly in the southeast area, where it replaces the carbonate rocks but not the shales. Locally, however, chert is also replaced by barite along zones of brecciation accompanied by jasper and veins of quartz.

Some surface exploration has been done, mainly by means of shallow bulldozer cuts, but not enough to delimit clearly the extent of the mineralization. It has been shown, however, that barite deposits occur over an area of about 20 acres.

OTHER PROSPECTS

A shallow prospect pit near the Blue Bell mine in north-central part NE¼NW¼ sec. 5, T. 37 N., R. 42 E. (pl. 1), exposes interbedded limestones and shales of the Comus formation. The limestone in the pit is partly replaced by barite.

A barite prospect known as the Thirsty Lode claim was located in May 1951 by Lyman Thomas and R. L. Taylor, north of Anderson Canyon in extreme SW¼ SW¼ sec. 30, T. 39 N., R. 42 E. (pl. 1). Here, fine-grained gray barite apparently replaces a limestone bed in the Preble formation. The bed strikes N. 20° W. and dips 40° northeast; the thickness and strike length of the barite replacement is not known.

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