3680 0071

(136) Item 71

Excerpted and condensed from:

U.S. Geol. Surv. Prof. Paper 431, 1964.

GEOLOGY AND MINERAL DEPOSITS OF THE OSGOOD MOUNTAINS QUADRANGLE, HUMBOLDT COUNTY, NEVADA

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 conspicious 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.

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 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 métamorphism 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 NE¼ SW¼NW¼ 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.

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* Rudemann
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

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.

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.

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 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 bound-

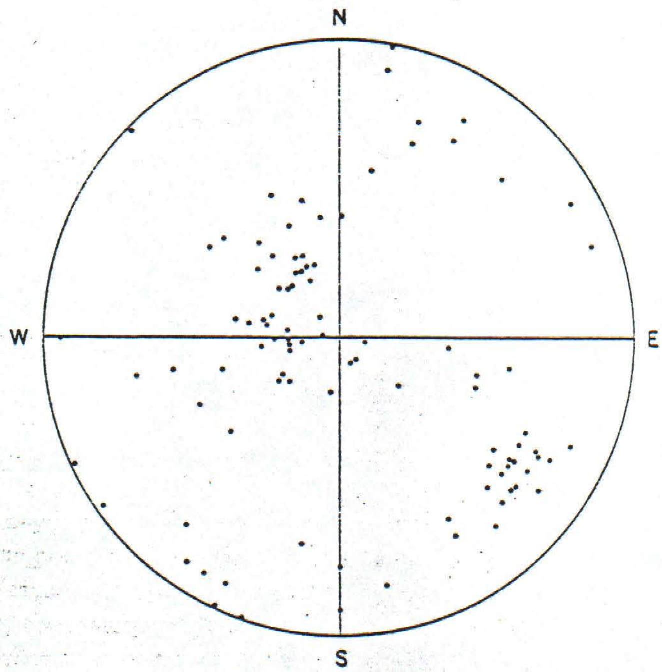

FIGURE 3.—Point diagram of poles of 95 joints measured in granodiorite stock of the Osgood Mountains. Plotted on upper hemisphere.

ary. 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.

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 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 modera 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 elsewhere 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 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 grad

[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.

tional 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 themselves. 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.

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-

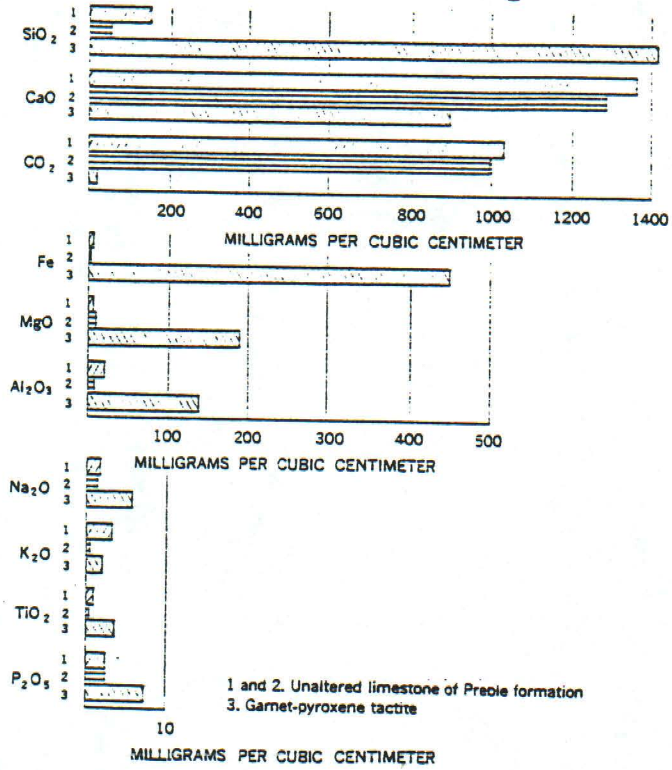

Figure 9.—Graph showing gain and loss of rock constituents in conversion of limestone to garnet-pyroxene tactite.

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.

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.

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.

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,[23] 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 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 2,300 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	2,984	----------	----------	$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	----------	10,732	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.

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.

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 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."